Time-dependence in mantle convection models featuring dynamically evolving plates

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SUMMARY
We present the findings from a study of 2-D Cartesian geometry mantle convection simulations carried out to determine how plate velocities and the surface and basal heat flux respond to an evolving plate geometry. We model flow for periods equating to hundreds of millions of years and find that while calculations that feature fixed plate geometries exhibit regular reversals in plate motion, this behaviour is absent in models featuring evolving plate boundaries. However, simulations featuring evolving plate boundaries and plates with either a fixed-thickness or temporally varying thickness are highly time-dependent and their globally averaged mean plate velocities are higher than in simulations featuring static plate geometries. Our models featuring evolving plate geometries assume that (1) young plates override old plates with the velocity of the younger plate and (2) that symmetric seafloor spreading occurs at divergent plate boundaries. Plate velocities are dynamically determined in accord with a force-balance modelling approach and the plate velocities determine the dynamic evolution of the plate boundaries according to criteria (1) and (2), above. Plates in our calculations are highly viscous and are treated as rigid blocks for the purpose of determining whether they will break according to a yield stress criterion. The modelling of plate rifting means that both the number and mean size of the plates in our calculations are time-dependent. We focus on isolating the influence of plate thickness on time-dependent flows and examine the time-dependence of global plate velocities and mantle and core heat flow. In an initial study of unit aspect ratio models we find that there is a transition in the character of time-dependent behaviour as the model plate thickness is increased. Plates that are comparable in thickness to the mean thickness of the thermal boundary layer exhibit periodic reversals. Plates that are much thinner than the thermal boundary layer exhibit no reversals. Intermediate thickness plates reverse intermittently. In aspect ratio 12 calculations, we find that the mean global plate velocity can exhibit variations of more than a factor of two over periods of tens of millions of years. We also find that the mean surface velocity of the plates and the mean global heat flux rise and fall in tandem (although high frequency variations in the global velocity are not reflected in the heat flux time-series). In a simulation spanning almost 3 Gyr we observe several instances of surface heat flux fluctuations of 40–50 per cent occurring within periods of 75–200 Myr.

Key words: flow reversals, heat flux, mantle convection, plate tectonics, time-dependence.

1 INTRODUCTION
The Earth is cooling by a different process than the other terrestrial planets because its upper cold thermal boundary layer (the lithosphere) participates in a convective flow extending deep into its interior (Grand 1994; Grand et al. 1997; van der Hilst et al. 1997).

In addition, surface deformation is confined to the narrow linear bands defining the tectonic plate boundaries. In the vast regions lying between the plate boundaries, the thermally activated rheology of the mantle makes the cold boundary layer strong and to first order each region defining a distinct tectonic plate moves across the surface with an almost uniform velocity. Plates, therefore, have a profound impact on heat transfer in the mantle and consequently, understanding terrestrial mantle convection requires the study of convection incorporating a system of dynamically evolving plates.

Attaining plate-like surface motion in self-consistent mantle convection experiments (featuring complex rheologies) has been an
active field of research for some time and promising advances have been made in this area (e.g. Zhong & Gurnis 1996; Moresi & Solomatov 1998; Trompert & Hansen 1998; Tackley 1998, 2000; Stein et al. 2004). However, obtaining plate-like surface motion over periods that equate with hundreds of millions of years when scaled to convection in the Earth remains a significant challenge. Yet, it is possible to investigate the feedback between mantle convection and plate-like surface motion with an alternative modelling approach. If the existence of the plates is accepted a priori and their qualities (e.g. uniform velocity and rigidity) are specified explicitly, mantle convection featuring plate-like surface motion can be modelled over extensive time periods (e.g. Monnereau & Quéré 2001; Bunge et al. 2002; Lowman et al. 2004; Quéré & Forte 2006).

Previous authors have studied the interaction between plate-like surface motion and convection by prescribing uniform velocity boundary conditions and fixed plate geometries (e.g. Parmentier & Turcotte 1978; Lux et al. 1979; Jarvis & Peltier 1981; Davies 1989; Bunge & Richards 1996; Jellinek et al. 2003). This work has shown that when plate speeds are comparable to the velocities induced by the buoyancy driven flow the plate length is mirrored by the convective wavelength of the system. In 2-D numerical studies featuring experiments performed in systems with a surface extent comparable to the Earth’s circumference, Gurnis & Zhong (1991) and Zhong & Gurnis (1993) showed that a single plate (supercontinent) with dynamically determined motion would instil its wavelength in the convecting mantle. More recent work (e.g. Lowman et al. 2001; Koglin Jr. et al. 2005) has demonstrated that the feedback between internally heated convection and dynamically determined plate velocities can result in a pattern of time-dependent plate motion characterized by intermittent to regularly reversing plate velocity. The flow reversals are driven by changes in the rotational sense of the overturn of the associated convection cells and have been observed in both unit aspect ratio calculations and wide multiple plate systems. Taken together, these studies indicate that plate scale profoundly influences the thermal structural scale in the underlying mantle.

Studies modelling systems with dimensions comparable to the circumference of the Earth and multiple evolving plates have been rare. In a convection model featuring a stiff lithosphere obtained by modelling a temperature-dependent rheology, Puster et al. (1995) obtained dynamically determined plate motion by specifying mobile weak zones (plate boundaries). These authors performed mantle convection experiments in an annulus with plates whose geometries evolved with time and found that the surface plates organized mantle flow, resulting in large-scale structures and a thermal angular power spectrum dominated by low angular degrees. In agreement with lower-mantle viscosity estimates obtained to fit geodynamic constraints using two-layer parametrizations (e.g. Hager 1984; Forte & Peltier 1987; Ricard & Vigny 1989), these authors also found that a 30-fold increase in lower-mantle viscosity resulted in plate sizes and velocities that were most consistent with values inferred from the geological record.

Specifying weak plate boundaries in a model with a stiff lithosphere (e.g. Gurnis 1989; King & Hager 1990) is one of several methods that have been used to obtain plate-like dynamically determined surface motion in mantle convection models. King et al. (1992) dubbed this modelling approach the material property method for modelling plates and showed that for steady-state scenarios it yields fundamentally equivalent convection solutions to an alternative method of obtaining plate-like surface motion, the so-called force balance method (Gable et al. 1991). Koglin Jr. et al. (2005) have shown that, with fixed geometries, the two plate modelling methods result in similarly behaving solutions, even for highly time-dependent flows. In this study, we employ the force-balance method for obtaining dynamically determined plate motion and introduce a new modelling approach that allows for obtaining dynamic plate boundary evolution featuring one-sided subduction and symmetric seafloor spreading. The study presented utilizes a 2-D numerical mantle convection model featuring multiple plates with dynamically determined velocities and dynamically determined plate boundary velocities to explore the feedback between an evolving system of plates and vigorous mantle convection. The experiments specify both internal and basal heating and a pressure-dependent viscosity. In addition, we are able to model time-dependent plate thickness by implementing a time-dependent viscosity that varies as a function of the horizontally averaged mean temperature. (Plate thickness does not vary laterally at a given time.) Our model also allows for the number of plates in a calculation to evolve, both as a result of plate consumption and plate rifting.

2 MODEL DESCRIPTION

We simulate convection in the Earth’s mantle by modelling infinite Prandtl number thermal convection in an incompressible Boussinesq fluid with a Newtonian rheology. Following the numerical approach of Gable et al. (1991) we solve the equations for the conservation of mass, momentum and energy in a 2-D Cartesian geometry using the hybrid spectral, finite-difference code MC3D.

Convective vigour in the calculations is determined by the Bénard-Rayleigh number, \(Ra_B\), (e.g. Chandrasekhar 1961) and the internal heating rate, \(H\). The full set of equations governing flow in these systems was previously described by Lowman et al. (2003). However, here we model geochemically homogeneous convection by implementing a depth-dependent internal heating rate. The layered heating simulates the reduction in volume of a finite thickness spherical shell as a function of radius. Specifically, we compare a spherical-shell segment and a Cartesian layer with the same thickness, \(d\), and equate the total heat production rate and the heat production rate at the surface in order to obtain a depth-dependent heat source distribution in the Cartesian case (Nettelfield & Lowman 2007).

2.1 Viscosity and plate thickness

All of the calculations presented feature a non-dimensional depth-dependent viscosity \(\eta_r(z)\) characterized by an upper-mantle viscosity of unity and an increase in lower-mantle viscosity with depth that follows a logarithmic trend. Rather than adopting any specific inferred viscosity profile for the mantle, we specify a depth-dependent viscosity that satisfies the widely accepted view that mantle viscosity increases sharply at the base of the transition zone, and that the total increase in viscosity is a factor of 30 (Hager 1984; Forte & Peltier 1987; Ricard & Vigny 1989).

Thus we set:

\[
\eta_r(z) = 1, \quad 0.769 \leq z \leq 1.00; \\
\eta_r(z) = 1 + \frac{29}{\log(100)} \log \left[ 1 + 99 \left( \frac{0.769 - z}{0.769} \right) \right], \quad 0 \leq z < 0.769, \\
\eta_r(z = 0) = 30.
\]

(1)

where \(z\) is the non-dimensional height above the base of the system. [Note \(\eta_r(z = 0) = 30.\)]

In an initial set of calculations we specify different uniform thickness plates where the plate viscosity is a factor of 1000 greater than the viscosity in the upper mantle. In the remaining calculations, we model a time-dependent viscosity by specifying a viscosity–depth relation that is dependent upon both mean temperature as
a function of depth and the pressure effect described by eq. (1). Previous work has shown that specifying a depth-dependent viscosity controlled by horizontally averaged temperature results in obtaining a good approximation for the mean temperature and heat flow observed in a fully temperature-dependent viscosity calculation (Sunder-Plassmann & Christensen 2000).

We calculate the temperature-dependence of viscosity based on an Arrhenius law (e.g. Ratcliff et al. 1998) given by:

$$\eta_T(z) = C_M \exp\left(\frac{1}{1 + \tilde{T}(z)} - 1\right) 2 \ln C_C. \quad (2)$$

where $\tilde{T}(z)$ is the average temperature at height $z$. $C_M$ is the magnitude of the non-dimensional viscosity at $\tilde{T} = 0$, and $2 \ln C_C$ is an activation parameter, so that $C_C$ determines the contrast in viscosity between fluid with a non-dimensional temperature of $\tilde{T} = 0$ (corresponding to a maximum viscosity) and $\tilde{T} = 1$ (corresponding to a minimum viscosity). Specifically, $C_M/C_C$ is the magnitude of the non-dimensional viscosity when $\tilde{T} = 1$.

In all of the calculations presented, the convecting layer is confined between isothermal horizontal boundaries. We specify $C_M = 1000$ in eq. (2) to give a non-dimensional viscosity value of 1000 at the model surface. We choose the contrast parameter $C_C$ to give non-dimensional upper-mantle viscosities approximately equal to 1. Consequently, $C_C$ varies between models that have different average temperatures; a higher average temperature will require a lower value of $C_C$, and vice versa. We model both the temperature-dependence and pressure-dependence of the rheology by multiplying eq. (2) by the depth-dependent function given in eq. (1).

We choose a non-dimensional viscosity value $\eta_n$ in order to define the depth at which the transition occurs from lithospheric plate to asthenospheric upper mantle. As the mean horizontal temperature and, therefore, viscosity stratification change over time, the depth at which this transition occurs must also change. Consequently, we obtain a time-dependent mean plate thickness. Our definition of the plate thickness is used to determine plate velocities as described below.

2.2 Plate evolution

We model multiple distinct plates with piecewise uniform surface velocities separated by narrow boundaries (e.g. Puster et al. 1995; Bunge & Richards 1996; Lowman et al. 2001; Monnereau & Quéré 2001; Quéré & Forte 2006). The plate geometry is known at all times so that the shear tractions acting on the base of any plate can be determined at any instant. The plate velocities are continually updated so that the integrated shear stress on the base of each plate is zero at all times. Consequently, plate motion is determined so that it neither adds nor subtracts energy from the system. As a result, we obtain dynamically determined plate velocities that reflect the time-dependent buoyancy distribution both within and beneath the plate at any instant (e.g. Davies 1986, 1989; Gurnis 1989; King et al. 1992; Puster et al. 1995; Zhong et al. 2000; Lowman et al. 2001; Monnereau & Quéré 2001; Quéré & Forte 2006). We do not attempt to model a chemically distinct continental lithosphere.

We utilize the evolving plate velocities to update plate boundary positions in calculations that model dynamically determined plate boundary evolution. For divergent boundaries we require the plate boundary to move with the average value of the velocities of the two neighbouring plates, thus simulating symmetric seafloor spreading (Vine & Matthews 1963; Morley & Larochelle 1964), and assume that new plate material introduced at divergent boundaries has an age of zero. For convergent boundaries, we determine the motion of the plate boundary based on the velocity of the youngest plate adjacent to the boundary.

Choosing to have the convergent plate boundary motion mirror the motion of the leading edge of the youngest plate is equivalent to relocating the boundary so that it moves with the velocity of the leading edge of the more buoyant plate. Alternatively, the condition may be viewed as one which forces the denser plate to subduct. To ensure this condition is satisfied we utilize the velocity history of the plates to determine and track age as a function of position at the surface of our calculations. Surface age is continually updated based upon the plate velocities at each time step. Our modelling approach ignores any forces associated with the collision of the edge of the plates and any effects of elasticity that would resist bending of the subducted plate (as would occur in the Earth). Moreover, we ignore subduction zone stress history. While recognizing that convergent plate boundary motion on the Earth is, therefore, a more complex process than is modelled here, we argue that by specifying that horizontal density gradients should implicitly control convergent plate boundary motion, we are determining boundary motion with the parameter that drives the convection itself.

At each time step we first update the plate geometry based on the plate velocities calculated at the previous time step. Subsequently, we recalculate a plate interaction coefficient matrix in order to determine the stresses induced on the mantle by plate motion (Gable et al. 1991). In total, the number of solutions of the flow equations at each time step is equal to the number of plates in the system, plus two. Implementing the same approach in a 3-D Cartesian system would require solving the flow equation $2n + 2$ times per time step, where $n$ is the number of plates. Specifying a time-dependent viscosity requires an equivalent number of multiple solutions to the flow equations, even for models with fixed boundary locations.

Plates may be eliminated from a calculation if they are subducted at one (or both) end(s) more quickly than they may be growing at another boundary. Prior to its ultimate consumption a plate becomes very small as it is consumed at a subduction zone. In such cases we often find that the stresses acting on the model plates may become very large; sometimes implying an order of magnitude or more increase in velocity. It is unlikely a very small plate on the Earth would exhibit such high velocities while maintaining uniform motion and rigidity. In order to avoid the numerical difficulties associated with trying to resolve plates as they narrow to widths of the order of the grid resolution, we eliminate shrinking plates once their width becomes less than 1 per cent of the width of the solution domain (i.e. about 400 km across). When this occurs, each neighbouring plate gains a fraction of the surface area previously covered by the formerly intervening small plate. The fraction gained is in proportion to the ratio of the sizes of the two neighbouring plates.

We introduce new plate boundaries by breaking a plate in two at any location where a predefined yield stress, $\tau_y$, is exceeded within the plate interior. To determine internal stresses we treat the plates as finite thickness blocks and calculate the horizontal forces on vertical planes within the plate. The horizontal force at a given location $x$ is equal to $\tau_x(x) \times d_{pl}$, where $\tau_x(x)$ is the mean normal deviatoric tensile stress at $x$ and $d_{pl}$ is the plate thickness. The horizontal forces arise from shear stress exerted on the base of the lithosphere. Thus, $\tau_x(x)$ in the plate interior can be determined by integrating the shear stresses from the location of the nearest plate boundary to the left (denoted $x_0$), to the position of interest, $x$, yielding

$$\tau_x(x) = \int_{x_0}^{x} \tau_x \, dx \, d_{pl}. \quad (3)$$
(Note that because the net stress on each plate is zero, integrating the stress from the nearest plate boundary to the right gives a value for $\tau_{yy}$ that is equal in magnitude but opposite in sign to the value defined in eq. 3). If $\tau_{yy}(x)$ corresponds to a tensile deviatoric stress greater in magnitude than $\tau_y$ then we specify that the plate brakes at the position $x$.

We specify a lithospheric yield stress, $\tau_y$, of 80 MPa based on the stress values observed in earlier mantle convection experiments (e.g. Gurnis 1988; Lowman & Jarvis 1999). However, estimates of lithospheric strength given by Kohlstedt et al. (1995) indicate that although this value may be appropriate for continental lithosphere it is likely too low for oceanic lithosphere. We shall discuss our choice further in our conclusions.

All of our calculations begin from a statistically steady initial condition obtained with a static plate geometry and satisfy the requirement that, given the experimental parameters, the model has evolved to a point where it does not exhibit any long-term heating or cooling. For a model that will feature evolving plates, the surface velocity history obtained during the calculation of the initial temperature field is used to determine the initial age of the surface.

3 Results

The calculations we focus on in this study fall into two categories: convection with a pressure-dependent viscosity, $\eta_P(z)$, and constant plate thickness (which we denote case D) and convection with a time-dependent viscosity $\eta(z) = \eta_P(z) \times \eta_T(z)$ and time-dependent plate thickness (denoted case TD). We indicate models with plate geometry evolution by adding a lowercase ‘e’ to the model name and distinguish our models with names of the form ($w$,$x$,$y$,$z$),$T$,$D$,$w$,$x$,$y$,$z$,$e$,$fi$,$re$,$fl$,$Y$,$x$), where $w$, $x$, $y$, and $z$ are numbers that refer to the non-dimensional width of each of the four plates featured in the calculation. In the case of an evolving geometry, these numbers refer to the initial condition of the plate geometry.

We first examine calculations featuring a single plate. We refer to the models as Model D1 and Model TD1, indicating a single plate with a width of $1\bar{d}$.

In the time-dependent viscosity calculations, we choose the values of the constant $C_C$ in order to obtain a non-dimensional viscosity value of approximately unity in the upper mantle. In aspect ratio 1 calculations, this means using a value of $C_C = 8000$, and in aspect ratio 12 calculations, $C_C = 5000$. In all such calculations, we use $C_M = 1000$ to give a non-dimensional viscosity of 1000 at the surface.

In each calculation, we specify a Rayleigh number of $Ra_B = 10^8$ (based on the viscosity value in the upper mantle) and a depth-dependent internal heating function $H(z)$, satisfying the spherical shell distribution condition described in the previous section, where $H(z = d) = 15$. This value is consistent with the estimated heating rate for the bulk silicate Earth inferred from chondritic meteorites (Stacey 1992). We specify 200 uniformly spaced vertical cells in each calculation; for aspect ratio 1 calculations there are 200 horizontal cells and for aspect ratio 12 calculations there are 2592 horizontal cells.

For stratified internal heating with a rate set at $H = 15$ at the surface of the system, the qualitative aspects of model evolution vary by very little in comparison to the behaviour observed in uniformly heat systems. Specifying $H = 9.22$ produces the same amount of total heating with a uniform distribution as we have prescribed with a stratified distribution. For the Rayleigh number here and $H$ in the range 5–15, Lowman et al. (2003) reported qualitatively similar time-dependent behaviour in convection models featuring plates and a stratified viscosity. However, these authors did not note our finding that the period of cyclic time-dependent behaviour increases with internal heating rate.

To gain insight into the influence of the parameters controlling time-dependent behaviour in our model, we first consider calculations that isolate the influence of plate thickness on flow time-dependence.

3.1 Aspect ratio 1 models

The first suite of calculations that we present have unit aspect ratio reflecting side walls and a single plate. Previous work (Lowman et al. 2001; Koglin Jr. et al. 2005) has shown that the combination of moderate internal heating rates and finite thickness plates results in frequent mantle flow reversals, where the mantle circulation and plate velocity regularly reverse direction. The authors showed that this behaviour is a robust feature in systems featuring stationary plate boundaries. The parameters controlling the regularity and onset of this phenomenon were shown to be the Rayleigh number and the internal heating rate (Lowman et al. 2003). In this study, we keep the viscosity model and heating mode fixed while varying plate thickness, in order to isolate the effect of the latter parameter on plate velocity time-dependence.

Fig. 1(a) shows representative out-takes from plate velocity time-series obtained for four different D1-type (i.e. constant plate thickness) calculations over a period of 0.005 diffusion times. The four

![Figure 1](image-url)
time-series correspond to calculations with plate thicknesses of 0.07d, 0.05d, 0.03d and 0d. In each case, the behaviour shown during the time period depicted in the figure is typical of the time-series obtained for the entire calculation. (Each calculation was performed for 0.1 diffusion times and reaches the thermally equilibrated state corresponding to the curves shown in less than half of that time.) In the case with a plate thickness of 0.05d (145 km, medium-dashed line) we obtain regular flow reversals. Our model behaves similarly to the case presented by Lowman et al. (2001) despite a slightly amended viscosity profile and the change to a depth-dependent internal heating rate. The principal difference in the result presented here and those of the previous study is a change in the period of the reversals. Increasing the plate thickness to 0.07d (203 km) does not change the regular nature of the flow reversals but slightly increases the periodicity. The thicker plate takes longer to reverse, and the peak values of velocity are also slightly smaller than in the 0.05d plate thickness case. However, significantly decreasing the plate thickness removes the regularity of reversals. At a thickness of 0.03d (87 km), reversals do occasionally take place, but they are interspersed with long periods in which no reversals take place. In these instances, the plate velocity occasionally decreases to almost zero but then returns to higher values without the occurrence of a change in direction. When a plate thickness of zero is specified no reversals occur.

We investigated flow plate velocity in unit aspect ratio calculations with plate thicknesses of \(n \times 0.05d\) for integer values of \(n\) between 0 and 14. We find the transition between regular and irregular reversals occurs between 0.035d and 0.040d (101.5 and 116 km, assuming a 2900 km thick system), and that there is another transition between models featuring irregular reversals and no reversals between 0.015d and 0.020d (43.5 and 58.0 km). We also find that as the plate thickness increases, the global average temperature increases. Fig. 1(b) shows the globally averaged temperature as a function of plate thickness in the calculations described above. Temperature increases monotonically with plate thickness so that the calculation featuring the thickest plate is approximately 33 per cent hotter than the calculation featuring no plate. A sharp discontinuity in the temperature trend occurs with the onset of periodic flow reversals. The datum point indicated with a circle corresponds to a calculation featuring a variable thickness plate, Model TD1 (discussed below).

The increase in temperature with plate thickness is explained by the fact that a thick viscous plate inhibits heat from reaching the surface of the system by advection. In addition, the thick plate suppresses the formation of boundary layer instabilities which can break away from the top of the system to cool the fluid interior. Consequently, a thick plate traps heat in the convecting system. The onset of periodic flow reversals acts to elevate temperatures even further because the reversals entail the surface velocity periodically dropping to zero and, therefore, retarding the flow of heat to the surface by advection.

Fig. 2 shows snapshots of the horizontally averaged temperature profiles in the upper part of the four D1-type calculations introduced in Fig. 1(a). The horizontal lines show the depth of the base of the plate in each case. The figures indicate that regular flow reversals occur when the mean temperature at the base of the plate is comparable to or greater than the temperatures in the upper mantle. That is, cases where the base of the viscous plate is at the bottom of the cold thermal boundary layer. When the base of the plate is placed inside the thermal boundary layer, flow reversals become less common and in the case of very thin plates, entirely non-existent. This is because the energy source that drives flow reversals forms well below the thermal boundary layer. Internally heated convection in a system featuring plates is characterized by a thermal structure in the bulk of the fluid between the horizontal thermal boundary layers that would drive a flow in the opposite direction to the motion driven by the cold boundary layer and downwellings. Specifically, at high \(Ra\) and internal heating rates, relatively broad regions filled with anomalously hot fluid form around the cold downwellings in the system. This buoyant region resists the downward pull of the sinking cold material and builds up around the downwellings, well below the upper thermal boundary layer. The head of the buoyant region extends into the shallow mantle where it sits over a column of warm material. Depths below the plates but above the hot envelope experience a strong push in an opposite sense of direction to the plate motion. Because shear stresses acting on the base of the viscous plate determine plate motion, flow reversals are only observed in cases where the base of the plate resides in the region where stresses associated with the buoyant envelope are at their greatest. As shown in Fig. 1, the period of regularly occurring flow reversals is sensitive to the depth at which the base of the plates is specified. We have further investigated the influence of plate thickness on reversal behaviour by performing several calculations with plates having thicknesses between 0.1d and 0.2d. We find that flow reversals persist in this range with the heating conditions that we have specified. However, the frequency of the flow reversals diminishes as the plate thickness is increased.

We have tested whether much higher rates of internal heating affect flow reversal frequency by affecting the mean temperature of the system. We find that flow reversals persist in calculations with internal heating rates up to three times greater than that we employ in the models presented in this paper. However, system temperatures increase dramatically as internal heating is increased and we find that flow reversal frequency decreases and becomes less regular with very high internal heating rates.

The extreme variations in plate velocity during a flow reversal result in similarly large variations in mean thermal boundary thickness. By specifying a stratified viscosity based on mean temperature as a function of depth we are able to model the thickening and thinning of the viscous plate that occurs with large changes in its velocity. For example, when the plates are moving fast and the thermal boundary layer is thin the depth to the base of the plate
decreases and similarly when the plate is moving slowly and the thermal boundary layer thickens the depth to the base of the plate increases.

Fig. 3 shows viscosity profile snapshots taken at intervals of 0.01 diffusion times for Model TD1. Also shown (by the block curve) is the viscosity in Model D1. The depth-dependence of the viscosity in Model TD1 is determined by the horizontally averaged temperature, resulting in a time-dependent viscosity stratification and plate thickness. The shear stresses used to determine plate motion (and define the plate thickness) are calculated at the depth below the surface where the viscosity attains a value of unity. Due to the pooling of cold material associated with downwellings, the viscosity temperature-dependence leads to a higher viscosity layer in the lowermost part of the mantle compared with calculations that feature a solely pressure-dependent viscosity. However, the viscosity at the core–mantle boundary is less than in the pressure-dependent case; it has a non-dimensional value of 3.75 \((C_M/C_C) \times 30\) compared to 30 in the purely pressure-dependent case.

We find that the introduction of a time-dependent viscosity stratification and, therefore, variable plate thickness (Model TD1) affects reversal behaviour. Fig. 4 shows the time-series of the plate velocity and the corresponding plate thickness for Model TD1. Irregular flow reversals occur in this calculation and the variable plate thickness ranges between approximately 0.015d and 0.085d (approximately 20 and 245 km when scaled to Earth values), with an average of 0.0388d (approximately 110 km). In the range that includes this value, our calculations with constant thickness plates showed that a transition occurs from time-dependent solutions exhibiting intermittent reversal behaviour to solutions featuring regular (periodic) reversals. Our one solution obtained with a variable thickness plate is, therefore, consistent with what was observed in the previously described solutions.

Based on the findings shown in Fig. 2, we suggest that the irregular flow reversals in Model TD1 can be explained by the fact that the thermal boundary layer thins during periods characterized by rapid plate motion and requires some period of time to thicken again after such a period. Thinning affects the calculated plate thickness and, therefore, the stress calculation that determines plate velocity. The result is that less predictable flow reversals occur in this model.

![Figure 3](image3.png)

**Figure 3.** Viscosity profiles at intervals of 0.01 diffusion times in Model TD1. The black curve shows the dependence on depth of the fixed viscosity values used in Model D1.

![Figure 4](image4.png)

**Figure 4.** Model TD1: (a) plate velocity time-series and (b) plate thickness time-series. The constant plate thickness used in Model D1 (0.05d) is indicated by the dotted line in the bottom figure.

### 3.2 Aspect ratio 12 models

In the remainder of this study, we focus on aspect ratio 12 calculations with periodic sidewalls. We first consider cases with fixed geometries featuring four plates. Previous studies showed that time-dependent flow featuring plate reversals similar to those observed in the unit aspect ratio models can appear in wide domains but that the behaviour is dependent on the arrangement of the plates, their sizes, and the initial conditions (Lowman et al. 2001). Here, we vary plate geometry and viscosity profile to investigate their effect on plate velocity and heat flux time-dependence in multiple plate models. Models (T)D4422 and (T)D4.52.51.53.5 feature plates with widths of 4d, 4d, 2d and 2d and 4.5d, 2.5d, 1.5d and 3.5d, respectively.

Table 1 gives the average global temperatures of the models presented in the remainder of this paper. In general, the models featuring fixed thickness plates are about 20 per cent warmer than the models featuring variable thickness plates. For both the D and TD-type models

<table>
<thead>
<tr>
<th>Model</th>
<th>Plate thickness</th>
<th>Temperature</th>
<th>T (2nd moment)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D4.52.51.53.5</td>
<td>0.05</td>
<td>0.7694</td>
<td>3.50 \times 10^{-6}</td>
</tr>
<tr>
<td>D4422</td>
<td>0.05</td>
<td>0.7937</td>
<td>2.10 \times 10^{-7}</td>
</tr>
<tr>
<td>D4422e</td>
<td>0.05</td>
<td>0.7912</td>
<td>4.79 \times 10^{-6}</td>
</tr>
<tr>
<td>TD4.52.51.53.5</td>
<td>0.0475</td>
<td>0.6945</td>
<td>1.59 \times 10^{-6}</td>
</tr>
<tr>
<td>TD4422</td>
<td>0.0480</td>
<td>0.6596</td>
<td>1.67 \times 10^{-7}</td>
</tr>
<tr>
<td>TD4422e</td>
<td>0.0464</td>
<td>0.6628</td>
<td>1.12 \times 10^{-5}</td>
</tr>
</tbody>
</table>
the calculations featuring evolving plate boundaries are comparable in mean temperature to calculations that adopt steady flow patterns. Thus we find that evolving plate boundaries allow the mantle to cool with a similar efficiency to a system featuring fixed plate boundaries.

Table 1 also gives the second moment calculated for each temperature time-series. The second moment value is an indication of the variability of the parameter measured by the time-series. Higher values indicate greater variations. The values in Table 1 indicate that in the calculations featuring evolving plates, the globally averaged temperature varies more than in the calculations featuring fixed plate boundaries. Similarly the calculations with static plate boundaries featuring four different size plates vary more in temperature than the other static plate boundary cases that feature pairs of equal size plates.

### 3.2.1 Fixed plate boundary models

Fig. 5 shows a temperature field snapshot representative of the steady convection pattern obtained in Model D4422 (which we introduce here as a reference model). The flow features a pair of aspect ratio six convection cells separating a major downwelling and upwelling positioned below convergent and divergent plate boundaries, respectively. A similar model was presented in earlier work (Lowman et al. 2001). In that study an upwelling developed between the pair of width 4$d$ plates and the flow was characterized by intermittent flow reversals of the two plates with widths of 2$d$. However, the possibility of obtaining the solution presented here was also identified. In either case, the symmetric geometry results in steady symmetric convection patterns. Lowman et al. (2001) suggested that wide plates arranged in symmetric geometries settle into flow patterns that cannot reverse (as we find is the case here).

Fig. 6 shows a sequence of temperature field snapshots for Model D4.52.51.53.5. Each frame is labelled with a non-dimensional time corresponding to an instant in the plate velocity time-series shown in Fig. 7. This choice of plate geometry does not allow for a pair of symmetric convection cells to develop where both a major downwelling and upwelling are positioned at plate boundaries. These snapshots indicate the time-dependence of the calculation by showing the change in position of the downwelling due to episodic reorganizations of the surface velocities caused by reversals of the two smaller plates. However, the time period between reorganization events is not as regular as the flow reversal events observed in our aspect ratio 1 calculations. We also note that in the absence of the conditions necessary to develop into a symmetric flow the system becomes characterized not only by its time-dependence but a lack of organization of its upwellings compared with Model D4422. Numerous mobile upwellings are present in the lower mantle as the constant changes in position of the downwellings cause lower mantle stagnation points to shift positions.
Fig. 7 shows the plate velocity time-series for Models D4422 and D4.52.51.53.5. The total period shown in each case is roughly equivalent to 7.3 Gyr. The top panel shows the velocities in Model D4422 and indicates that while the velocities are highly time-dependent none of the plates changes direction during the entire period.

Model D4.52.51.53.5 features episodic changes in the direction of the velocity of the two smallest plates, which causes the flow pattern to change as the major downwelling changes position. Although the two largest plates do not change direction, they experience episodic changes in velocity, from almost zero to non-dimensional speeds close to 8000 (approximately 6 cm yr\(^{-1}\)) and vice versa. The greatest changes in velocity occur when a reversal is taking place; the change in direction of one plate directly affects the magnitude, and occasionally the direction, of the other three plates. Consequently, the mean surface velocity of the system is also very time-dependent (as discussed further below). However, compared with the mantle transit time, reversals occur infrequently; in dimensional terms the reversals take place at intervals separated by almost 900 Myr.

We return to calculations that feature a time-dependent stratified viscosity (and therefore, plate thickness) and examine the effects on time-dependence in aspect ratio 12 systems.

Fig. 8 shows a temperature field snapshot from Model TD4.52.51.53.5. In contrast to Model D4.52.51.53.5, Model TD4.52.51.53.5 adopts a remarkably steady flow pattern for a period comparable to the age of the Earth. The calculation is characterized by a pair of convection cells. As observed in Model D4422, Model TD4.52.51.53.5 features a single downwelling and the presence of the pressure-dependent viscosity appears to be the dominating factor in determining the system wavelength.

The aspect ratio 12 calculations featuring variable plate thickness are cooler than those with a fixed plate thickness (Table 1). The overall coolness of the system may result from the fact that the decreased viscosity of the lower part of the plate allows for heat to be advected closer to the surface at all locations. This effect becomes more significant as convection cell width is increased and was not clearly exhibited in the comparison of aspect ratio one calculations. A less viscous upper thermal boundary layer also allows for instabilities to form more readily at the surface of the system. When the resulting cold drips detach from the upper boundary layer they sink into the fluid interior and provide a means of cooling the entire system.

The top panel of Fig. 9 shows time-series of the plate velocities from Model TD4.52.51.53.5. The largest plate in the system does not change direction, although its velocity fluctuates in magnitude by up to 50 per cent. The three small plates balance the motion of the larger plate and have velocities that fluctuate in magnitude substantially, however, they only exhibit a single brief reversal cycle. The bottom panel in Fig. 9 shows the plate thickness in Model TD4.52.51.53.5 as a function of time. The fluctuations in the thickness of the plate in this calculation are much smaller than in Model TD1 and the mean value is close to the thickness specified in Model D4.52.51.53.5.

3.2.2 Evolving plate boundary models

We now consider an aspect ratio 12 calculation with a dynamically evolving plate geometry, Model D4422e. The calculation is started with the same initial plate geometry and temperature field (corresponding to $\tau = 0.000$ in the calculations) as Model D4422. The only difference between the calculations is the evolving plate boundaries. The time-dependent characteristics of the models, such as average surface velocity and the surface and basal heat fluxes are examined below.

Fig. 10 depicts a sequence of temperature fields and the corresponding surface age and plate boundary positions in Model D4422e. The period that elapses between the first and final frame in the figure is approximately 136 Myr. Like Model D4.52.51.53.5, Model D4422e is extremely time-dependent. Both models feature shifting downwellings and a variety of transient upwellings. Like the other D-type models, Model D4422e exhibits particularly vigorous downwellings due to its high mean temperature. In addition, it is characterized by very long wavelength flow, and typically features only one major downwelling. Thus it appears that even with the presence of multiple evolving plate boundaries, a significant increase in lower-mantle viscosity relative to upper-mantle viscosity strongly influences the convective wavelength of this plate–mantle system. Moreover, we observe in a qualitative sense that convergent plate boundary motion has the dominant effect on convective flow patterns and that upwelling positions appear relatively unaffected by the motion of divergent plate boundaries.
The surface age plots included in Fig. 10 show several examples of symmetric seafloor spreading associated with the presence of passive divergent plate boundaries (i.e. boundaries not associated with active mantle upwellings). The tic marks on the vertical age axes correspond to increments of 0.001 non-dimensional time units (i.e. 366 Myr). Thus, the surface ages we obtain are much higher than the mean age of the oceanic plates. This is consistent with the mean surface velocities of approximately 1.5 cm yr\(^{-1}\) that prevailed for a long period prior to the sequence shown and then increased to an average of over 3.5 cm yr\(^{-1}\) during the period featuring the plate breakup events. The low surface velocities were observed during a period featuring only two plates. A sustained period featuring large plates leads to an increase in convective cell wavelength that results in increased temperature and reduced velocities (e.g. Lowman et al. 2001; Grigné et al. 2005).

The sequence shown in Fig. 10 features two examples of plate breakup events. A breakup occurs during the intervening period between the timing of the top two panels in the figure; the plate bounded between \(x \approx 4.4d\) and \(8.4d\) breaks at \(x \approx 7.7d\) close to the major downwelling (see the second panel of the figure). The flow around the downwelling in the figure (which is formed by the merger of two downwellings) is much more vigorous than anywhere else in the calculation, and the small plates in the vicinity of the downwelling have much larger plate velocities than the larger plates. However, flow below the large plates close to the downwelling is also much more vigorous than the flow far from the convergent plate boundary. Consequently, the magnitude of the tractions along the base of the plate increases rapidly as the convergent boundary is approached. The plate yield stress is exceeded because the tractions at the end of the plate nearest the boundary are particularly large even though they are confined to a relatively small area. This phenomenon can cause sections of plate to break-off of the end of a large plate when particularly vigorous downwellings develop, as in this sequence.

The breakup between the third and fourth panels in Fig. 10 occurs at \(x \approx 2.4d\). This breakup is somewhat different to the case described above in that a large plate breaks at a location far from any major downwelling or upwelling. Breakup occurs because the plate is constrained to move at a constant velocity which is significantly greater than the fluid below it at one end and substantially less than the fluid below at the other end. Because the plate is large, the integrated effect of the tractions associated with the fast and slow moving upper-mantle flow is enough to overcome the yield stress in the plate interior. Consequently, the plate breaks above the point where the subplate flow makes the transition from moving slower than the plate to moving faster.

The panels in Fig. 11 show the stress conditions in Model D4422c at the time of the two breakup events described above. The figure shows the non-dimensional shear stress, \(\tau_{xz}\), at the base of the plates (dashed curve) as well as the mean horizontal deviatoric normal stress magnitude (vertical solid bars) in the plates. The latter quantity is only shown at locations where its magnitude is a local maximum. Positive values represent tensile stress and negative values compressional stress. The horizontal dotted line indicates the non-dimensional yield stress value. The normal stresses and yield stress have been divided by a factor of five relative to the shear stresses in order to improve the figure quality (i.e. vertically exaggerate the shear stresses relative to the normal stresses). Vertical arrows indicate the location of plate boundaries. Negative shear stress values indicate that upper-mantle motion is pulling the plate in the positive \(x\) direction while positive shear stress values indicate mantle flow is...
pulling the plate in the negative \(x\) direction. The breakup events described in the discussion of Fig. 10 occur when the deviatoric normal stress magnitude exceeds the yield stress, \(\tau_y\). Fig. 11(b) shows that there were several locations where the large plate in the calculation had become at risk of breaking.

Fig. 12 compares the mean surface velocity and heat flux time-series from each of the aspect ratio 12 D-type models examined. Also shown in the top panel is a time-series indicating the total number of plates at any given instant in Model D4422e. In dimensional terms these time-series cover almost 1.5 Gyr. During the time period shown Model D4422e features 33 breakup events, however, about half of these breakup events are examples of a small segment breaking off of the end of a larger plate only to subsequently be consumed. The total number of plates present in the model at any one time varies between 2 and 7.

The top panel indicates very different evolution in Model D4422e in the final 0.001 units of the time period examined compared with the preceding 0.003 diffusion times. We suggest that the initial period is atypical and corresponds to a period in which the model’s evolution is dominated by the initial long wavelength convection pattern of Model D4422. As mentioned above, during the middle of the calculation the model is dominated by a pair of large plates. However, shorter wavelength features eventually develop once the large plates breakup.

As expected, the average surface velocity and surface heat flux are strongly correlated. Model D4422 exhibits a surface heat flux and mean plate velocity that fluctuate only minimally during the entire period examined. Model D4.52.51.53.5 exhibits periodic increases in mean surface velocity and surface heat flux of up to 100 and 50 per cent percent, respectively. The fluctuations are associated with reversals of the small plates in the model. Basal heat flux variations of approximately 50 per cent are also observed due to the influence of plate reorganization events. The mean surface velocity and heat fluxes also vary considerably in Model D4422e. In addition, the plate velocities in Model D4422e fluctuate much more rapidly than in the cases with static plate boundaries, as do the surface and basal heat fluxes. The largest variations in surface velocity and heat flux coincide with occasions where plates are either consumed or created. In both Model D4.52.51.53.5 and Model D4422e basal heat flux increases by up to 50 per cent with the arrival of new cold material at the bottom of the system after the opening of a new subduction zone (or the rapid migration of an existing subduction zone). Heat flux varies particularly rapidly in Model D4422e as plate boundary motion allows cold material to spread over large regions of the bottom boundary and numerous unorganized upwellings appear which thin the surrounding basal thermal boundary layer, replacing the stabilizing influence on basal heat flux of a single focussed upwelling. During periods in which it features at least four plates, Model D4422e delivers significantly greater surface heat flow than Model D4422. Our findings indicate that, in general, time-dependent convection patterns produce higher mean surface heat fluxes.

Although TD-type models exhibit much steadier convection patterns than D-type models in cases with static plate geometries, we find that once plate boundary evolution is enabled TD-type models are as time-dependent as D models. Fig. 13 shows a sequence of temperature field snapshots, and corresponding surface age plots and plate boundary positions from Model TD4422e. Non-dimensional times corresponding to the snapshots of the thermal fields are shown below each panel. In general, Model TD4422e features one or two convergent plate boundaries and several passive divergent plate boundaries. The high viscosity lower mantle results in a long wavelength flow pattern but the temperature-dependence of the viscosity allows for the formation of numerous instabilities in the lower thermal boundary layer. The downwellings in Model TD4422e are less vigorous overall than those in Model D4422e because the former model is cooler than the latter. In addition, the plates in Model TD4422e exhibit higher velocities (and accordingly, younger ages) than the plates in Model D4422e. The result is that the downwelling morphology in Model TD4422e is quite distinct from that exhibited by the downwellings observed in Fig. 10. When a convergent plate boundary in these calculations is stationary, a supply of cold downwelling material descends without interruption and a smooth continuous downwelling is obtained. If the convergent plate boundary is in rapid motion, the cold sinking material breaks away from the upper thermal boundary layer in distinct drips that sink in a chain of cold blobs. The downwellings in Model TD4422e are typically of the second type because of the larger velocities exhibited in that model.

Plate breakup events provide illustrative examples of model evolution. One of the plates featured in the sequence shown in Fig. 13 rifts at \(x \approx 9.5d\), during the intervening period between the top two panels shown in the figure. In contrast to the rifing cases observed in Model D4422e this breakup is triggered by an upwelling. Indeed, because Model TD4422e is cooler than Model D4422e it features more vigorous upwellings. Consequently, it exhibits examples of plate rifting caused by flow associated with active upwellings. At

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**Figure 12.** The number of plates present in Model D4422e as a function of time (a); corresponding time-series of average surface velocity (b), surface heat flux (c) and basal heat flux (d) from Models D4422, D4.52.51.53.5 and D4422e.
$x \approx 7.4d$ a different sort of breakup event occurs between the bottom two panels in Fig. 13. This event is more similar to the rifting observed in Model D4422c. As the large plate confined between $x \approx 3.9d$ and $9.6d$ grows it extends to cover regions of the underlying mantle moving at very different velocities. Differences in velocity between uniform velocity plates and the underlying mantle exert shear stresses on the base of the plate. Even moderate magnitude tractions can integrate to produce very large internal stresses in the plate when they are exerted over a large area. Consequently, very large plates are susceptible to breaking. In this case, the plate is approximately 17 000 km wide at the time it breaks.

At the times corresponding to the two breakup events described in the preceding discussion of Model TD4422e, Fig. 14 shows the non-dimensional shear stress at the base of the plate (dashed curve) and the magnitude of the mean deviatoric normal stress in the horizontal direction at locations where the magnitude attains local maximums (vertical bars). Positive normal stresses represent tensile stress and negative normal stresses are compressional. As in Fig. 11 the normal stresses have been divided by a factor of five in this figure, relative to the shear stresses, in order to improve the figure quality. Vertical arrows show the position of the plate boundary positions. The horizontal dashed line indicates the non-dimensional yield stress value (also reduced by a factor of 5). Significant shear stress associated with the vigorous upwelling observed in the third panel of Fig. 13 is evident in the top panel of Fig. 14 and the vertical solid bar marking the stress maximum is clearly greater in magnitude at this location than at any other local stress maximums (marked by other vertical bars). The bottom panel of Fig. 14 indicates that there are a number of locations in the interior of the large central plate where interior stresses are very close to the yield stress indicating numerous potential rifting points in the plate as it grows to cover a very large area.

Figs 15 and 16 present the same information as Fig. 12 but for the TD-type models that we have examined. We present two separate figures in order to preserve the same horizontal scale as in Fig. 12 and facilitate a straightforward comparison of the different time-series. Again, there appears to be a period of adjustment in the evolving plate case, Model TD4422e, that lasts about 0.003 diffusion times (Fig. 15). After this period the model is characterized by more frequent plate consumption and rifting, resulting in sharp changes in the mean global plate velocity and a large degree of variability in the global surface heat flux. The total number of breakup events in Model TD4422e is 17 over the final 1.8 Gyr of the evolution of the model, and the total number of plates in the calculation ranges between 2 and 7.

Fig. 16 compares the evolution of three TD-type models after several mantle overturns. As expected, Models TD4422 and TD4.52.51.53.5 show relatively uniform and similar mean surface velocities. In contrast, the evolving boundary case, Model TD4422e, shows fluctuations in mean global velocity of more than a factor of three. The mean global heat flux is not as variable as the surface velocity but clearly fluctuates by a much greater amount with the introduction of evolving plate boundaries. In this case, the maximum surface heat flux observed exceeds the minimum heat flux by over 50 per cent.

In contrast to the surface heat flux, the basal heat flux does not vary significantly between the three calculations. The maximum basal heat flux observed in the case with evolving boundaries is approximately 18 per cent greater than the minimum flux observed. This is actually less than in Model TD4422. The most likely explanation for the small variation in the basal heat flux compared with the surface heat flux is that the downwellings in the TD-type calculations...
Figure 15. The number of plates present in Model TD4422e as a function of time for $0 \leq t \leq 0.004$ (a); corresponding time-series of average surface velocity (b), surface heat flux (c) and basal heat flux (d) from Models TD4422, TD4.5, 2.5, 1.5, 3.5 and TD4422e.

4 DISCUSSION AND CONCLUSIONS

Previous studies of internally heated mantle convection featuring plates with fixed geometries identified the occurrence of periodic mantle flow reversals (Lowman et al. 2001; Koglin Jr. et al. 2005). Our findings show that the flow reversal periodicity is dependent on both the fixity of the plate geometry and the plate thickness. Periodic flow reversals are not obtained unless the viscous plates in the system are at least comparable in thickness to the thickness of the upper thermal boundary layer. The presence of flow reversals is dependent on plate thickness, size and geometry, as well as Rayleigh number and internal heating rate (Lowman et al. 2003). Flow reversals occur in systems featuring single plates (with reflecting side boundaries) and multiple plates (with periodic side boundaries). In the systems studied, the reversal of one plate is reflected in the motion of the other plates due to the coupled motion of all of the plates. However, reversals of single plates are possible without simultaneous reversals of any other plates.

We find that neither flow reversals of the type described above nor any other temporally periodic flow pattern behaviour is exhibited in systems with evolving plate boundaries. Rather, the inclusion of evolving plate boundaries in an internally heated mantle convection model results in a highly unpredictable time-dependent system (specifically, exhibiting shorter period variability in the convective pattern, mean surface velocity and surface heat flux) in comparison with systems that include static plate boundaries. The first of two types of models examined (D-type) includes temporally invariant plate thickness and exhibits flow reversals in both aspect ratio one, single plate calculations and aspect ratio 12, multiple plate systems featuring non-symmetric plate geometries. The second of the model types (TD-type) examined here includes time-dependent plate thickness and a time-dependent mantle viscosity stratification and is not conducive to obtaining flow reversals, even in static geometry cases. Despite promoting fundamentally different behaviour in systems that feature specified fixed plate geometries, the D- and TD-type rheologies both allow very similar behaviour when plate boundary motion is enabled. Plate boundary motion introduces a new level of time-dependence into the mantle system and results in more frequent fluctuations in boundary heat flux than are observed in systems where plate geometry is fixed. We suggest that this result is not model specific and that as a result of plate boundary
motion, the presence of symmetric seafloor spreading and one-sided subduction will increase the time-dependence of the thermal characteristics in any coupled plate–mantle system.

In our study, basal heat flux fluctuations show a wide range of variation (like surface heat flux fluctuations) and are clearly affected by the vigour of the system downwelling. The basal heat fluxes in Models D4.52.51.53.5 and D4422e increase by over 50 per cent as they climb from a minimum value to a maximum. However, after an initial period of adjustment (see Fig. 15) the basal heat flux in our Model TD4422e does not fluctuate any more significantly than the basal flux in calculations featuring steady convection patterns; Model TD4422 and TD4.52.51.53.5 (Fig. 16). Moreover, the basal heat flux maximum never exceeds the minimum by more than 25 per cent in any of these models. Surface heat flux increases by more than 50 per cent in both the D- and TD-type calculations.

The D-type models exhibit greater fluctuations in basal heat flux because the downwelling in these models are both colder relative to the ambient mantle and more focussed (due to their slower motion) than the downwelling in the TD-type models. Consequently, when new downwellings arrive at the base of these systems they deposit a particularly concentrated cold blanket on the hot bottom boundary so that the basal heat flux rises rapidly. The TD-type models do not exhibit flow reversals (in the static geometry cases) nor sluggish plate boundary motion and so the downwelling material in these calculations does not arrive at the bottom of these models either intermenttently nor in concentrated parcels. Moreover, a consequence of the rapid plate boundary motion in these calculations is that the downwellings breakup into numerous small parcels that descend more slowly (and therefore, warm-up) in comparison with a single large sinking parcel.

Although we are confident that our calculations that feature evolving plate boundaries better simulate evolution of the mantle than calculations in which boundary evolution is absent, it is debatable whether D- or TD-type models better reproduce mantle conditions. The D-type models allow colder more vigorous downwellings to develop, as might be obtained with a fully temperature-dependent viscosity. However, the constant plate thickness in these models holds heat in the mantle and, therefore, produces stronger temperature contrasts between subducted material and the ambient mantle than is expected in the Earth. In addition, the requirement of a temporally constant mechanical lithosphere thickness is an idealization. Consequently, we suggest that of the cases presented here, our Model TD4422e is the most applicable model to the Earth but that variations in basal heat flow approaching the magnitudes observed in Model D4422e may be more Earth-like. Future studies featuring a fully temperature-dependent rheology will be required to settle this issue.

The time-dependent convection patterns in our calculations that include evolving plate boundaries at first appear similar to the rapidly changing convection patterns found in systems featuring free-slip surface conditions (e.g. Weinstein et al. 1989; Hansen et al. 1992). However, internally heated convection in a stratified viscosity system like the Earth has been shown to adopt very long wavelength, stable convection patterns in the absence of plates (e.g. Hansen et al. 1993; Bunge et al. 1996, 1997). In addition, the systems modelled here differ from free-slip surface systems due to the insulating effect of the thick plates, which stabilize the upper thermal boundary layer and inhibit heat from reaching the surface by advection. By trapping heat and hence pockets of buoyant material in the system interior, plates produce time-dependent flow featuring dramatic events.

(When plate boundary motion is prohibited the most dramatic of events is manifested, namely, a flow reversal.) When boundaries are permitted to move plate reversals are not observed but periods of rapid increases and decreases in global plate velocity occur. Rapid increases in velocity are driven mostly by downwelling evolution. However, plates can rapidly decelerate as they override anomalously hot regions of the mantle, where flow is driven counter to the direction of plate motion.

One surprising finding from both Models D4422e and TD4422e is the dominance of the systems by only one or two downwelling, even when as many as seven plate boundaries are present. Indeed, the temperature field snapshots in Figs 10 and 13 do not indicate a clear correlation between the scale of the overlying plates and underlying convection cell wavelength. This result is in contrast to the findings of previous studies that have featured either fixed plate boundaries or a single plate (e.g. Gurnis 1988; Gurnis & Zhong 1991; Zhong & Gurnis 1993). The stratified viscosity in the systems presented here likely plays a role in maintaining the long wavelength signature of the convection (e.g. Bunge & Richards 1996; Bunge et al. 1997). When plate boundaries are not allowed to migrate, the length-scale of the plate becomes instilled in the mantle. Our models with mobile plate boundaries indicate that because the mantle cannot organize itself during the timescales associated with plate boundary reorganization this finding may not be applicable to the Earth. In contrast to the apparent absence of a correlation between plate scale and convective wavelength there does appear to be a relation between plate number and surface heat flux (Grigné et al. 2005), particularly in Model D4422e. This correlation is explained by the fact that there are more divergent plate boundaries in the system at times when there are more plates and the presence of divergent plate boundaries increases surface heat flow.

Without the availability of a mechanism to break plates, a system featuring plate evolution will eventually converge on a two plate paradigm. The Earth has not converged to such a state due to the occurrence of plate rifting as part of its natural evolutionary cycle. We have simulated plate rifting by specifying a plate yield stress and implementing the breakup of a plate if this yield stress is exceeded. Our choice of an 80 MPa yield stress is guided by laboratory experiments (e.g. Kohlstedt et al. 1995). In fact, these experiments suggest that a more appropriate yield stress may have a higher value (e.g. Tackley 2000), particularly for oceanic lithosphere. However, the fact that our calculations feature between 2 and 7 plate boundaries within a region having a lateral extent nearly equal to the circumference of the Earth appears to support 80 MPa as a reasonable choice for the model yield stress. Previous studies (e.g. Puster et al. 1995) modelled evolving plate geometries in a dynamically self-consistent mantle convection model but constrained their calculations to feature a specified number of plates at all times. A system subjected to such a constraint might be expected to evolve to a state that is dependent on the fixed plate number. By specifying a yield stress value criterion in a range that results in a time-dependent number of plates, we claim that we have added a dimension of realism to our models which may be important for determining variations in time-dependent quantities.

In general, we find that the plate yield stress and plate boundary velocity criteria that we specify result in systems in which large plates are susceptible to rifting due to the effect of integrating basal shear tractions over large areas. The pull of sinking material at convergent boundaries contributes to deviatoric tensile stresses in plate interiors and contributes to plate rifting. We find that the presence of active upwellings are not a prerequisite for plate breakup, however, in cases where they are present plate rifting tends to occur preferentially over sites associated with the most vigorous upwellings. Following plate rifting, the motion of upwellings originating deep
in the mantle does not subsequently follow the divergent boundary motion. In contrast to convergent boundaries, divergent boundaries are passive features that appear to have little effect on the evolving flow. We note that our models do not include distinctly continental lithosphere and the plate rifting events featured in the calculations are not meant to simulate continental breakup.

Newly introduced plate boundaries must initially be divergent in our models. One way in which to convert divergent plate boundaries to convergent plate boundaries is a reversal in the motion of a plate. Without flow reversals it appears that these systems are destined to feature many divergent plate boundaries and typically one dominant convergent plate boundary. The limitation of the models to two dimensions may play a role in inhibiting the conversion of plate boundaries. Non-convergent plate boundaries may be converted to convergent boundaries more easily if less dramatic events than 180° flow reversals are required for the transformation. Future 3-D studies, in either a Cartesian or spherical geometry, may exhibit different behaviour and global characteristics to these 2-D models if they allow new convergent boundaries to form more readily.

The limitation of our study to 2-D models must also have an effect on the time-dependence exhibited in the experiments. Although we have established that the combination of internally heated convection and thick plates results in highly time-dependent flows, by requiring that flow reorganization events must necessarily involve flow reversals, we have likely retarded the frequency at which these events may appear. Indeed, flow reorganization events that occur in 3-D systems, like the Earth, do not require changes in plate motion of 180°. King et al. (2002) observed flow reorganization events in a 3-D convection model with static plate geometry approximately every 200 Myr. In contrast, we observe flow reversals in Model D4.52.51.53.5 at intervals of approximately 900 Myr. Consequently, we note that the 2-D results presented here probably represent a lower bound on the degree of time-dependence expected in a 3-D system. The implementation of migrating plate boundaries in 3-D mantle convection models could result in even more time-dependent surface velocities and rapidly varying surface and possibly core heat fluxes than is indicated by these 2-D calculations. This work has shown that the time-dependence of a system like the Earth’s mantle is profoundly influenced by the motion of its surface. Our studies featuring static plate boundary models demonstrate peculiarities in the behaviour of fixed surface geometry systems that are not likely relevant to the Earth. For example, the evolution of steady flows, flow reversals and general quasi-periodic behaviour all require static plate boundaries. In contrast, models featuring evolving boundaries do not exhibit periodicity and are far from steady.

Due to the absence of flow reversals in calculations featuring evolving plate boundaries, the results presented here cast doubt on whether flow reversal behaviour has been manifested in the Earth’s history (for example, as the mechanism driving the periodic assembly and dispersal of continents). However, our findings do not rule out the possibility that mantle driven periodic flow reorganization events (King et al. 2002) may occur in a 3-D system with evolving plate boundaries. Whether plate reorganization events result from deeply placed driving forces or tectonic events (e.g. continental collisions) remains unclear (Richards & Lithgow-Bertelloni 1996). Studies incorporating fully evolving dynamic plates in 3-D convection models will be required in order to address this issue.

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Time-dependent mantle convection featuring evolving plates


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