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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]

[a] CCFS ARC Centre of Excellence and Macquarie Planetary Research Centre, Department of Earth and Planetary Science, Macquarie University, Sydney, Australia
[b] Rice University, Houston, TX, USA
[c] Monash University, Melbourne, Australia
[d] The University of Melbourne, Melbourne, Australia[1]

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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 Pearse (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)). 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).

* Corresponding author.
[1] Now at.

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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 as 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:

$$\tau_{ij,j} - p_{,i} = f_i \tag{1}$$

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

$$\tau_{ij} = 2\eta\dot{\varepsilon}_{ij} \equiv \eta(v_{i,j} + v_{j,i}) \tag{2}$$

This momentum equation is subject to the incompressibility constraint:

$$v_{i,i} = 0 \tag{3}$$

And the energy equation is given by:

$$\partial T/\partial t + v \cdot \nabla T = \kappa\nabla^2 T + Q(t) \tag{4}$$

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

$$x_i = dx_i' \qquad T = \Delta T T' \qquad \eta = \eta_0\eta' \tag{5}$$

where d is the layer depth (2,890,000 m, Fig. 1), η_0 is a reference viscosity (1e21 Pa.s), and ΔT 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:

$$\eta = A \cdot e^{-T_1 T} \tag{6}$$

The viscosity varies from 1 at the bottom (where the reference basal temperature is 1), to 3×10^4 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:

$$\tau_{\text{yield}} = B_0 + B_z \cdot z \tag{7}$$

The default non-dimensional parameters are set to $B_0 = 1 \times 10^5$ and $B_z = 1 \times 10^7$ (these equate to a cohesion of ~12 MPa, and a coefficient of friction of ~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

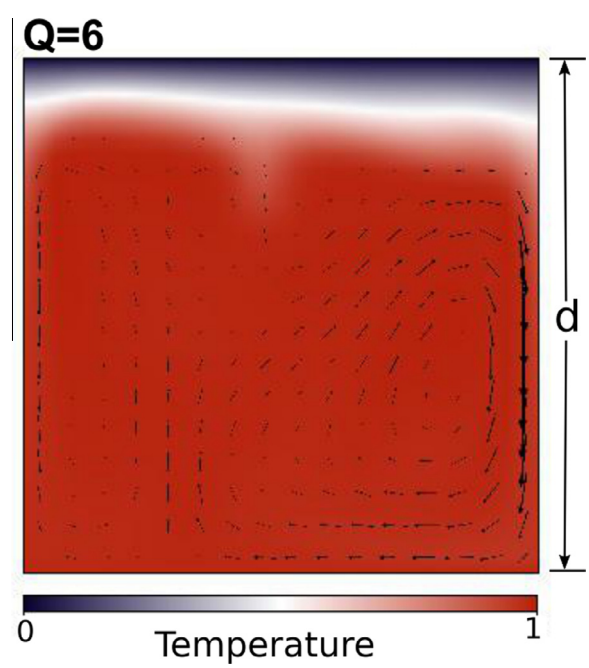

Fig. 1. Basic model configuration for the parameters outlined in Table 1, and an internal heating rate $Q = 6$. Shown are the temperatures field, and velocity field (arrows). The Rayleigh number for this simulation is 1×10^7. The system incorporates viscoplasticity, and is in a steady-state "hot-stagnant-lid" regime.

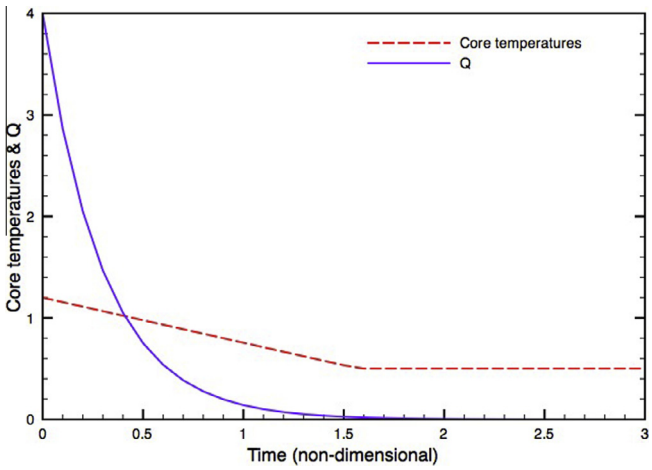

Fig. 2. Evolutionary curves for heat production Q, and core temperatures. These evolving conditions are imposed on our evolutionary models. See text for details.

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:

$$Q(t) = Q_0 e^{-\lambda t} \tag{8}$$

where $Q_0 = 2.25$ and $\lambda = 3.34239$, are set to give equivalent present day heat production values, and a decay over time t 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 Ga ($T_{base} = 1.2$ 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.

The resolution is 64 × 64 (×64) nodes for the Cartesian 2 and 3 dimensional calculations. We have included an Underworld input as an online supplement which also outlines the parameters used, and allows duplication of the basic results. Within it are the modifications required to the basic underworld code to enable evolutionary calculations, which should assist replicability.

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 (Tb, Ts) at: Ra = 1e5, $\Delta\eta$ = 1e4, and Tb = 1, and Ts = 0. Viscosity is temperature, and depth dependent. The yield strength σ_{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 σ_{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 0 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

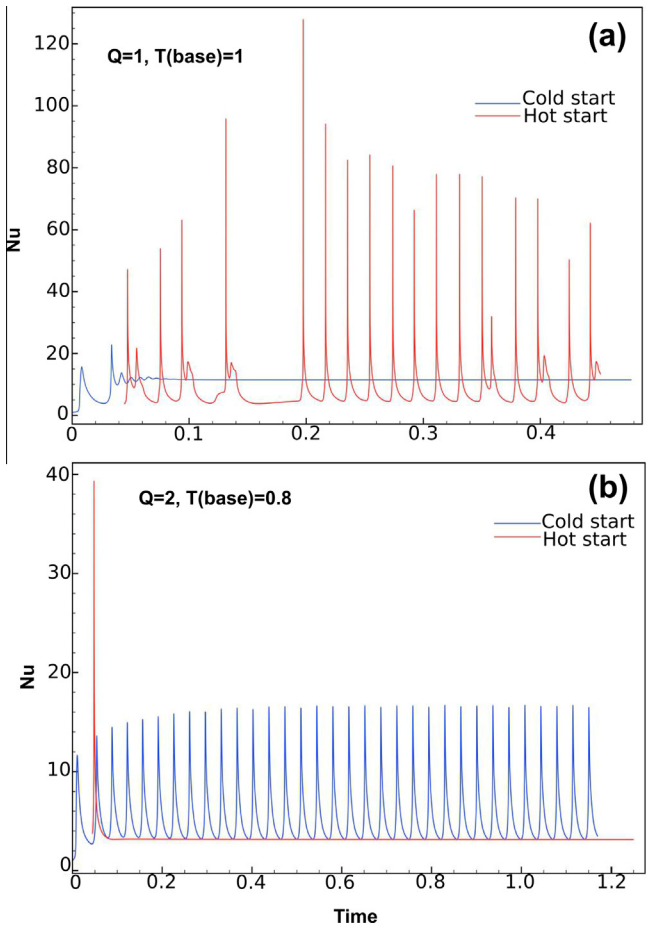

Fig. 3. (a) Nusselt number versus time for two simulations with identical system parameters, but different starting conditions. System parameters as per Table 1, with $Q = 1$ and $T_{base} = 1$. The initial condition for the cold-start case is a linear temperature gradient between the top and the bottom, 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(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).

Ra		1×10^7
Viscosity	$\eta = \eta_0\, e^{-T1T}$	
η_0		3×10^4
T_1		10.31
Yield parameters		
B_0		1×10^5
B_z		1×10^7
Thermal conditions		
Q		0.5*
T_{base}		1.0
T_0		0.0

* 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_{basal} = 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.

(t = 0.1–1) to ensure that a specific tectonic regime exists for the parameter values being examined. The modelling domain consists of 32 × 32 × 32 grid cell elements for each of the 12 spherical caps and boundary conditions are free slip. Each run is divided into a viscosity plot (grey shells are regions of high viscosity "plates" and yellow bands are regions of active yielding) and thermal profiles from the CMB to surface. All other parameters are held constant.

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 $Q \sim 15$, 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 tectonothermal 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 $t = 1.0$ 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_{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

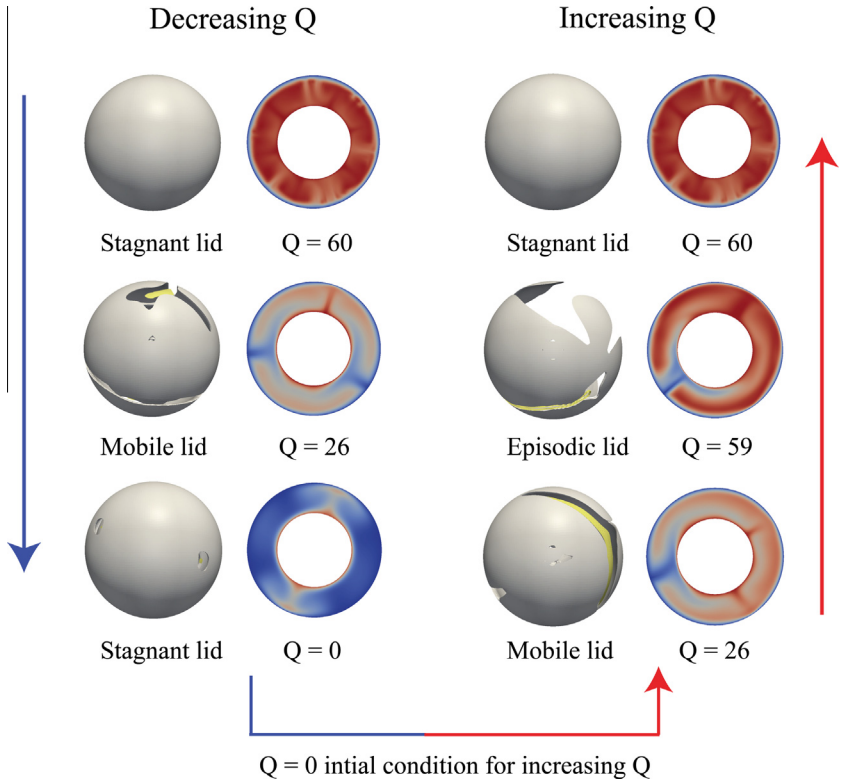

Fig. 4. Effect of initial conditions in a 3D spherical model, using CitcomS. The temperature-dependent viscosity contrast is 1×10^4, and is also depth dependent. Temperature of the CMB is 1, and the yield strength is 42,500. Internal heating (Q) is varied from 60 to 0 in the decreasing pathway, and 0 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.)

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 $\sim t = 0.3$, 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 $t \sim 1.75$, 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 ($B_0 = 1 \times 10^6$, $B_z = 1 \times 10^8$). 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, transiting into a quiescent stagnant lid regime only when the CMB temperatures have declined to $\sim$0.5 (e.g. at $t \sim 2.0$, 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, transiting into a regular mobile-lid mode only after $t \sim 0.9$ (e.g. Fig. 8e). The decline in CMB temperatures and Q result in a transition into cold stagnant lid convection at around $t = 1.6$.

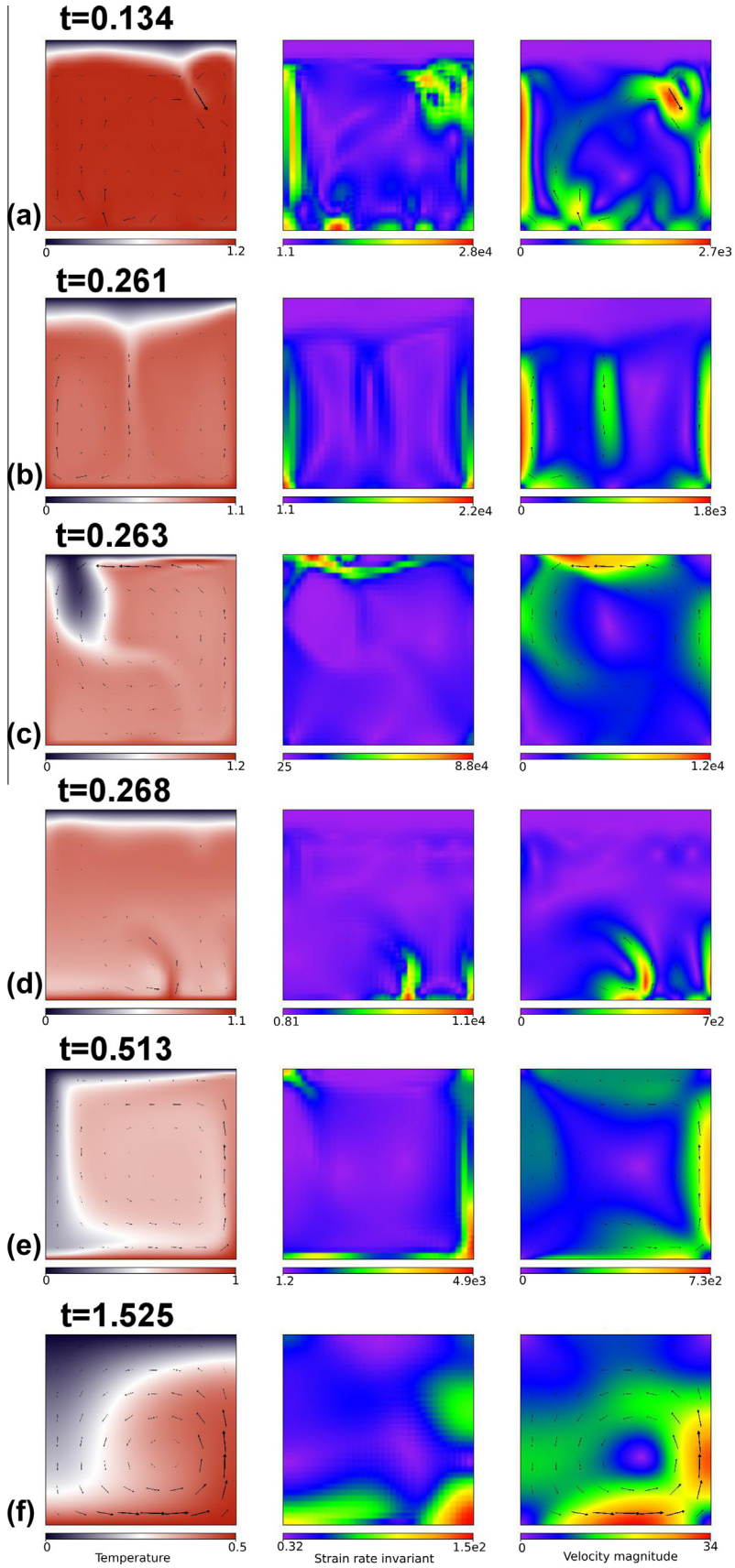

Fig. 5. Time evolution of a simulation with parameters outlined in Table 1, with imposed decaying mantle heat production, and basal temperature, as per Fig. 2. The initial condition is from a previous stagnant-lid simulation, equilibrated at $Q = 8$ (equivalent to a hot early Hadean). The systems begins in a stagnant regime, before evolving into an episodic regime (e.g. $t = 0.263$). Eventually, with waning CMB temperatures, the system transits into a plate tectonic mode ($t = 0.513$, equivalent to ~5Gyr), before winding down into a senescent stagnant-lid mode ($t = 1.525$, ~15 Gyr; Note arrow size rescales to system maximum each timestep).

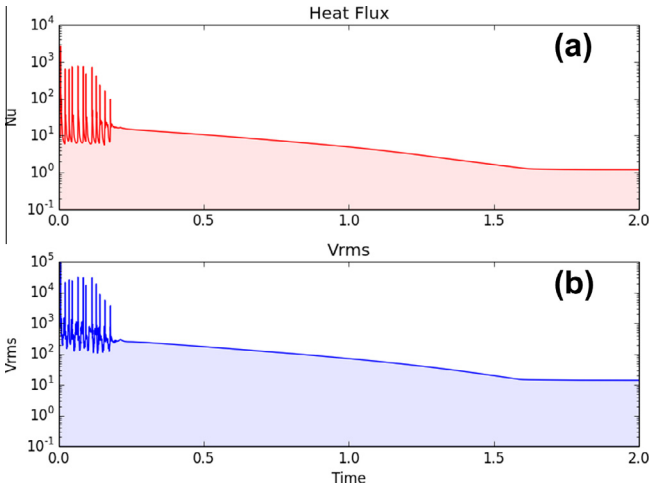

Fig. 6. Nusselt number vs non-dimensional time for the simulation shown in Fig. 5 (a) log linear plot of the full extent of Nu range, including extreme peaks during initial overturn events, and (b) root-mean square velocity of the system, which demonstrates the evolution from episodic convection to mobile lid ($t \sim 0.35$), and eventually to cold stagnant lid convection ($t \sim 1.7$).

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- > 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 dramatic 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 significant 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 evolutionary stage, Lenardic and Crowley (2012) note that, for equivalent 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 condition equilibrated at $Q = 2$. With this starting condition, we explore a number of different evolving heat production functions. Exponential heat decay is of the form described in Eq. (8). In each case, we vary the prefactor Q_0 for our heat production to be 2.0, 4.0, 6.0, and 8.0. The decay constant λ 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 production through time leads the systems rapidly into a mobile-lid regime, and heat flow smoothly follows the decay in heat production 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 nonlinearities of the tectonic regime, particularly at early times, modulate that trend.

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

Variations in the evolution of core–mantle boundary temperatures 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 constant λ 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 evolution 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 conditions, 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.

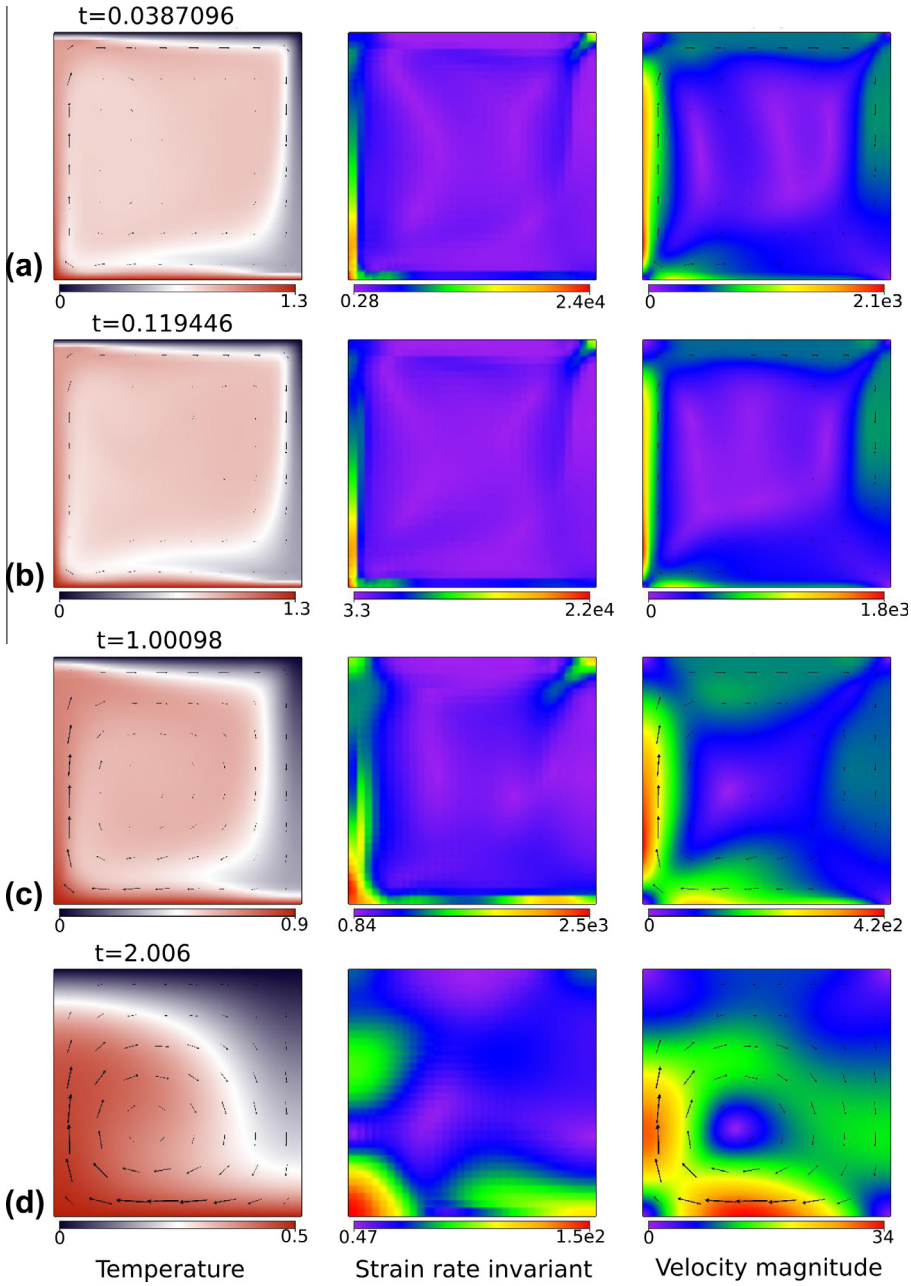

Fig. 7. Time evolution of a system with the default values listed in Table 1, but with an initial condition taken from a previous stagnant-lid simulation with $Q = 2$. The model transits rapidly into a steady plate-tectonic mode (a), which continues, with a gradually thickening thermal boundary layer (b, c), until the simulation eventually transits into a cold stagnant-lid mode (d).

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

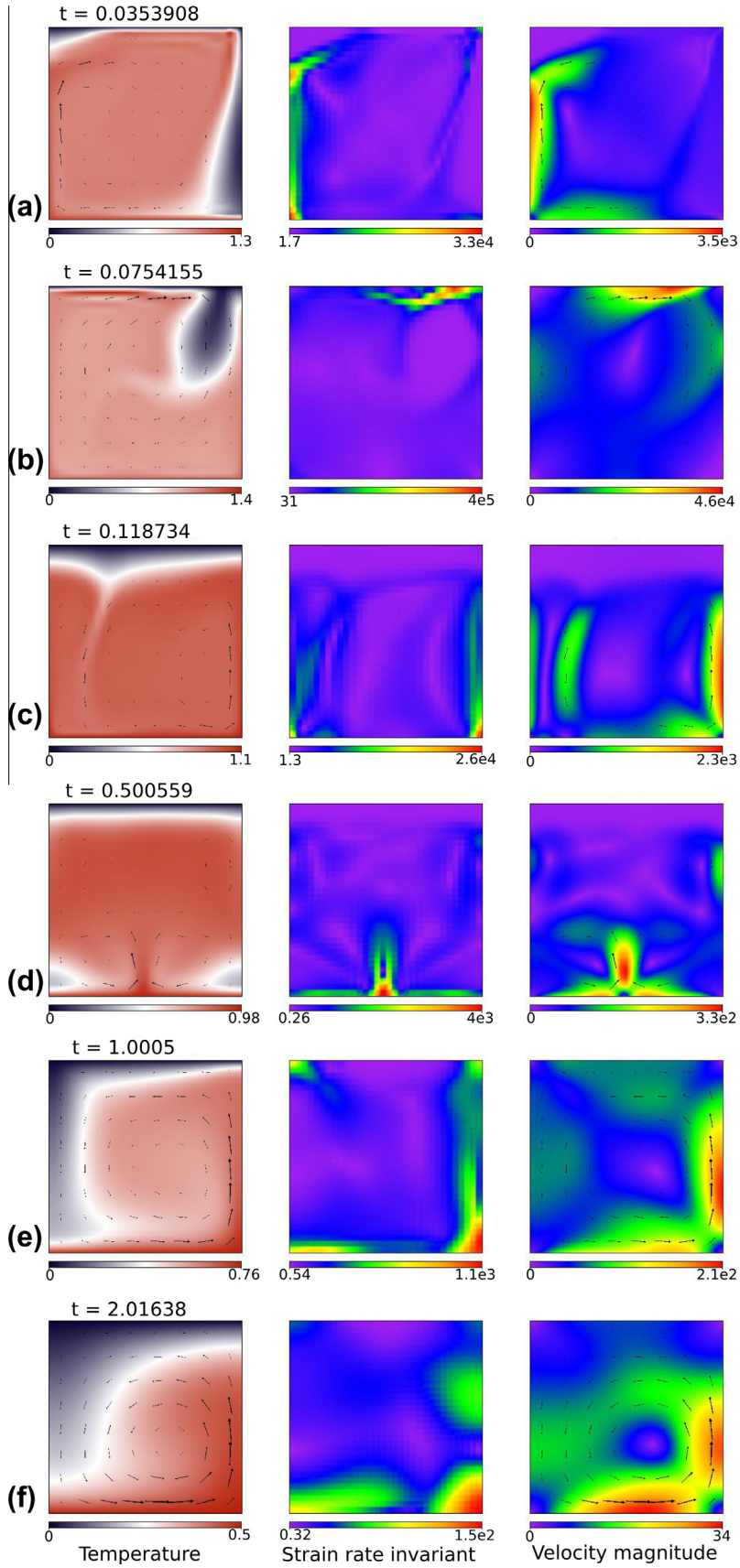

Fig. 8. Time evolution of a system with the default values listed in Table 1, but with an initial condition taken from a previous stagnant-lid simulation with $Q = 6$. The model begins in a stagnant mode, before transiting into an episodic regime (a–d). At $t \sim 1.0$ the systems evolves into a plate-tectonic regime (e), which eventually wanes into a cold stagnant lid regime (f).

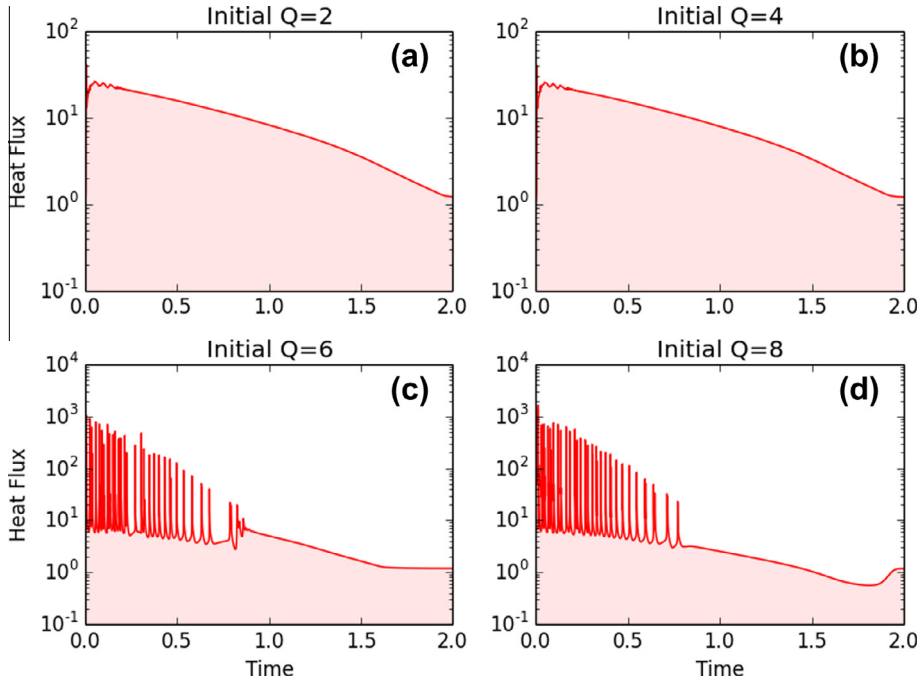

Fig. 9. Time evolution of Nusselt number, for four simulations with the same system parameters (Table 1), heat production decay rates, and core-temperature evolution, but with different starting conditions (initial conditions (IC) for $Q = 2,4,6$ and 8; the initial $Q = 2$ and $Q = 6$ curves are from the simulations shown in Figs. 7 and 8). For an initial condition with a comparatively cold temperature field (equilibrated at $Q = 2$ or 4), the system rapidly enters a steady plate tectonic mode, which it remains in until decaying into a cold stagnant lid mode at $\sim t = 2.0$. In contrast, the models with a "hot" start (initial conditions equilibrated at $Q = 6$ or 8) rapidly enter an episodic regime, which dominates their evolution, until they pass into a plate-tectonic regime at $\sim t = 0.8$, and a cold-stagnant lid mode at $\sim t = 1.6$.

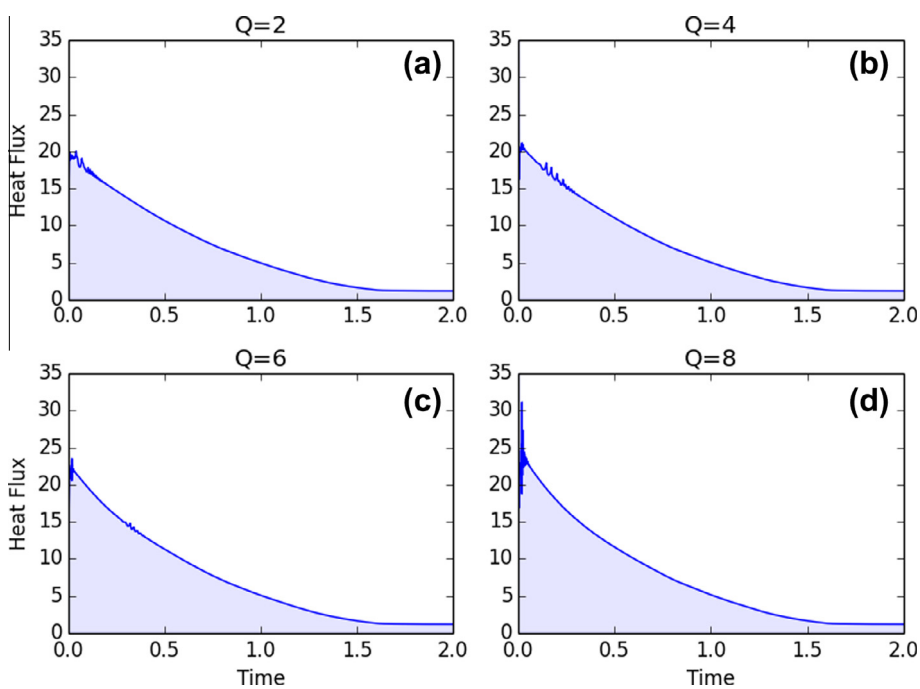

Fig. 10. Effect of varying mantle heat production values on evolution, for simulations with same system parameters (Table 1), the same starting conditions (previous stagnant-lid simulation with $Q = 2$). The prefactor Q_0 values (Eq. (7)) are set to 2, 4, 6 and 8. The decay constant λ was kept at 3.34239. Higher internal heating affects the early evolution, particularly the time-dependence in the $Q = 8$ curve. However, all simulations rapidly progress into a mobile-lid regime, and Nu decays, following the heat production decay curve, until they pass into a stagnant mode at $\sim t = 1.63$.

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 ~650 km deep on the Earth (e.g. Wood et al.

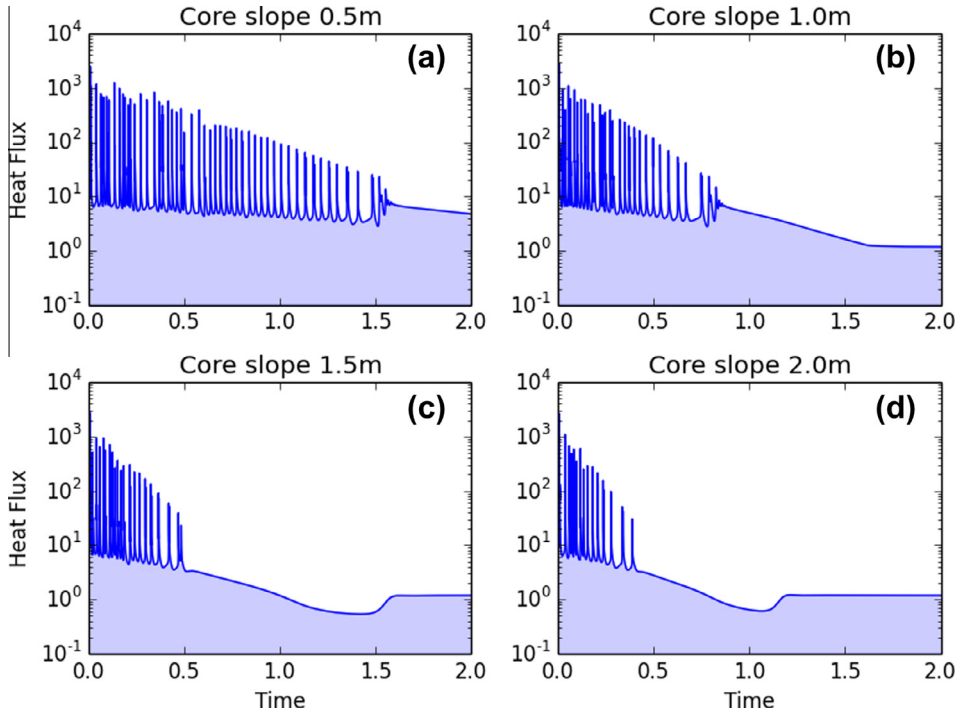

Fig. 11. Effect of core temperature evolution. Coreslope = 1 m equates to the gradient of the graph in Fig. 2. A coreslope of 0.5 m equates to a slope half that of Fig. 2 – meaning that the core evolution lifetime is twice as long. The results show that a protracted core-temperature evolution can sustain significant internal buoyancy forces for longer, leading to a longer period of active tectonism. Conversely, rapidly decaying core temperatures (e.g. Coreslope = 2 m) results in a correspondingly shorter period of active tectonism.

(2006)), 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 (Hoink 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; Hoink 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

our Cartesian models here this is superfluous, as the total basal heat flux is higher due to the larger bottom surface area, relative to a spherical model (O'Farrell and Lowman, 2010). This is clearly an avenue for future work, our first pass goal in this paper was to assess the effect an evolving core has on the mantle, without any implicit feedback. The observed effects are significant, and underscore the importance of understanding where in a tectonothermal evolutionary arc a planet lies, rather than considering time-independent factors, such as size, in isolation.

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; Weller 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 senescent 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

Butler, S., Peltier, W., 2002. Thermal evolution of earth: models with time-dependent layering of mantle convection which satisfy the urey ratio constraint. J. Geophys. Res. (Solid Earth) 107 (B6) (ESE-3).
Butler, S., Peltier, W., Costin, S., 2005. Numerical models of the earth's thermal history: Effects of inner-core solidification and core potassium. Phys. Earth Planet. Inter. 152, 22–42.
Calvert, A.J., Sawyer, E.W., Davis, W.J., Ludden, J., 1995. Archaean subduction inferred from seismic images of a mantle suture in the superior province. Nature 375, 670–674.
Canup, R.M., 2004. Simulations of a late lunar-forming impact. Icarus 168, 433–456.
Condie, K.C., Pearse, V., 2008. When Did Plate Tectonics Begin on Planet Earth? The Geological Society of America, Boulder, CO, p. 295p.
Condie, K.C., Kroner, A., 2008. When did plate tectonics? Evidence from the geological record. In: Condie, K.C., Pearse, V. (Eds.), When Did Plate Tectonics Begin on Planet Earth? The Geological Society of America, Boulder, CO, pp. 281–294.
Condie, K., O'Neill, C., 2010. The Archean–Proterozoic boundary: 500 My of tectonic transition in Earth history. Am. J. Sci. 310, 775–790.
Condie, K.C., O'Neill, C., Aster, R., 2009. Evidence and implications of a widespread magmatic shut down for 250 Myr on Earth. EPSL 282, 294–298.
Crowley, J.W., O'Connell, R.J., 2012. An analytical model of convection in a system with layered viscosity and plates. GJI 188, 61–78.
Davies, G.F., 1980. Thermal histories of convective Earth models and constraints on radiogenic heat production in the Earth. J. Geophys. Res 85 (251), 7–2530.
Davies, G.F., 1993. Conjectures on the thermal and tectonic evolution of the Earth. Lithos 30, 281–289.
Debaille, V., Brandon, A.D., O'Neill, C., Yin, Q.Z., Jacobsen, B., 2009. Early martian mantle overturn inferred from isotopic composition of nakhlite meteorites. Nat. Geosci. 2 (8), 548–552.
Debaille, V., O'Neill, C., Brandon, A.D., Haenecour, P., Yin, Q.-Z., Mattielli, N., Treiman, A.H., 2013. Stagnant-lid tectonics in early Earth revealed by 142Nd variations in late Archean rocks. EPSL 373, 83–92.
Elkins-Tanton, L.T., Parmentier, E.M., Hess, P.C., 2003. Magma ocean fractional crystallization and cumulate overturn in terrestrial planets: implications for Mars. Meteorit. Planet. Sci. 38 (12), 1753–1771.
Escartin, J., Hirth, G., Evans, B., 1997. Effects of serpentinization on the lithospheric strength and the style of normal faulting at slow- spreading ridges. Earth Planet. Sci. Lett. 151, 181–190.
Foley, B.J., Bercovici, D., Landuyt, W., 2012. The conditions for plate tectonics on super-Earths: inferences from convection models with damage. Earth Planet. Sci. Lett. 331, 281–290.
Hoink, T., Lenardic, A., 2009. Long wavelength convection, Poiseuille–Couette flow in the low-viscosity asthenosphere and the strength of plate margins. GJI 180, 23–33.
Hoink, T., Lenardic, A., Richards, M., 2012. Depth-dependent viscosity and mantle stress amplification: implications for the role of the asthenosphere in maintaining plate tectonics. GJI 191, 30–41.
Hopkins, M., Harrison, T.M., Manning, C.E., 2008. Low heat flow inferred from >4 Gyr zircons suggests Hadean plate boundary interactions. Nature 456, 493–496.
Jellinek, A.M., Jackson, M.G., 2015. Connections between the bulk composition, geodynamics and habitability of Earth. Nat. Geosci. 8 (8), 587–593.
Karato, S.I., 2014. Some remarks on the models of plate tectonics on terrestrial planets: From the view-point of mineral physics. Tectonophysics 631, 4–13.
King, S.D., 2015. Mantle convection, the asthenosphere, and Earth's thermal history. Geol. Soc. Am. Spec. Pap. 514 (SPE514-07).
Korenaga, J., 2010. On the likelihood of plate tectonics on super-Earths: does size matter? Astrophys. J. Lett. 725 (1), L43.
Lenardic, A., Jellinek, A.M., Moresi, L.-N., 2008. A climate change induced transition in the tectonic style of a terrestrial planet. Earth Planet. Sci. Lett. 271, 34–42.
Lenardic, A., Crowley, J.W., 2012. On the notion of well-defined tectonic regimes for terrestrial planets in this solar system and others. Astrophys. J. 755 (2), 132.
Moore, W.B., 2008. Heat transport in a convecting layer heated from within and below. J. Geophys. Res. 113 (B11407).
Moore, W.B., Webb, A.A.G., 2013. Heat-pipe earth. Nature 501 (7468), 501–505.
Moore, W.B., Lenardic, A., 2015. The efficiency of plate tectonics and nonequilibrium dynamical evolution of planetary mantles. Geophys. Res. Lett.
Moresi, L., Solomatov, V., 1998. Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. Geophys. J. Int. 133, 669–682.
Moresi, L., Quenette, S., Lemiale, V., Meriaux, C., Appelbe, B., Mühlhaus, H.B., 2007. Computational approaches to studying non-linear dynamics of the crust and mantle. Phys. Earth Planet. Inter. 163, 69–82. http://dx.doi.org/10.1016/j.pepi.2007.06.009.
Moyen, J.-F., van Hunen, J., 2012. Short-term episodicity of Archaean plate tectonics. Geology 40, 451.
Nakagawa, T., Tackley, P.J., 2012. Influence of magmatism on mantle cooling, surface heat flow and urey ratio. Earth Planet. Sci. Lett. 329, 1–10.
Nakagawa, T., Tackley, P.J., 2015. Influence of plate tectonic mode on the coupled thermochemical evolution of earth's mantle and core. Geochem. Geophys. Geosyst. 16, 2015G. http://dx.doi.org/10.1002/C005996.
Noack, L., Breuer, D., 2014. Plate tectonics on rocky exoplanets: Influence of initial conditions and mantle rheology. Planet. Space Sci. 98, 41–49.
Noack, L., Breuer, D., 2013. First-and second-order Frank-Kamenetskii approximation applied to temperature-, pressure-and stress-dependent rheology. Geophys. J. Int. http://dx.doi.org/10.1093/gji/ggt248.
Nimmo, F., Stevenson, D.J., 2000. Influence of early plate tectonics on the thermal evolution and magnetic field of Mars. J. Geophys. Res. 105, 11969–11979.
Nimmo, F., Price, G.D., Brodholt, J., Gubbins, D., 2004. The influence of potassium on core and geodynamo evolution. GJI 156, 363–376.
O'Farrell, K.A., Lowman, J.P., 2010. Emulating the thermal structure of spherical shell convection in plane-layer geometry mantle convection models. Phys. Earth Planet. Inter. 182 (1), 73–84.
O'Farrell, K.A., Lowman, J.P., Bunge, H.P., 2012. Comparison of spherical-shell and plane-layer mantle convection thermal structure in viscously stratified models with mixed-mode heating: implications for the incorporation of temperature-dependent parameters. Geophys. J. Int., ggs053
O'Neill, C., Jellinek, A.M., Lenardic, A., 2007a. Conditions for the onset of plate tectonics on terrestrial planets and moons. EPSL 261, 20–32.
O'Neill, C., Lenardic, A., Moresi, L., Torsvik, T., Lee, C.A., 2007b. Episodic Precambrian subduction. EPSL 262, 552–562.
O'Neill, C., Lenardic, A., 2007. Geological consequences of super-sized Earths. Geophys. Res. Lett. 34 (19).

O'Neill, C., Lenardic, A., Griffin, W.L., O'Reilly, S.Y., 2008. Dynamics of cratons in an evolving mantle, invited paper for a *Lithos* special volume on continental volcanism. Lithos 102, 12–24.

O'Neill, C., Lenardic, A., Coltice, N., Moresi, L., 2013. Mantle convection and outgassing on terrestrial planets. In: Mackwell, S.J. et al. (Eds.), Comparative Climatology of Terrestrial Planets. Univ. of Arizona, Tucson, pp. 473–486. http://dx.doi.org/10.2458/azu_uapress_9780816530595-ch19.

O'Neill, C., Debaille, V., 2014. The evolution of Hadean-Eoarchaean geodynamics. Earth Planet. Sci. Lett. 406, 49–58.

Reese, C.C., Solomatov, V.S., Moresi, L.-N., 1999. Non-Newtonian stagnant lid convection and magmatic resurfacing on Venus. Icarus 139, 67–80.

Richards, M., Yang, W.-S., Baumgardner, J., Bunge, H.-P., 2001. Role of a low-viscosity zone in stabilizing plate tectonics: implications for comparative terrestrial planetology. Geochem. Geophys. Geosyst. (2000GC000115)

Sandu, C., Lenardic, A., McGovern, P., 2011. The effects of deep water cycling on planetary thermal evolution. J. Geophys. Res. 116, B12494. http://dx.doi.org/10.1029/2011JB008405.

Solomatov, V.S., 1995. Scaling of temperature-and stress-dependent viscosity convection. Phys. Fluids 7, 266–274.

Stamenković, V., Breuer, D., 2014. The tectonic mode of rocky planets: Part 1–Driving factors, models & parameters. Icarus 234, 174–193.

Stein, C., Schmalzl, J., Hansen, U., 2004. The effect of rheological parameters on plate behaviour in a self-consistent model of mantle convection. Earth. Planet. Sci. Lett. 142, 225–255.

Stein, C., Lowman, J.P., Hansen, U., 2012. The influence of mantle internal heating on lithospheric mobility: Implications for super-Earths. Earth Planet. Sci. Lett. 361, 448–459.

Stern, R.J., 2008. Modern-style plate tectonics began in Neoproterozoic time: an alternative interpretation of Earth's tectonic history. In: Condie, K.C., Pearse, V. (Eds.), When Did Plate Tectonics Begin on Planet Earth? The Geological Society of America, Boulder, CO, pp. 265–280.

Stevenson, D.J., 1990. Fluid dynamics of core formation. In: Newsom, H.E., Jones, J.H. (Eds.), Origin of the Earth. Oxford University Press, New York, pp. 231–250.

Tozer, D.C., 1972. The present thermal state of the terrestrial planets. Phys. Earth Planet. Int. 6, 182–197.

Turcotte, D.L., Schubert, G., 1982. Geodynamics: applications of continuum physics to geological problems. John Wiley, New York, p. 450.

Valencia, D., O'Connell, R.J., Sasselov, D.D., 2007. Inevitability of plate tectonics on super-earths. Astrophys. J. 670, L45–L48.

van Heck, H.J., Tackley, P.J., 2011. Plate tectonics on super-Earths: Equally or more likely than on Earth. EPSL 310, 252–261.

Weller, M.B., Lenardic, A., 2012. Hysteresis in mantle convection: Plate tectonic systems. GRL 39. http://dx.doi.org/10.1029/2012GL051232.

Weller, M.B., Lenardic, A., O'Neill, C., 2015. The effects of internal heating and large scale climate variations on tectonic bi-stability in terrestrial planets. Earth Planet. Sci. Lett. 420, 85–94.

Wood, B.J., Walter, M.J., Wade, J., 2006. Accretion of the Earth and segregation of its core. Nature 441 (7095), 825–833.

Yuen, D.A., Balachandar, S., Steinbach, V.C., Honda, S., Reuteler, D.M., Smedsmo, J.J., Lauer, G.S., 1995. Non-equilibrium effects of core-cooling and time-dependent internal heating on mantle flush events. Nonlinear Processes Geophys. 2, 206–221.

Zhang, S., O'Neill, C., 2015. The early geodynamic evolution of Mars-type planets. Icarus.

Zhong, S., Gurnis, M., 1995. Mantle convection with plates and mobile, faulted plate margins. Science 267, 838–843.

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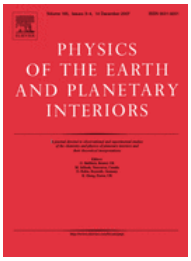

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