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Influence of supercontinents on deep mantle flow

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ABSTRACT

The assembly of supercontinents should impact mantle flow fields significantly, affecting the distribution of subduction, upwelling plumes, lower mantle chemical heterogeneities, and thus plausibly contributing to voluminous volcanism that is often associated with their breakup. Alternative explanations for this volcanism include insulation by the continent and thus elevated subcontinental mantle temperatures. Here we model the thermal and dynamic impact of supercontinents on Earth-like mobile-lid convecting systems. We confirm that insulating supercontinents (over 3000 km extent) can impact mantle temperatures, but show the scale of temperature anomaly is significantly less for systems with strongly temperature-dependent viscosities and mobile continents. Additionally, for continents over 8000 km, mantle temperatures are modulated by the development of small-scale convecting systems under the continent, which arise due to inefficient lateral convection of heat at these scales. We demonstrate a statistically robust association between rising plumes supercontinental interiors for a variety of continental configurations, driven largely by the tendency of subducting slabs to lock onto continental margins. The distribution of slabs also affects the spatial positioning of deep mantle thermochemical anomalies, which demonstrate stable configurations in either the sub-supercontinent or intraoceanic domains. We find externally forced rifting scenarios unable to generate significant melt rates, and thus the ultimate cause of supercontinent breakup related volcanism is probably related to dynamic continental rifting in response to mantle reconfiguration events.

1. Introduction

The formation and breakup of supercontinents may impose an important recurring signal on the geological record (Condie, 2004). Their formation is associated with massive, widespread orogenesis, their final breakup accompanied by widespread volcanism (Condie, 2004; Condie et al., 2009-this issue). The voluminous volcanism during their breakup cannot be attributed to passive rifting, the volume of volcanism is often locally 3–4 times thicker than mid-ocean ridge volcanic sequences (Korenaga and Jordan, 2002). As a result, this anomalous volcanism has been ascribed to mantle plumes (Storey et al., 1995; Ernst and Buchan, 2001), or anomalously hot mantle beneath the supercontinent (Anderson, 1982; Coltice et al., 2007, 2009-this issue).

Either association suggests a peculiar coupling of supercontinent formation to mantle structure and dynamics. Supercontinents place a severe constraint on the global distribution of subduction; subduction cannot occur within the supercontinent, and will likely migrate to the margins of supercontinents. Thus the mantle beneath the centre of a supercontinent will not be cooled by subducting slabs for a period, allowing the build-up of heat beneath (Anderson, 1994; Lowman and Gable, 1999; Lowman and Jarvis, 1999). At the onset of supercontinent dispersal, these anomalously hot temperatures could contribute to the massive volcanism accompanying breakup (Coltice et al., 2007). This scenario has two implications: 1) that there is a finite length scale for continents, above which the sheer scale of the continent will insulate the mantle beneath it from subduction, i.e. a lower cut-off limit for the size of a supercontinent. This has been investigated in some detail in previous modelling (Lowman and Jarvis, 1993, 1995; Coltice et al., 2007). 2) There should be a minimum amount of time required for temperatures to rise to sufficient levels to generate the volume of volcanism observed. Neither of these implications have been systematically explored in detail for Earth-like convecting systems with mobile plates.

Continental blocks modulate mantle temperatures in two ways. First, the growth of continents themselves results in the depletion of heat producing elements, such as U, Th and K, from the mantle. Under some conditions this can be a first order effect, and tends to cool the mantle, having been previously explored in detail in O'Neill et al. (2005). However, continents also act as thermal insulators by preventing efficient convective heat loss to the surface. This effect has been the subject of a number of previous works (Lenardic, 1997, 1998; Lenardic and Moresi 1999, Jellinek and Lenardic, submitted).

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One important result from these studies is that increasing continental extent effectively insulates the mantle, resulting in increased mantle temperatures—not just beneath the continent, but globally.

Larger continent extents also affect mantle flow patterns, and their potential influence is illustrated in Fig. 1, which shows the change in flow patterns for laboratory convection models for increasing continental (i.e. thermal insulator) extents. Total continent extent for the Earth is ~37% (Lenardic and Moresi, 1999). At low continental extents ($L<40\%$), the flow geometry essentially scales with the size of the continent. At large extents ($L>60\%$), the flow scales with the scale of the oceanic region. At $L\sim40\%$, there was no preferred wavelength of flow. However, at $L=60\%$, the flow was tank filling—i.e. the effect of the continent on flow extends the total width of the modelling domain. At this point, the effectiveness of free-lid convection in the non-continental area in removing the system's heat is undermined by the shear scale of the continent, which influences surface convective patterns to the edge of the convection domain, and thus reducing the efficiency of lid recycling and resulting in higher mantle temperatures.

Much previous supercontinent modelling (Gurnis, 1988; Zhong and Gurnis, 1993; Lowman and Jarvis, 1999; Lenardic and Kaula, 1995, 1996; Yoshida et al., 1999; Honda et al., 2000; Coltice et al., 2007) has focussed on simple isoviscous or weakly temperature dependent systems. Such models have demonstrated temperature increases beneath supercontinents for plausible timescales, for a variety of basal to internal heating ratios and depth-dependent mantle viscosity structures.

Isoviscous models do not, however, encapsulate the temperature range or dynamics of mobile-lid systems—of which plate tectonics is an example—with strongly temperature dependent viscosities, which play an important regulatory role for internal temperatures. For example, in a such a system, increases in mantle temperature will result in lower mantle viscosities, which will convect faster, carrying away any excess thermal anomaly and heavily regulating the scale of temperature variations within the system. Furthermore, Lenardic (1998) has shown that large temperature variations beneath continents are minimized by advective mixing and lateral transport of heat to oceanic domains. They present cases showing that increasing continental extent can impact global mantle temperatures. While their models incorporate large-scale continents, these end-member cases do not preserve the continent-oceanic area ratios—shown to be important for mantle temperatures. So for the super-continent problem, it is important to study the effect of large-extent continents in systems which have a vigorously overturning oceanic lithosphere (>40% area extent, Jellinek and Lenardic, in review). Furthermore, the magnitude of mixing depends on the continent being able to drift. Gurnis (1988) showed continents tend to propagate away from upwellings and congregate on downwellings, and this mobility is an important consideration in calculations of subcontinental mantle temperature (Lenardic, 1997). A continent locked above a long-lived upwelling as a result of the models' boundary conditions is unlikely to reflect realistic mantle temperatures beneath supercontinents in Earth-like systems. This is not to say such a configuration will not occur, but it is crucial that if it does arise, it is because of the interaction of the continent with buoyancy forces in the mantle, and not the influence of boundary conditions.

On the other hand, a simple model for plume-related volcanism also suggests an underlying correlation between supercontinents and deep mantle dynamics. Why are large igneous provinces associated

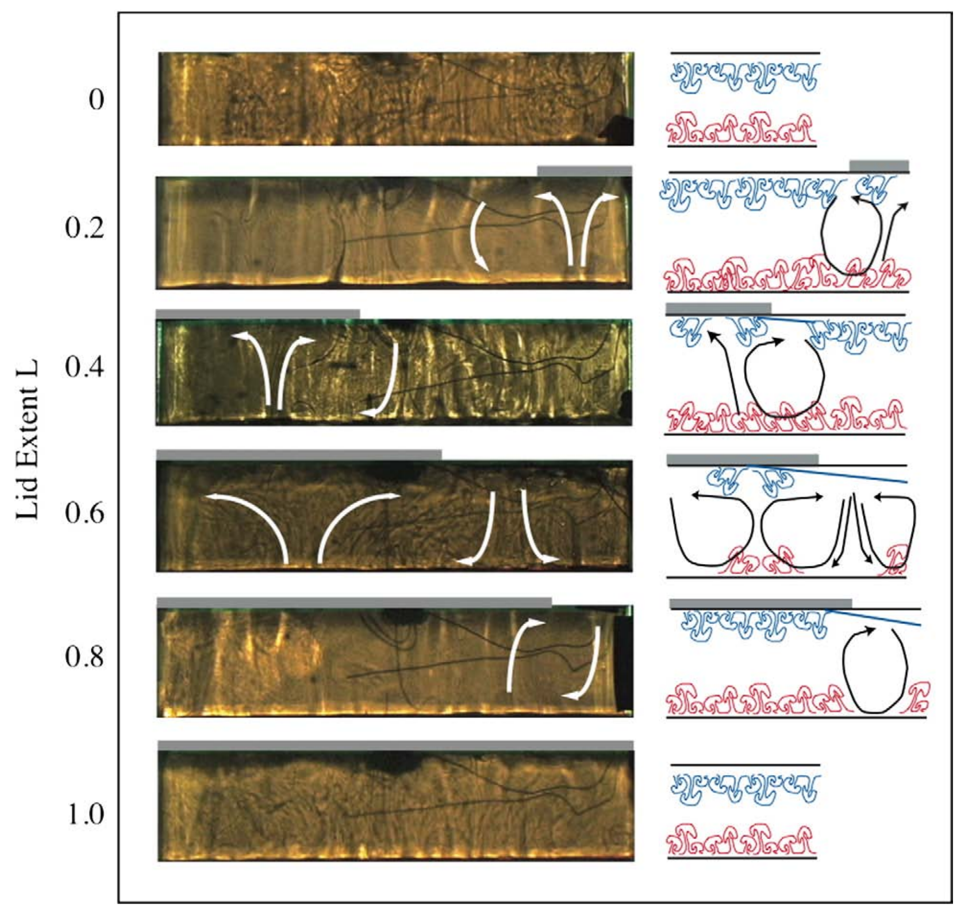

Fig. 1. Laboratory convection shadowgraphs under a lid of variable extent (L) from Jellinek and Lenardic (in review). Fluid is a Newtonian aqueous corn syrup, and the thermal boundary layers are shown diagramatically on the right. Arrows show time-averaged velocity field. For lid extents greater than 0.6, the thermal impact of the lid extends to the side boundaries of the modelling domain, affecting the heat transfer properties of the entire system (see Jellinek and Lenardic, in review for details and discussion).

Table 1

Symbol	Property	Value
Mantle		
ρ_m	Mantle density	3400 kg m^{-3}
g	Gravitation acceleration	9.81 ms^{-2}
α	Thermal expansivity	3×10^{-5} K^{-1}
T_{CMB}	Temperature at the CMB	~4000 +/− 600 K
ΔT	Non-adiabatic T contrast	2555 K
d_m	Depth of convecting mantle	2895 km
κ	Thermal diffusivity	10^{-6} m^2 s^{-1}
k	Thermal conductivity	3.5 W m^{-1} K^{-1}
C_p	Specific heat	1200 J K^{-1} kg^{-1}
H	Mantle heat production	5×10^{-12} W kg^{-1}
$\eta(T)_m$	T-dependent viscosity	10^{19}– 10^{24} 10^{24} Pa s^{-1} (FK approx[a])
–	Upper mantle visc contrast	$0.5*\eta(T)$
–	Lower mantle visc contrast	$5*\eta(T)$
dP/dT	Clapeyron slope at 670 km	−2.5 MPa/K
$\Delta\rho/\rho$	Density change across transition	10%
μ	Friction coefficient (oceanic)	0.15
$T_{sol}(0)$	Solidus at 0 GPa	1150 °C
M_{sol}	Linear solidus slope	0.9 °C/MPa
$\Delta T_{liq\text{-}sol}$	Solidus/liquidus ΔT	500 °C
L	Latent heat of melting	200 kJ/kg
Continent (as per mantle unless otherwise stated)		
ρ_c	Continent density	3000 kg m^{-3}
H_c	Continental heat production	10×10^{-12} W kg^{-1}
$\eta(T)_c$	T-dependent continent viscosity	$10^3\times\eta(T)_m$
μ_c	Friction coefficient (continental)	0.65
d_c	Continental thickness	200 km
Lower mantle thermochemical piles (as per mantle unless otherwise stated)		
$\Delta\rho_{lmtc\text{-}1}$	Pile density difference	$1.25\rho_m$
$\eta(T)_{lmtc\text{-}1}$	Pile viscosity—model 1	5×10^{21} Pa s^{-1}
$\eta(T)_{lmtc\text{-}2}$	Pile viscosity—model 2	8×10^{23} Pa s^{-1}

For references see Schubert et al. (2001).
[a] Frank-Kamenetski approximation viscosity profile (Lenardic et al., 2003).

(spatially and in time) with the onset of supercontinent rifting? Either plumes are determining the loci of extension; i.e. a continent under tensile stress will rift if an upwelling plume arrives beneath it, or plumes are in some way localized by the same processes that ultimately result in supercontinent breakup.

The arrival of a plume head would result in significant buoyancy induced stresses within the lithosphere, thermally weakening the lithosphere, and thus rifting would be associated spatially and temporally with the arrival of a plume (Courtillot et al., 1999; Hill, 1991; Richards et al., 1989). This is not necessarily an unusual postulate: ridge jumps and microcontinent formation have been observed in relation to plume dynamics (Müller et al., 2001). It does, however, suggest a more active role for mantle plumes in continental dynamics than has been suggested (Anderson, 1982). On the other hand, if upwelling plumes are passively advected into plate spreading centres associated with continental breakup, then the coincident timing of plume arrival and rifting is extremely fortuitous.

In either case, the reason for the association of upwelling plumes and continental interiors is not clear. Qualitatively, it has been argued that the predilection of subduction to occur on continental margins (Gurnis, 1988; Lowman and Jarvis, 1996) instigates a large-scale return flow away from the margins—in continental interiors or in intraoceanic basins. However, large Rayleigh number (Ra) convection appropriate to the Earth is intrinsically unsteady, and it has not been demonstrated that this relationship holds quantitatively for such systems incorporating a plate-like mode of convection and an Earth-like viscosity structure.

Recently, it has been postulated that the surface distribution of plume initiation and their subsequent distribution is strongly correlated with two geoid highs in the Pacific and under Africa (Burke and Torsvik, 2004; Burke et al., 2008). These geoid anomalies are associated with extremely slow S-wave velocities, and the sharp structure and lack of P–S velocity correlation suggests a compositional, rather than purely thermal, origin (Ni et al., 2002). The shape and distribution of these thermochemical anomalies can be explained by the history of subducted slabs over the last 200 Ma (McNamara and Zhong, 2005; see also Zhang et al., 2009-this issue); subducting slabs seem to have "bulldozed" the thermochemical material into piles, which constrains their physical properties to being dense viscous blobs (McNamara and Zhong, 2005). The existence of such lower mantle structures can provide for the long-lived stability of plume conduits in the lower mantle (Jellinek and Manga, 2004), as well as providing the loci for their formation and ascent (Burke et al., 2008; Zhao, 2009-this issue). Furthermore, these lower mantle anomalies have been suggested to reflect the past distribution of supercontinents, due to association of supercontinents, and lower mantle blobs, with subduction distribution (Maruyama et al., 2007; Rino et al., 2008). This positional correlation between the sub-African anomaly and Pangaea certainly seems reasonable (Burke et al., 2008), however, the association between the Pacific anomaly and Rodinia is more speculative. Zhong et al. (2007) present models illustrating the reconfiguration of mantle convection in response to supercontinent formation, with showing the development of an active upwelling and degree-2 convection underneath a supercontinent, and demonstrated the spatial correlation between deep geoid anomalies, the onset of degree-2 convection, and supercontinent position.

In this paper we investigate the impact of supercontinents in mobile-lid convecting systems, with Earth-like viscosity structures, efficiently overturning oceanic domains, and 'mobile' continents. We address three important outstanding questions regarding the effect of supercontinents on deep mantle dynamics. 1/ Is there a finite length scale above which subduction cannot efficiently cool the subcontinental mantle, and if so, what are the characteristic mantle flow patterns associated with this regime? 2/ What is the spatial relationship between supercontinent position and the deep loci, and eruption localities, of mantle plumes? 3/ What is the effect of supercontinent position on deep mantle heterogeneities, and how does this affect the local distribution of mantle plumes? These questions have not been addressed from the context of Earth-like convecting systems incorporating supercontinents and plate behaviour, and the purpose of this paper is to explore the interaction of the components of such systems in models incorporating the necessary physics to reproduce this mode of convection.

2. Method

We will address these questions using a 2D Cartesian finite element mantle convection code (Ellipsis). The code is capable of simulating compositionally distinct supercontinents and mantle melting. While the two-dimensional restriction is severe, as it doesn't incorporate important details of 3D mantle flow, it is important to understand the interaction between competing factors in this complex system first. Supercontinent position, plume localisation, evolution of mantle heterogeneities, and magnitude of mantle melting can all be addressed in 2D models, and the goal here is to obtain a quantitative understanding of the importance of each of these factors.

The code itself solves the standard convective equations for energy and Stokes flow subject to an incompressibility constraint, and is more thoroughly outlined in Moresi et al. (2003) and O'Neill et al. (2006). It employs Langrangian integration points on a fixed mesh, the points are equivalent to material points and carry information on composition, melt depletion history, strain history etc.

We use the Frank-Kamenetskii (1969) approximation for mantle viscosity, which varies over five orders of magnitude. The rigid plates deform by plastic failure once a critical yield stress is reached (which is defined using the Byerlee criterion). Deformation of failed material is achieved by employing a modified viscosity (see Moresi and Solomatov, 1998, for details). The material parameters in this study are outlined in Table 1. For simulations with deforming continents, the relevant continental rheology is outlined in Table 1 and in Fig. 2. For the simpler models, the continents are held fixed and act as rigid blocks. This affect is achieved by using high viscosity and yield stress values. Though held in-situ, the free-slip top and bottom boundary conditions (BCs), and the periodic side BCs for the mantle, mean the model is essentially Lagrangian with respect to the continent's reference frame, and dynamically equivalent to a moving continent in a mantle reference frame (Gurnis, 1988; Gurnis and Zhong, 1991).

To simulate mantle melting, we use the parameterization of McKenzie and Bickle (1988), which calculates the percentage melt produced based on the supersolidus temperature (see Fig. 2). We use the solidus of Takahashi et al. (1993) and assume a 500 °C temperature difference between the liquidus and solidus. We assume that once melt is produced in the mantle, it migrates from the mantle system, leaving behind a residuum that is melt depleted, quantitatively equivalent to the volumetric fraction of melt produced. The particle will no longer melt unless the highest melt fraction it has previously reached is exceeded, then another fraction of melt is produced equal to the difference, and the depletion value increases to this new highest melt fraction. Full details of this implementation are given in O'Neill et al. (2005), and it is shown schematically in Fig. 2.

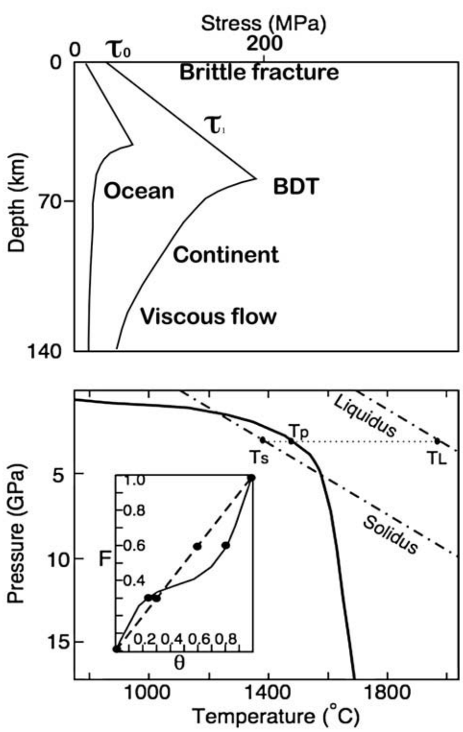

Fig. 2. a) Representation of the strength of the oceanic and continental lithosphere in the simulations presented. The materials follow a Frank-Kamenetski temperature dependent viscosity flow law (Moresi and Solomatov, 1998). In the shallower regions, the stresses in the cold highly viscous lithosphere exceed the yield stress τ_y ($\tau_y = \tau_0 + \tau_1 P$), and the lithosphere deforms in a plastic manner (see Moresi and Solomatov, 1998 for details). The primary difference between the oceans and the continents in these models is the material parameters τ_0 and τ_1, which are high for the continents to hold them rigid, and lower for the oceanic lithosphere to permit mobile-lid convection. b) Parameterization of melting used in the simulations. A typical geotherm is shown as the dark solid line. The solidus and liquidus (from Takahashi et al., 1993) are marked. The supersolidus temperature θ is defined as the relative degree the geotherm is between the solidus and liquidus ($(T_p - T_s)/(T_L - T_s)$). c) (inset) the melt fraction F can be determined from the supersolidus temperature using a parameterization of melting, such as that of McKenzie and Bickle (1988), shown as a dark line the linear relationship from Jaques and Green (1980) is shown as dashed.

3. Results

3.1. Insulation, plumes, and mantle temperatures

The simulations presented here have large aspect ratios, though not large enough to preserve this ratio of continental to global surface area, which would be ideal, but require 11 × 1 aspect ratio models at minimum. This has implications for modelling supercontinents in that while the scale of supercontinents is large (>10,000–12,000 km in lateral extent), the total continental volume of the Earth is unchanged. However, at these aspect ratios the models are very computationally intensive. Instead, as we have argued, the thermal and geometrical characteristics of the flow should be adequately resolved, for representative supercontinent extent/mantle depth ratios, provided the continental extent does not exceed 60%. We present models at varying aspect ratios that preserve this system constraint. The initial conditions for these models are imported from previous low-resolution runs to steady state with a continent imposed.

An example of mobile lid convection with a supercontinent is shown in Fig. 3. This is an example of a long-timescale simulation that ran for >3 Gyr. We have adopted a Langrangian reference frame with respect to the continent; i.e. the continent is held fixed and mantle is free to move around it. The continental extent is 50% (equivalent to an extent of 5780 km), the side boundary conditions are periodic, and all the material parameters are summarized in Table 1. The time series shown are for the temperature field, and melt depletion field, calculated as described in the methodology. The depletion is a history dependant parameter, and gradually increases over the lifetime of a simulation as more material passes through the melting zone.

As in previous models incorporating an endothermic transition at 670 km with plausible Clapeyron slopes, and highly viscous subducting slabs, slabs penetrate into the lower mantle with great frequency, though they do interact with, and are occasionally hindered by, the 670 km phase transition. Notably, though, the distribution of descending slabs is strongly tied to the continental margins, with subduction zones migrating to these zones where they either affix to the margins, or become extinct (particularly when a rolling back subduction zone encounters a continent). The net result is the distribution of subducting slabs in the mantle is very strongly correlated with the margins of continents.

This also has an effect on the distribution of upwelling plumes. Plumes originate from over-thickened hot thermal boundary layer material at the lower boundary. Subducting slabs induce a flow field around themselves, which tends to herd active upwelling features away from the zones of downwelling. Slabs encountering the core–mantle boundary (CMB) will propagate laterally due to the background flow field, and the gravitational instability of descending material above. As slabs at the CMB plough sideways, they bulldoze any lower viscosity plumes they encounter away from the zone of downwelling. Furthermore, cold slabs at the CMB hinder the generation of a hot, thick thermal boundary layer required for plume initiation, and thus suppress plume formation. The net result of these effects is, over long time periods, an anti-correlation between the average position of subducting slabs (continental margins) and the more likely position of plumes (continental or oceanic interiors). Over the course of several mantle overturns, initiating plumes will entrain significant amounts of older depleted slab material, depending on plume position and depletion of the source, as shown in the evolution of melt depletion diagrams.

Fig. 4 shows the evolution in temperature, and lower mantle velocities, for the model shown in Fig. 3. We should stress that the model was run at constant heat production and basal temperatures, and so represents a steady-state equilibrium convection calculation, not an evolutionary model. This is appropriate as here we are examining the dynamics of these systems under a regime similar to Earth today, not examining the history-dependent dynamics of

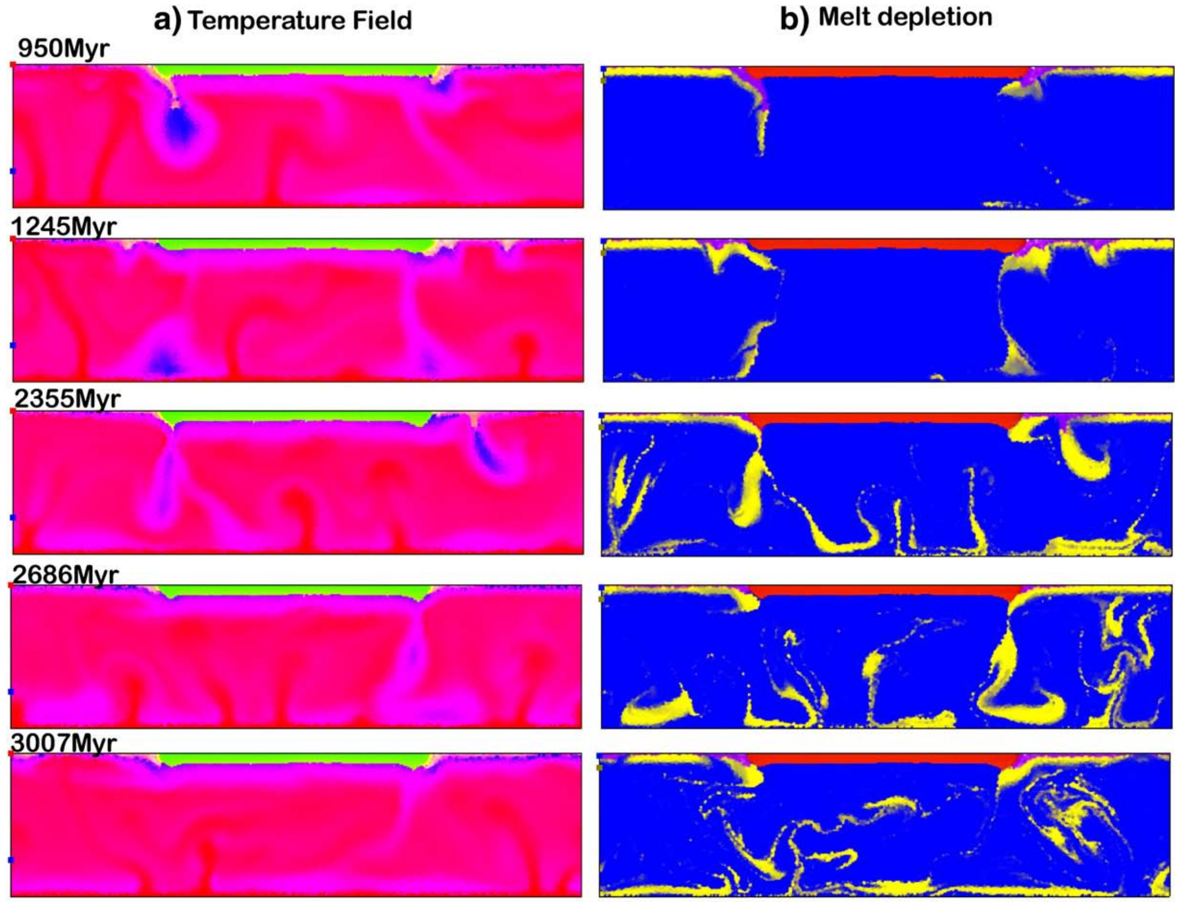

Fig. 3. Time series of a convecting system with parameters outlined in Table 1, showing the evolution of the temperature field (left) and melt depletion field (right). A continent of fixed extent, occupying 50% of the modelling domain (i.e. 5780 km) is fixed in place by a surface velocity condition, and by virtue of its large viscosity and yield strength. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

thermal evolution simulation, which is beyond the scope of this paper. Here we define the "continental interior" as the region of the continent more than 500 km from a margin, and thus are not overly affected by subduction zone dynamics. Fig. 4a shows the evolution of sublithospheric potential upper mantle temperatures within the continental interior, compared to the global upper mantle average. The upper mantle average is strongly affect by subducting slabs, which contribute to both its long-term average of 1350–1400 °C potential temperatures, but also the periodic low-temperature spikes, which reflect periods when cold subducting slabs dominate the upper mantle temperatures. In comparison, the continental interior experiences on average somewhat higher temperatures (1425–1450 °C) than the upper mantle average, which reflect its isolation from subduction, and also the proclivity of plumes to upwell in this zone and elevate the regional asthenospheric temperatures. To put this in perspective, local variations in temperature swamp this difference throughout the course of the simulation. Shown in Fig. 4b is the vertical velocity of the mantle at a depth of 2167.5 km (shown as a blue square on the left hand side of the temperature cross sections of Fig. 3). This depth is deep enough to uniquely diagnose lower mantle dynamics, but elevated enough to avoid the large lateral motions associated with the lower boundary layer. The vertical velocity is positive for upwellings, negative for downwellings. The global average must be zero—for an incompressible medium, the amount of material rising through this depth must equal the amount of material descending through it. However, in regions of plume concentration, the average velocity will be positive, for regions of active subduction and downgoing slabs, the average will be negative. The average for the whole continent is close to zero—this reflects the combined effects of subducting slabs at the continental margins, and upwelling plumes in the interior. However, the average vertical velocity for the continental interior is significantly positive (~0.1 cm/yr). Note this is not the upwelling velocity of individual plumes, which may rise much faster, but the combined average for the entire interior. This elevated average reflects the propensity of mantle plumes to rise within the continental interior. Thus in this one simple model we have seen 1/ how continents impose dynamical constraints on the mantle, encouraging upwelling plumes beneath them by virtue of their influence on subduction zone dynamics; And 2/ enable hotter than average temperatures in the mantle beneath the continental interior, as a result their isolation from subduction, and the continual addition of hot plume material to the asthenosphere beneath the continent.

How do these effects vary for increasing continental extent? Fig. 5 illustrates the change in flow patterns for increasing continental extent. For small continents (2890 km), no portion of the subcontinental mantle is ever really free of the influence of subduction. Downgoing subducting slabs are present beneath large portions of the continent for much of its history, and little opportunity exists for development of stable plumes beneath the interior for this configuration (Lowman and Jarvis, 1995, 1999). At the other extreme, for continents larger than ~8000 km, small-scale convection becomes an important mechanism for upper mantle heat transport. The inability of hot material beneath such large continents to be advected into oceanic regions, and minimal subduction-driven lateral flow, allows the thermal boundary layer beneath the continent to thicken and cool, in some cases forming drips which penetrate into the lower mantle (see Fig. 5f, centre of continent). These non-subduction density anomalies drive small-scale upper mantle flow with cell sizes ~500–

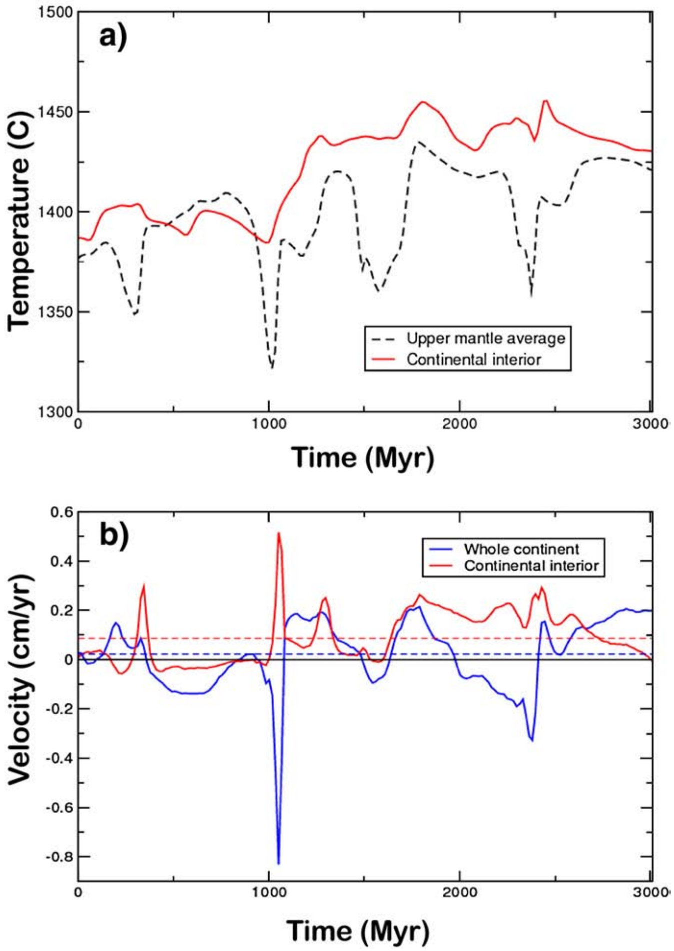

Fig. 4. a) Evolution of upper mantle average potential temperatures (dashed, defined to be within the depth range of 200–600 km) from the model in Fig. 3, and upper mantle temperatures in this same depth range beneath the continental interior (subcontinental upper mantle more than 500 km away from the continental margin). b) Evolution of the vertical velocity field at a depth of 2167.5 km (blue squares to the left of the temperature field diagrams in Fig. 3). Upwards is positive. The average for the whole mantle is 0 (mass flow up and down must be balanced). Vertical velocities beneath the whole continent are affected by subducting slabs and rising plumes. Vertical velocity beneath the continental interior is effected primarily by the rising plumes in this example.

800 km. This is an important difference between mid-size (3000–8000 km) continents, and supercontinents (>8000 km extent). Lateral mantle flow beneath mid-size continents, driven by ultimately by subduction, allows the advection of heat and hinders the development of cold TBL anomalies, whereas the large extent of supercontinents requires the development of these anomalies to allow heat transport in these isolated regions. The snapshots in Fig. 5 are for different times in each given simulation, and thus the development of the depletion field is at different stages for each. Large aspect ratio simulations take significantly longer to run, and thus are only allowed to evolve for shorter timescales (~1 Gyr rather than 3 Gyr for 4×1 simulations). These models start from an initial steady state (i.e. output from a previous long-period calculation). Given that supercontinents form and breakup on the order of <500 Myr, this is more than enough time to discern the effects of scale in supercontinent formation. The amount of melting generated beneath the continent in these examples is minimal due to its thick stable architecture.

Fig. 6 illustrates the ratio of interior subcontinental mantle temperature, to the upper mantle average. Here we have allowed the simulation to evolve to a statistical steady state, then taken the time average of these temperatures. Since these simulations were performed at different continental/oceanic extents (which affects the global upper mantle temperatures; Jellinek and Lenardic, in review), we normalize the sub-continental temperatures to eliminate this effect. There is no systematic trend in sub-continental temperatures, compared to the upper mantle average, with increasing continental extent. In fact for larger continents, the temperatures are if anything diminished. This is due the development of a thick, cool boundary layer in the continental interior due to its isolated conditions, and the development of small upper-mantle scale convection cells beneath the continent, which enables cooling of the subcontinental asthenospheric mantle by mixing the TBL back in (Solomatov and Moresi, 2000). Since the oceanic upper mantle remains hot and is able to be advected all the way to the surface, this dominates the temperature ratio parameter, swamping local subduction effects and temperature perturbations beneath the continents, and so the temperature ratio is less than one for some continental extents.

The subcontinental vertical velocities at 2167.5 km depth also show little systematic trend with increasing continent size. The smallest continents are always dominated by subduction, and so the vertical velocities beneath it reflect the effect of downgoing slabs beneath it at depth. For larger continents, the interior velocities are isolated from descending slabs, and mostly reflect the predominance of active upwelling plumes beneath the continental interior, due to the dynamics previously discussed. However, this plume focussing seems to reach a peak at around 6000 km, after which the dominance of plumes in the lower mantle velocity field beneath continents is diminished. The reason for this is that at scales >6000 km, active downwellings form in the continental interior, and these counteract the effect of upwelling plumes in the "average" vertical velocity. To summarize this, plumes are focussed beneath the interior of sizable (>3000 km) continents by the dynamics of continental-margin subduction zones, but above the 3000 km threshold, there is no systematic increase in plume activity with increasing continental size. Beyond about ~6000 km in extent, small-scale downwellings develop in the sub-continental thermal boundary layer, enabling the establishment of non-subduction related convection systems.

The average deep mantle (2167.5 km) vertical velocity for a simulation involving a 5780 km wide continent is shown in Fig. 7. Two snapshots of the temperature field, from times ~1 Gyr apart (Fig. 7a and b), illustrate the correlation of subducting slabs with the continental margins, and also the tendency of plume ascent to occur either in the continental interior or mid-oceanic regions—away from active subduction. This is shown somewhat more robustly in Fig. 7c, which plots the time-averaged vertical velocity at 2167.5 km depth for the entire simulation, against distance across the modelling domain. The position of the continent is the same as in the snapshots in Fig. 7a and b. The negative long-term average vertical velocity near the continental margins is again due to the propensity of subduction to occur there. The positive average vertical velocity in the continental interior, and in the intra-oceanic regions, reflect the dominance of plume ascent locally in these regions. In this example, the continent is too small for intra-continental small-scale convection to play a large role.

3.2. Deep mantle heterogeneities

The previous section's results shed light on the relationship between deep mantle dynamics and continental position and size. The third question posed in this paper, is how supercontinents may be related to the distribution deep mantle chemical heterogeneities, and how this might affect plume positions. Fig. 8 illustrates a similar model setup to previous models, with a 4×1 aspect ratio and a 5780 km wide continent. This time, though, a low viscosity, dense, deep mantle heterogeneity is added (material properties in Table 1). The deep chemical layer is shown as purple in the temperature plots, or green in the depletion plots. Due to its low viscosity in this example, the pile is initially split by dense, viscous subducting slabs, and then

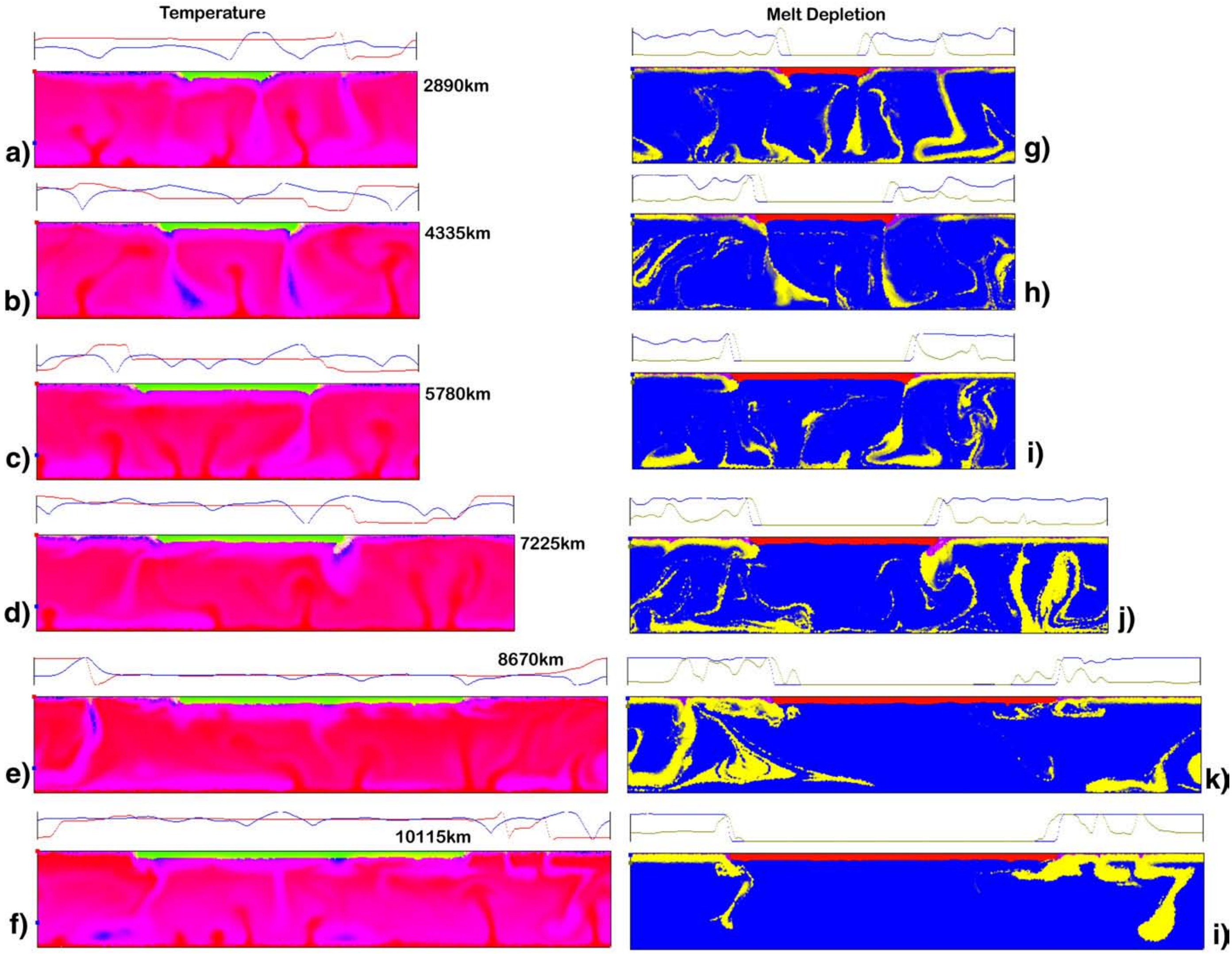

Fig. 5. Snapshots from convecting systems with variable continental extents. Larger aspect ratios are needed for larger continental extents (listed) to avoid edge effects. Shown are the temperature field (a–f) and the depletion field (g–l). Snapshots are different times in the evolution of these models, large aspect ratio models did not run as long as smaller models. Parameters in Table 1. Curves above temperature plots represent normalized horizontal surface velocity (red) and deep mantle vertical velocity (blue, positive down). The curves above the depletion plot represent the value of depletion at depth indicated by small boxes on the left of the depletion field plots. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

herded by laterally propagating slabs at the CMB into two piles—one beneath the continent, and the other directly beneath the centre of the oceanic region. These positions are, however, quite time-variable, and the blobs will move considerably over the course of a simulation due to the vagaries of subduction. Interestingly, the mechanisms by which the chemical layer is thickened also tend to modify the thermal boundary layer, which becomes thickened over the piles themselves. Thus there is a tendency for plume initiation and ascent to be strongly correlated the positions of these piles. More specifically, the mechanical boundary posed by the piles to thermal boundary layer thickening means that the TBL becomes overthickened right at the edge of the piles, and plume initiation is strongly correlated with these edges. This is exactly what has been suggested by Burke et al. (2008), who identified a correlation between reconstructed plume-head related volcanics (LIPS), and the edges of the thermochemical piles beneath the Pacific and Africa. Furthermore, the position of upwelling plumes becomes strongly tied to these piles, and so is quite stable with respect to the position of the piles. This is in-line with previous work by Jellinek and Manga (2004) and McNamara and Zhong (2004), who suggested from laboratory and numerical experiments that plume stability, and thus the hotspot reference frame, might be intimately tied with deep mantle chemical structures.

Previous work by NcNamara and Zhong (2004, 2005) argued that while the African pile, which has a ridge-like, sharp-edged morphology (Ni et al., 2002), can be explained by the interaction of subducting slabs with the a low viscosity material, the Pacific pile, which is more rounded in shape, is more consistent with a higher viscosity material. In this vein we have also simulated the evolution of thermochemical piles with a viscosity 80% of a slab at the surface at 0 °C (i.e. 80% of the largest viscosity in the system), shown in Fig. 9. Despite the strength of these blobs, they are still shunted laterally by descending slabs, and more importantly are dissected and split by slabs arriving at the CMB. The main difference between this and the previous simulation is the thermochemical piles retain their form, including sharp edges, more effectively than weaker, less viscous piles. The correlation between initiating mantle plumes and the edges of piles remains, as does the stability of plumes affixed to these piles. In this example, the piles actually migrate from the continental regions into the intra-oceanic domain, due to the effect of laterally propagating slabs at the CMB. This illustrates that the position of deep mantle heterogeneities is

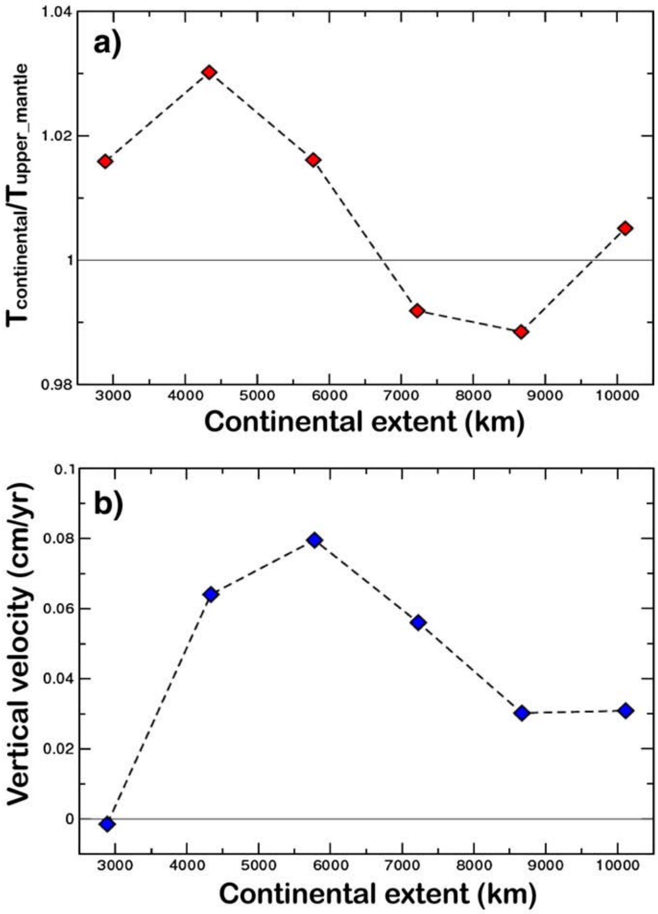

Fig. 6. a) Variation in subcontinental upper mantle potential temperatures (normalised to the upper mantle average for each simulation) against continental extent. Each simulation was run to statistical steady state, and averages of temperature field taken. Temperatures are normalized to avoid systematic variations in upper mantle temperatures due to changes in the ratio of continental extent to modelling domain lateral extent (see discussion in text). b) Variation in subcontinental interior vertical velocity at 2167.5 km depth, vs continental extent. Positive is upwards. The vertical velocity field beneath small continents is dominated by subduction, in mid-range continents by rising plumes, and in the largest continents by plumes and intracontinental small-scale convection.

quite fluid, with stable configurations possible both in the intro-oceanic areas, or beneath stable continents, as both regions are away from active subduction. The system may change which configuration its adopts many times over course of a simulation. The net effect on the average position of plumes, calculated from the average vertical velocity of the simulation, is shown in Fig. 10. Here we have plotted the time-averaged vertical velocity at 2167.5 km, for a model with no thermochemical piles (as in Fig. 7), and for the two models with low and high viscosity piles shown in Figs. 8 and 9. Again, the velocity field is dominated by subduction near the edges of the continent, and upwelling plumes in the continental interior and intra-oceanic regions. The picture is not really changed significantly when thermochemical piles are added. The work of Jellinek and Manga (2004) predicts that due to the viscous coupling of plumes with thermochemical piles, plume ascent velocity should be less, in these models this is difficult to ascertain as we are looking at long-timescale average temperatures. Despite their effect on plume stability, the piles themselves are quite mobile, and thus the generation of plumes in the continental interior is quite transient, though statistically favourable on average. The basic conclusion that upwellings are concentrated into the base of the continents is not affected D" dynamics in the models presented here.

3.3. Melt production and rifting

One prominent omission from the previous results is the lack of melting beneath a thick, stable continent. Understanding this melting is critical not just from a geological perspective but also for its potential atmospheric and climatic impact (e.g. Santosh and Omori, 2008). The lack of melt beneath thick stable continents in these models is not surprising, even hot, upwelling mantle beneath a thick continental lithosphere will not generate much melt. Voluminous volcanism is only associated with supercontinent rifting, and thus is not expected to be a feature of stable supercontinent configurations. To explore the link between rifting and volcanism in these models, we have simulated the breakup of a previously stable supercontinent in Fig. 11. The starting point for each model is a supercontinent simulation that has already evolved to steady state (as in Fig. 3). Thus the subcontinental mantle is already elevated in temperature, and statistically plumes are stable under continent at the beginning of rifting. The spreading in the cases presented is controlled by surface velocity conditions, with a very slow initiating rift (relative velocity between the blocks of 0.25 cm/yr, Fig. 11a,b) contrasted against a fast rift scenario (Fig. 11c and d). The degree of melting for these examples is shown in Fig. 11e. Despite similar initial conditions, the generation of melt in these two scenarios is primarily governed by spreading rate, a result which is not entirely surprising (e.g. McKenzie and Bickle, 1988). The melt generation rates are under 40 m^2/yr for the slow rift, or >300 m^2/yr for the fast rift. We use the cross-sectional melt area

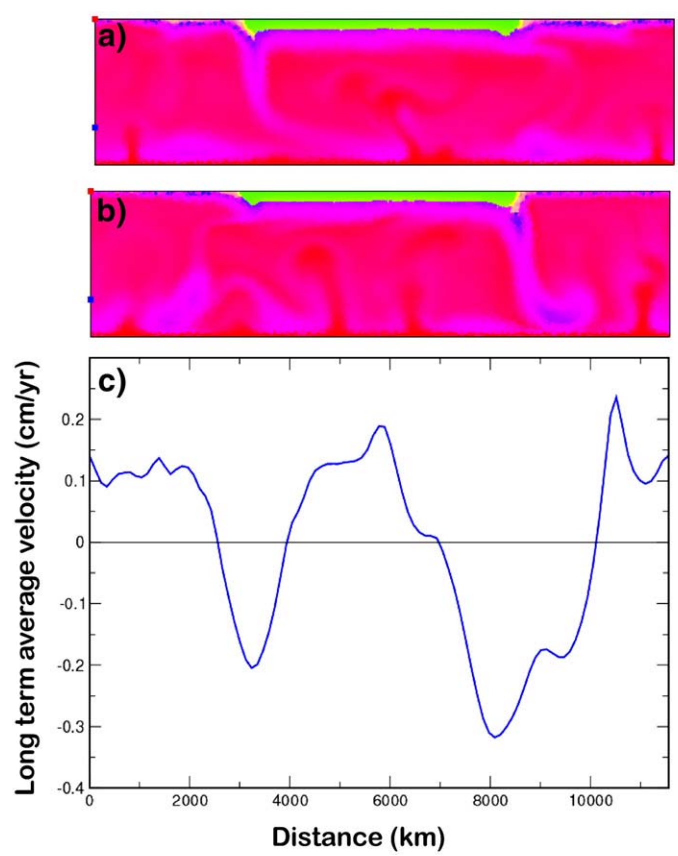

Fig. 7. a) and b) show two snapshots of a mobile lid convecting system with a fixed rigid continent of 5780 km extent. c) Long term steady state vertical velocity at 2167.5 km depth (shown by the blue box in a and b). The margins have a consistently negative vertical velocity over long time periods, consistent with downgoing slabs dominating the velocity field in these regions. Both intraoceanic in continental interior regions have consistently positive anomalies, representing the counterflow to the dowgoing slabs, and dominated by the localisation of upwelling plumes in these regions. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

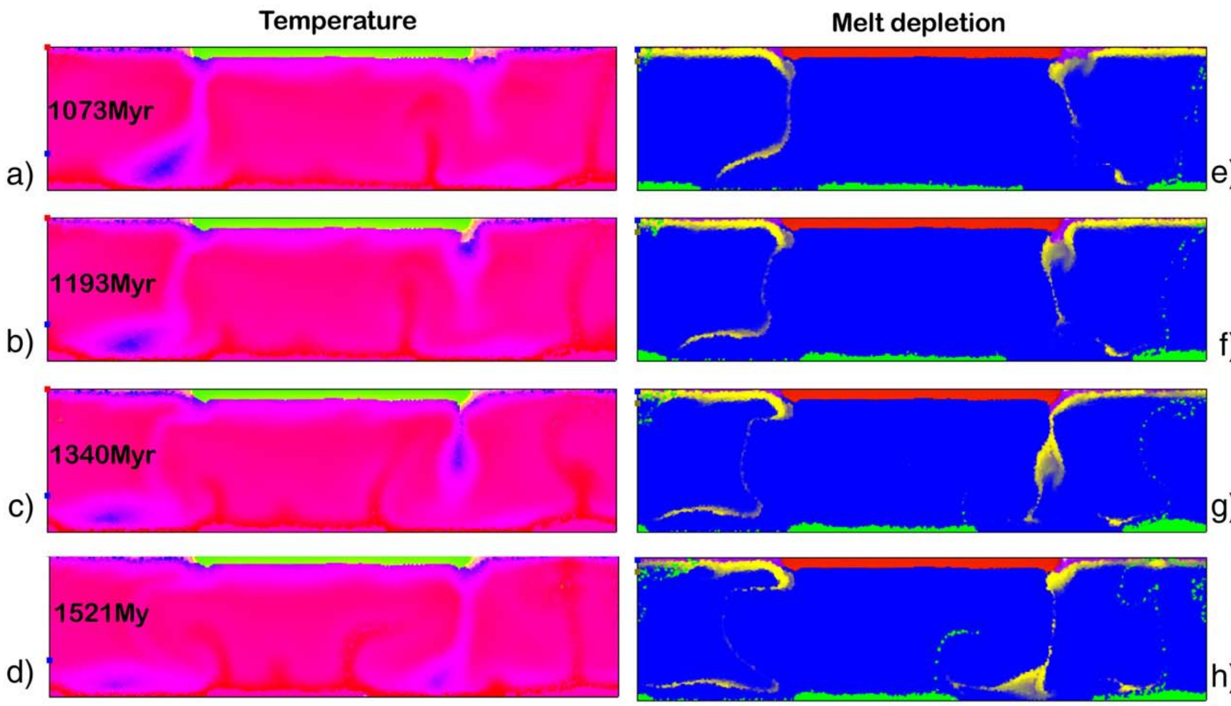

Fig. 8. Evolution of the temperature (a–d) and melt depletion (e–f) fields for a simulation with a 5780 km wide continent, and a dense, low viscosity chemical boundary at the base of the mantle (dark purple in a–d, green in e–h). Material properties in Table 1. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

rather than volume, as these are 2-dimensional models. For comparison, the melt generation rate of a mid-ocean ridge cross-section spreading at 5 cm/yr with 7 km thick crust is 700 m^2/yr. So the melt volumes are not entirely out of line with passive spreading margins, and we have not explained the origin of the voluminous volcanism during supercontinent breakup (~2000 m^2/yr). Importantly, the advection of plumes into the active rift zone is not a dominant feature of these models, and so scenarios involving the passive advection of plume conduits into an independent rift seem problematic. One of the major limitations of these models is that they do not account for dynamic rifting in response to plume upwellings, or upper mantle–lower mantle breakthrough events. Both of these scenarios link a major melt-generation event with a causative tectonic precursor to rifting, and this interaction should be explored in further work.

4. Discussion

A number of previous works have illustrated the basic mechanism supercontinent insulation: when the scale of a supercontinent is several times that of the convecting layer (mantle depth), subduction at the continental periphery cannot cool the interior and the sub-continental mantle heats up (Gurnis, 1988; Zhong and Gurnis, 1993; Lenardic and Kaula, 1995, 1996; Lowman and Jarvis, 1999; Coltice et al.,

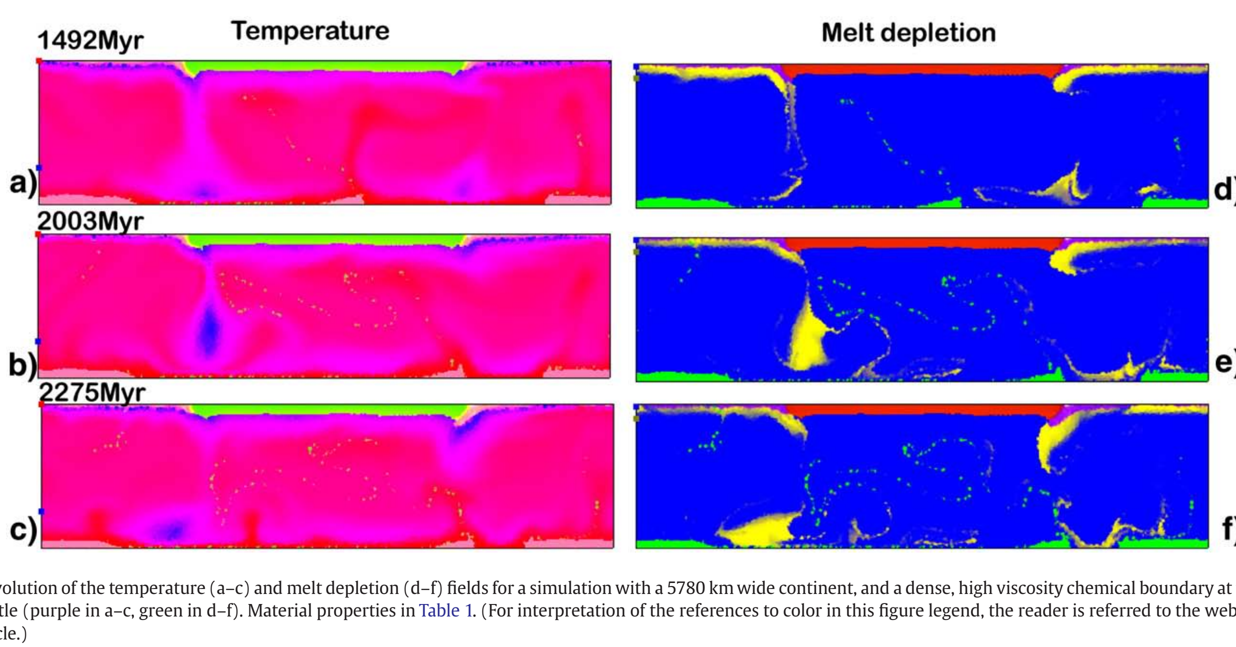

Fig. 9. Evolution of the temperature (a–c) and melt depletion (d–f) fields for a simulation with a 5780 km wide continent, and a dense, high viscosity chemical boundary at the base of the mantle (purple in a–c, green in d–f). Material properties in Table 1. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

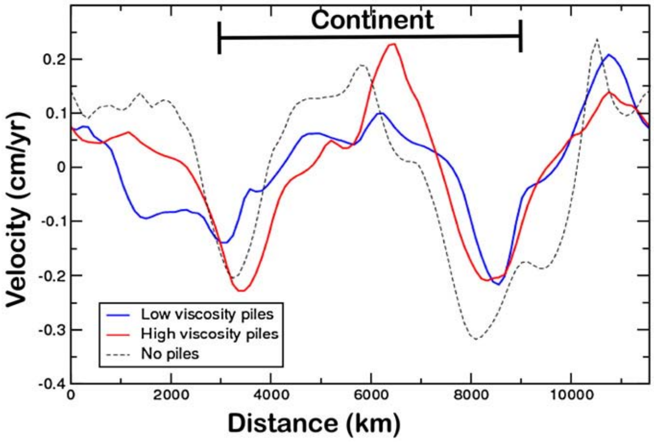

Fig. 10. Time averaged vertical velocity for the simulations shown in Figs. 3, 8 and 9, for no piles, low viscosity piles and high viscosity piles. Deep mantle upwellings are dominant in the continental interior region in all cases.

2007). Many of these early models used isoviscous, or weakly temperature dependent viscosities. Strongly temperature-dependent viscosities impose a strong regulatory constraint on mantle temperatures; hot mantle convects faster and looses heat efficiently, cool mantle is more sluggish and inefficiently looses heat, allowing time for radiogenic heating to heat the mantle up (Tozer, 1974). Thus the temperature variance in a system with strongly temperature dependent viscosities is a lot less than for equivalent isoviscous systems. This has a large impact for sub-supercontinent temperatures. While the basic mechanism—isolation from subduction—is still at work in the models presented here, the degree that sub continental temperatures are elevated is only 25–50 °C, far less than the ~200 °C observed in some isoviscous models. Also, the range in temperatures seen in any given simulation is >50 °C, and thus larger than the systematic sub continental temperature increases. These fluctuations are more due to time-dependent high-*Ra* convection rather than continental size (Phillips and Bunge, 2005, 2007). Thus, it is difficult to explain the large scale volcanism observed accompanying supercontinent breakup through this mechanism alone. Comparison with Cooper et al. (2006), who plot continental heat flow against extent for a wide range of key parameter variations, suggest this result is robust. Furthermore, there is no systematic increase in subcontinental temperatures with increasing continental extent. Jellinek and Lenardic (in review) showed that in the limit of large *Ra* that lateral temp variations from continent to ocean tend to zero, due to advective mixing. In these models, this advective mixing is not only lateral, but includes small scale convective systems beneath the continent. Once a continent exceeds ~6000 km laterally, instabilities in the cold subcontinental thermal boundary layer form, and the resulting small-scale convection systems efficiently transport heat beneath the subcontinental lithosphere, ameliorating further temperature rises. This idea of small-scale convection beneath supercontinents has parallels in Korenaga and Jordan (2002), who explored small-scale convection scenarios as an alternative to plumes in explaining subcontinental melting.

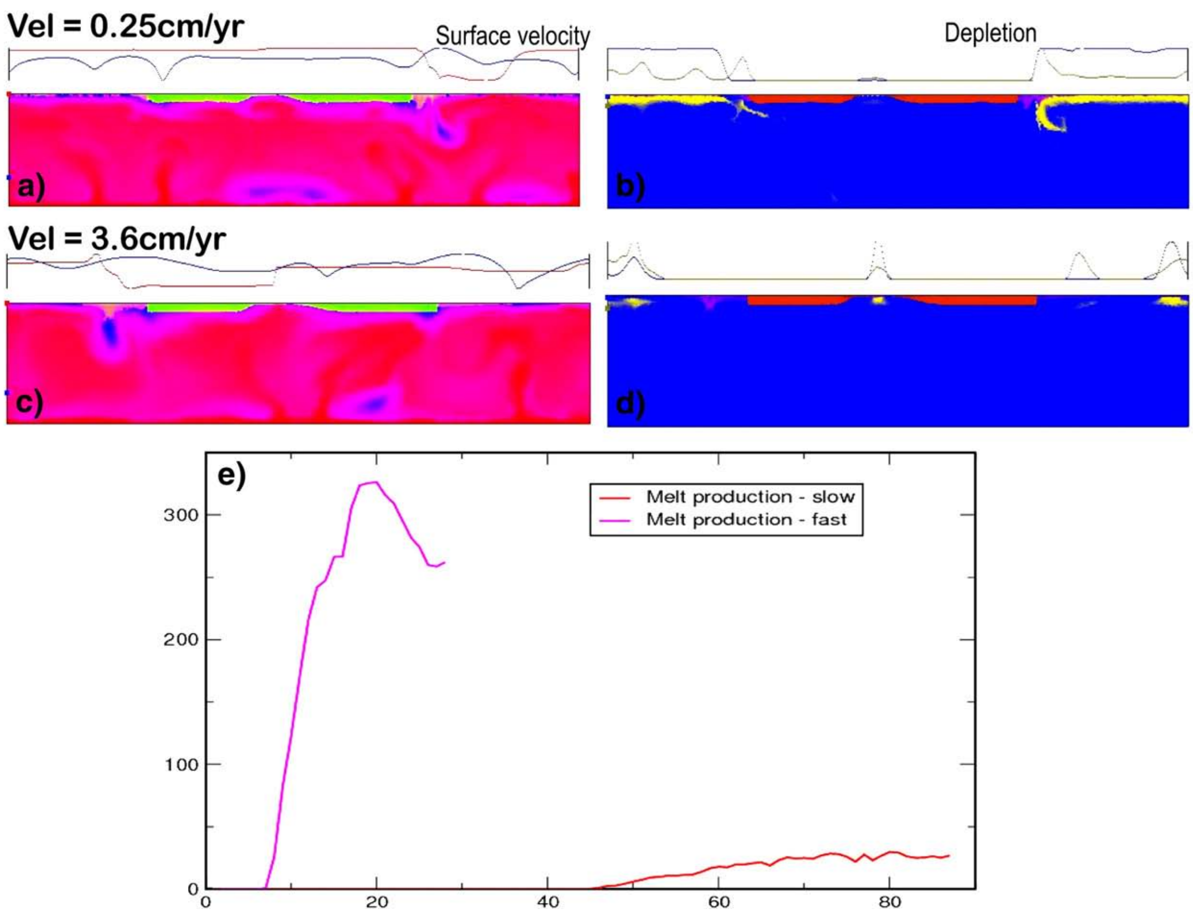

Fig. 11. Snapshots of temperature and depletion fields for early stage melting during continental rifting at two different spreading velocities, 0.25 cm/yr (a, b and 'slow' in e) and 3.6 cm/yr (c, d and 'fast' in e). The starting condition was from a previous simulation which had evolved to a hot subcontinental temperature condition (e.g. Fig. 3). e) Melt production vs time for the two simulations shown in a and c. The maximum melt rates, at >40 and ~320 m^2/yr are significantly less than that for mid-ocean ridge spreading (700 m^2/yr at 5 cm/yr), demonstrating the control of rift velocity in these passive spreading models, and supporting the case for more active rift mechanisms (e.g. plumes) in explaining voluminous volcanism during continental breakup.

Another important aspect of these models is they incorporate a mixed-heating approach, with reasonable internal heat production values, and a large non-adiabatic temperature gradient across the mantle. Lowman and Jarvis (1999) show this to be important for the timescale of mantle heating beneath a supercontinent; and the ratio of internal to basal heating is also important for the style of mantle flow—e.g. plumes do not operate in insulated, purely internally heated systems. Indeed, that was the starting point for the models of Coltice et al. (2007) who specifically looked at models without mantle plumes. We have not explored the effects of changing this ratio further here, but have adopted sensible values for internal/basal heating (Table 1), and thus the style of flow we observe here for Earth-like conditions should be applicable to mantle advection beneath large continents. Furthermore, systems with significant internal heat generation and temperature-dependent viscosities tend to have internal temperatures which are strongly controlled by these effects—mitigating to a large degree the limitations of a 2D Cartesian geometry, at least from the point of view of the thermal characteristics of the system (O'Neill et al., 2007). Clearly the 2D limitation is a dynamical constraint, but the behaviour of such systems imitates 3D simulations in most important respects, and 3D global spherical simulations at the resolution required to image individual plumes is very computationally demanding (e.g. Zhong et al., 2007). The fundamental conclusions of this paper are unlikely to be affected by the geometry of the problem.

So to return to our first question: subduction cannot cool efficiently cool the subcontinental upper mantle for continental extents >3000 km. However, in a strongly temperature-dependent regime, with drifting continents (in respect to the mantle), the temperature contrasts associated with this are less than ~50 °C. For continental extents >6000 km, advection of material from beneath the continent is not viable, and small-scale drip instabilities form in the subcontinental lithosphere, precipitating non-subduction related convection beneath the continent, which modulates any further temperature increases.

The second question we posed is what is the relationship between supercontinent position and rising mantle plumes? At one extreme we have the argument that plume heads initiate continental breakup (Hill, 1991). This does not explain why plumes should be concentrated beneath a supercontinent, if one accedes that indeed they are. The idea that subduction zones lock onto continental margins, and this marginal subduction concentrates upwellings into continental interiors, has antecedents going back to the models of Gurnis (1988). Indeed, Lowman and Jarvis (1999) used the relationship that subduction often occurs at continental margins as a starting condition for their models. However, in most simulations at realistic Rayleigh numbers (*Ra*), and in a mobile-lid convection regime (i.e. strongly temperature dependent viscosity with fault zones), subduction zones are not static but evolve constantly through time (e.g. Zhong and Gurnis, 1995). On top of that, at high *Ra*, upwellings are intrinsically unsteady, and large differential motions can occur between the mantle and the surface continents. Thus the relationship between continental margins, subduction, and continental interior plumes is non-trivial in more complicated dynamical regimes, and here we have demonstrated that the suggested qualitative relationship between subduction zones and margins, and plumes and continental interiors, holds quantitatively over long-timescale simulations and a large range of continental configurations. Over long period simulations, the deep mantle flow field beneath continental margins is dominated by downgoing subducting slabs, and the flow field beneath the continental interior distal to these slabs, and in intra-oceanic domains, is dominated by rising plumes. This relationship holds for all continents larger than 3000 km across; though small-scale convection becomes increasingly important for continents larger than ~6000 km.

Honda et al. (2000) showed that supercontinents can affect the dominant degree flow pattern in the mantle, and Zhong et al. (2007) demonstrated that placing a supercontinent over a degree-1 mantle downwelling—akin to supercontinent formation over closing ocean basins—can cause a reconfiguration of the mantle flow pattern. The supercontinent causes a disruption to lateral flow in the upper thermal boundary layer—the plates—and a new downwelling forms at the edges of the supercontinent. In response to this, an upwelling initiates beneath the continent, which could potentially lead to supercontinent dispersal and massive volcanism. This sequence is analogous to the observations in these models of stable subduction at the edges of a supercontinent, and a focussing of plumes in the stable continental interior and in the intraoceanic regions. The behaviour of dense chemical layers in these models is also influenced by the dynamics of descending slabs at the CMB. McNamara and Zhong (2005) showed that the structure of the thermochemical piles beneath Africa and the Pacific is a product of subduction history in these regions for the last 119 Ma. These piles have already been demonstrated to influence plume stability (Jellinek and Manga, 2004; McNamara and Zhong, 2004), and we find a large correlation between where these features occur and plume position in our models. In line with previous work, we also find these piles anchor plumes into position, and migration of plume conduits at the CMB is primarily related to movement of the thermochemical piles themselves. What is more, plume initiation is strongly correlated with the edges of these features, as this is where local thickening of the lower thermal boundary layer is likely to be focussed. This behaviour supports the idea of Burke et al. (2008) that most hotspots initiate at the edges of low shear wave anomalies beneath Africa and the Pacific.

5. Conclusions

The formation of supercontinents can have a first order effect on mantle flow. Supercontinents restrict the distribution of subducting slabs, which show a strong correlation with continental margins due to the migration of subduction zones to these regions. This spatial correlation then impacts the distribution mantle return flow, and specifically plumes, to continental interiors or intraoceanic settings away from descending slabs. This conclusion is robust for continents larger than 3000 km. At scales over 8000 km, the comparable advection and diffusion timescales results in inefficient lateral advection of material beneath a continent results in the formation of drip instabilities beneath the interior continental lithosphere, resulting in small scale convection cells and efficient heat transport beneath the continental interior. The spatial influence on subduction also affects the distribution of lower mantle thermochemical anomalies with respect to the supercontinent—stable configurations occur beneath the continental interior or in the intraoceanic regions, and these are coupled tightly with plume initiation, which occurs largely at the edges of such blobs, and with plume stability, which is enhanced by the anchor they provide, as previously found by Jellinek and Manga (2004). The supercontinent heating mechanisms discussed in this paper (warm plume-fed asthenosphere, subduction isolation, and plume clustering beneath a supercontinent) are not in themselves sufficient to explain supercontinent breakup volcanism volumes, for externally forced rifting. This suggests a more direct involvement of the mantle in supercontinent breakup (Ernst, 2009-this issue), and proposes the ultimate question of the cause of supercontinent breakup volcanism involves a complex interaction between deep mantle processes, non-linear mantle dynamics, and continental rift mechanisms.

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