Focussing of stress by continents in 3D spherical mantle convection with self-consistent plate tectonics

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[1] Previous mantle convection studies with continents have revealed a first-order influence of continents on mantle flow, as they affect convective wavelength and surface heat loss. In this study we present 3D spherical mantle convection models with self-consistent plate tectonics and a mobile, rheologically strong continent to gain insight into the effect of a lithospheric heterogeneity (continents vs. oceans) on plate-like behaviour. Model continents are simplified as Archaean cratons, which are thought to be mostly tectonically inactive since 2.5 Ga. Long-term stability of a craton can be achieved if viscosity and yield strength are sufficiently higher than for oceanic lithosphere, confirming results from previous 2D studies. Stable cratons affect the convective regime by thermal blanketing and stress focussing at the continental margins, which facilitates the formation of subduction zones by increasing convective stresses at the margins, which allows for plate tectonics at higher yield strength and leads to better agreement with the yield strength inferred from laboratory experiments. Depending on the lateral extent of the craton the critical strength can be increased by a factor of 2 compared to results with a homogeneous lithosphere. The resulting convective regime depends on the lateral extent of the craton and the thickness ratio of continental and oceanic lithosphere: for a given yield strength a larger ratio favours plate-like behaviour, while intermediate ratios tend towards an episodic and small ratios towards a stagnant lid regime.
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1. Introduction

[2] Continents impose a strong heterogeneity in the upper thermal boundary layer of the Earth's mantle - thermally and mechanically, which has a strong effect on surface heat flow [*Lenardic et al.*, 2005; *Cooper et al.*, 2006; *Grigné et al.*, 2007] and enforces flow reorganisation [e.g., *Lowman and Jarvis*, 1993, 1999], that affects convective wavelength [e.g., *Zhong and Gurnis*, 1993; *Guillou and Jaupart*, 1995; *Phillips and Coltice*, 2010].

[3] Although giving important insights into the physics of the Earth, all models used in these studies were simplified either by the lack of oceanic plates or by the assumption of simple rheologies or rigid, immobile continents or 2D geometry. The combination of both - plate tectonics and mobile continents - in 3D still remains a challenging task. The only study we are aware of that considers both features is the one of *Yoshida* [2010]. In his study he investigated the effect of weak continental margins (like *Lenardic et al.* [2003] in 2D), which surround the stronger interior and protect it against drastic deformation. Nevertheless he could model only a part of a sphere, which does not capture long wavelength flow components, and a short integration time.

[4] In this study we present a set of 3D spherical simulations with self-consistently generated plate tectonics that interacts with a continent floating at the top of the mantle. We concentrate on the question of how simple continents representing the strong Archaean cratons affect the global strength of the lithosphere by investigating thermal and mechanical aspects associated with the presence of continents and which depend on their lateral and vertical extent. In doing so we first describe our model in section 2. In sections 3–4 we present our results, which we discuss in section 5.

2. Numerical Model

[5] Our model is based on the one described by *van Heck and Tackley* [2008]; numerical details are given by *Tackley* [2008]. In order to integrate continents into the set of equations a modified momentum equation is necessary, such that the resulting (non-dimensionalized) Boussinesq equations are:

$$\vec{\nabla} \cdot \vec{u} = 0 \quad (1)$$

$$-\vec{\nabla} P + \vec{\nabla} \cdot \left[\eta\left(\partial_i u_j + \partial_j u_i\right)\right] = Ra(T - RC)\vec{e_r} \quad (2)$$

$$\partial_t T + \vec{u} \cdot \vec{\nabla} T = \nabla^2 T + Q, \quad (3)$$

where $\vec{u}$, P, η, T, C, $\vec{e_r}$ and t represent velocity, pressure, viscosity, temperature, composition, radial unit vector and time respectively. The three controlling parameters are the internal heating rate Q, the buoyancy ratio R, which is the ratio of the density difference $\Delta\rho$ of continental material to the thermal density variation $\rho\alpha\Delta T$ and the Rayleigh number $Ra = \alpha g D^3 \Delta T \rho / \kappa \eta_0$. In these definitions α is the thermal expansivity, ΔT the temperature drop across the lithosphere, g the gravitational acceleration, D the mantle thickness, κ the thermal diffusivity and η_0 a reference viscosity obtained at $T = 1$.

2.1. Continent Formulation

[6] Continental material is represented by a continuous compositional field ($0 \le C \le 1$, with $C = 1$ being continent), which influences the flow, as the buoyancy ratio $R < 0$ for a

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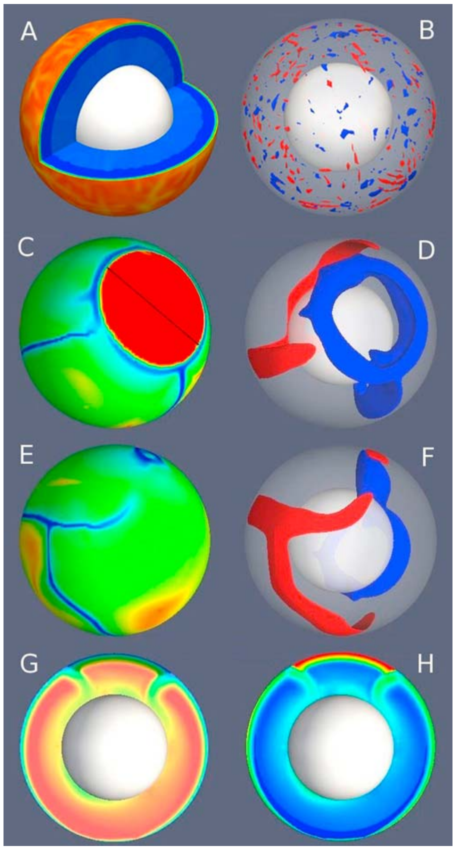

Figure 1. Comparison of 2 cases without a craton and a craton of the size of present-day Asia. The surface yield strength is $\sigma_Y \approx 360$ MPa. (a) Viscosity field without a craton. Red colour represents highly viscous, blue colour weaker material. (b) Contour plot of the residual temperature. Red contour: temperature 0.1 hotter than horizontal average. Blue contour: temperature is 0.1 colder. (c–f) show the same figures with a craton included for the front and back half of the sphere. (g–h) 2D slices of temperature and viscosity field, respectively, along the black line specified in Figure 1c. The craton is visible as grey transparent material in Figure 1g.

buoyant continent (see equation (2)). Following *Lenardic et al.* [2003] we simplify the model continents as Archaean cratons. These are initially spherical frustums with a homogeneous composition. After initialization, cratons are treated self-consistently and differ from normal mantle only in terms of buoyancy and rheological properties (see below). The compositional field is tracked by using the tracer ratio method [*Tackley and King*, 2003].

2.2. Rheology

[7] The viscosity η is given by an Arrhenius law to account for its temperature dependence in combination with a composition-dependent pre-factor for considering viscosity differences between continental and oceanic material:

$$\eta_T(T, C) = \exp[\ln(\Delta\eta_C)C] \cdot \exp\left[\frac{\tilde{E}_A}{T+1} - \frac{\tilde{E}_A}{2}\right], \qquad (4)$$

where $\Delta\eta_C$ and $\tilde{E}_A$ define the ratios $\eta(T, 1)/\eta(T, 0)$ and $\eta(1, C)/\eta(0, C)$, respectively. $\tilde{E}_A$ is set to 23.03, which allows for 5 orders of magnitude viscosity variation in the interval $0 \leq T \leq 1$. This is probably less than on Earth, but still allows for the formation of a rigid lid on top of the mantle [*Solomatov*, 1995]. Plastic yielding is considered to be the weakening mechanism necessary to observe plate-like behaviour [*Moresi and Solomatov*, 1998; *Tackley*, 2000a], i.e., material behaves viscously as long as the stresses remain below the yield stress (σ_Y), which increases linearly with depth and has a similar composition-dependence as the viscosity (see equation (5)). If the yield stress is reached, material deforms plastically and the viscosity is decreased to the yield viscosity η_Y:

$$\eta_Y = \frac{\sigma_Y(d, C)}{2\dot{\epsilon}} = \exp[\ln(\Delta\sigma_Y)C] \cdot \frac{\sigma_Y(0,0) + d \cdot \sigma_Y'}{2\dot{\epsilon}}. \qquad (5)$$

Here d is the depth and $\dot{\epsilon}$ is the 2nd invariant of the strain-rate tensor. $\Delta\sigma_Y$ defines the ratio $\sigma_Y(d, 1)/\sigma_Y(d, 0)$ to account for different yield stresses of continental and oceanic material. σ_Y' is the yield stress gradient with depth, which is fixed at 0.83 MPa km^{-1}, while the surface value is to be varied. The effective viscosity η_e is then given by the average of the two contributions from equations (4) and (5): $\eta_e(T, C, \sigma) = [1/\eta_T + 1/\eta_Y]^{-1}$.

[8] A low-viscosity asthenosphere, which is thought to improve plate-like behaviour [*Tackley*, 2000b; *Richards et al.*, 2001], is implemented following [*Tackley*, 2000b]: Once material is hotter than its solidus temperature, its viscosity is reduced by a factor $\Delta\eta_M = 0.1$. Here a linear increase of the solidus with depth is assumed: $T_{sol}(d) = T_{sol}(0) + d \cdot T_{sol}'$. The surface value $T_{sol}(0)$ is set to 0.6 and the depth gradient T_{sol}' is 2, which corresponds to 800 K and 0.9 K km^{-1} assuming a temperature drop of 1300 K across the lithosphere.

3. Simulations and Results

[9] All cases were run in 3D spherical geometry on a Yin-Yang grid [*Tackley*, 2008] with resolution 64 × 192 × 32 × 2, but one test case with resolutions 96 × 288 × 48 × 2 and 128 × 384 × 64 × 2 did not show significant changes. As there is not yet a strict consensus when plate tectonics began and if some cratons already existed at this onset time [e.g., *Stern*, 2005; *Condie and Kröner*, 2008], it is reasonable to assume pre-existing cratons, which are inserted at the beginning of the simulation. The Rayleigh number in all cases is 10^6 (based on a reference viscosity of $\eta_0 = 10^{23}$ Pa s), the internal heating rate Q is 20.5 ($\approx 5 \times 10^{-12}$ W kg^{-1}), and the density difference between continental and normal mantle material is about -100 kg m^{-3} (R = -0.4). The surface boundary is isothermal, while the core-mantle boundary is insulating. Both boundaries are free slip.

[10] In order to assure the longevity of cratons since the late Archaean we follow the results of various studies in 2D cartesian geometry [e.g., *Lenardic and Moresi*, 1999; *Lenardic et al.*, 2003]. Their basic result is that the buoyancy of continental lithosphere cannot assure craton stability. Instead, a rheologic contrast is much more important - in particular higher viscosity and yield strength, which can by justified with the relative dehydration of cratonic lithosphere [e.g., *Karato*, 2010]. We found that these stability criteria do not change significantly for 3D spherical geometry and that craton longevity can be achieved if the viscosity of cratonic

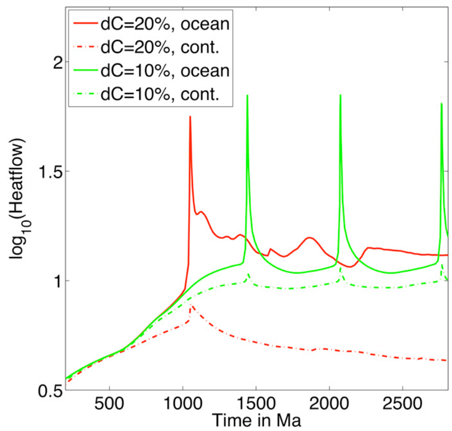

Figure 2. Time series of mean surface heat flow separated into oceanic (solid lines) and continental parts (dashed lines), used to demonstrate the thermal blanketing effect. Different colours indicate different craton thicknesses (green: 10% of mantle depth, red: 20% of mantle depth).

lithosphere is two orders of magnitude and its yield strength one order of magnitude higher than for oceanic lithosphere (similar to, e.g., *O'Neill et al.* [2008]).

4. Influence of Cratons on the Convective Regime

[11] It has been shown previously how the convective regime (stagnant lid vs. plate-like) depends on the yield strength of the lithosphere [e.g., *Moresi and Solomatov*, 1998; *Tackley*, 2000a, 2000b; *Stein et al.*, 2004; *van Heck and Tackley*, 2008]. However, none of these models included continents and a major shortcoming was the fact that the plate-like regime was only observed for a yield stress of 100–200 MPa, which is significantly less than observed in laboratory experiments [e.g., *Kohlstedt et al.*, 1995]. We now investigate how the presence of a craton affects which regime develops for a broad range of yield stresses.

[12] Figure 1 clearly shows that the presence of a craton alone can change the convective regime. Figure 1 displays the viscosity field (Figure 1a) and the residual temperature (Figure 1b, see figure caption for explanation) for a case without a craton. The surface viscosity is very high and quasi uniform, but it is very low in the interior due to the high temperature. Spatial variations in temperature are small and randomly distributed. Figures 1c–1h shows the result for the same case, but with a craton of the size of present-day Asia (8.6% of surface area). Large viscosity contrasts exist between the oceanic plates and the localized plate boundaries. Additionally a transitional regime can occur, which is characterized by short-lasting mobilization events that interrupt periods of stagnant lid. In the present study it occurs at intermediate lithospheric strength occur with an almost constant period that decreases with increasing size of the craton.

[13] More generally one can expect two competing mechanisms that control which regime develops. On the one hand continents act as thermal insulators. This does not mean that continental material transports heat less efficiently [*Lowman and Jarvis*, 1999], but that their great thickness reduces the conductive heat transport [*Lowman and Jarvis*, 1993], such that the heat flow over continents can be much smaller than in oceanic regions. This can easily be seen in Figure 2: the continental heat flow is always smaller than the oceanic heat flow and the difference between them increases with increasing continental thickness. Thus, thermal blanketing heats up the subcontinental mantle and leads to a decrease in viscosity and convective stresses. Consequently it is harder to exceed the yield stress and create plate boundaries. On the other hand continental and oceanic lithosphere have significantly different thicknesses and rheologies. These contrasts may focus stresses at the continental margins and allow for yielding in the lithosphere [*Smith and Mosley*, 1993].

[14] Both mechanisms are strongly affected by the geometric dimensions of the continent. For this reason, changes in the lateral continental extent and the thickness ratio of continental and oceanic lithosphere will be investigated in the following sections.

4.1. Continental Size

[15] *Lenardic et al.* [2005] developed a scaling which shows that thermal blanketing - for the parameter combinations used in the present study - is mainly controlled by continental coverage (and not by continental thickness or conductivity. This dominance of the coverage effect also holds if heat producing elements are concentrated in the continental crust [*Cooper et al.*, 2006]. In order to investigate how variations in continental coverage affect thermal insulation and the convective regime, we ran a set of cases with varying yield strength of oceanic lithosphere and varying lateral size of the craton (i.e., changing the total continental coverage). Results are given in Figure 3a: with no craton (relative continental area is zero) the transition between plate-like and stagnant-lid behaviour occurs at a relatively low yield strength of 240–300 MPa (note that the episodic regime does not exist in this case). Once a craton is included the plate-like regime can be observed for higher yield strength (up to 480–540 MPa), even if the continent is small. The episodic regime now occurs for a broad range of intermediate yield strengths up to 700–800 MPa, where the exact value increases slightly with the continental area. The transition between plate-like and episodic behaviour is hardly affected by continental size, as long as continental coverage does not exceed ≈20%. For larger continents this transition is observed at slightly smaller yield stresses (≈350–400 MPa). This may be explained by a trade-off between thermal blanketing and the heterogeneity in the lithosphere. The former counters continuous plate-like behavior, but it can overcome the advancing effect of stress focussing only for the largest continents.

4.2. Continental Thickness

[16] Seismic tomographic results show that cratons can be as thick as 200–250 km [e.g., *Gung et al.*, 2003], which agrees well with studies on electrical conductivity data [*Hirth et al.*, 2000]. However, these estimates cannot be used directly in this study as the Rayleigh number in the model, which determines the thickness of the thermal boundary layer, is a factor of about 20 lower than expected for the Earth ($Ra_E \approx 2 \times 10^7$ [*Schubert et al.*, 2001]): assuming an oceanic boundary layer thickness of $\delta_E = 100$ km

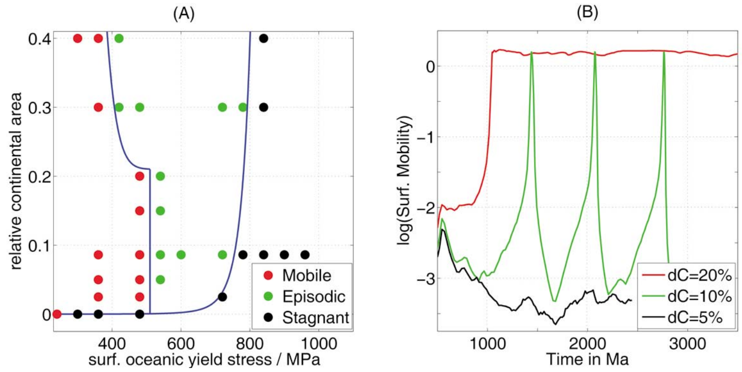

Figure 3. (a) Regime diagram in the parameter space of oceanic yield strength (at the surface) and relative continental area. Each dot represents one calculation, while its color indicates the developed convective regime: (red) mobile/plate-like, (green) episodic, (black) stagnant lid. Blue lines separate different regimes approximately. (b) Time evolution of the surface mobility [*Tackley*, 2000a] for cases with $\sigma_Y \approx 480$ MPa and 8.6% relative continental area. Colors indicate the respective continental thickness (dC) relative to the mantle depth.

for the Earth, its model analogue needs to be scaled up as $\delta = (Ra_E/Ra)^{1/3} \cdot \delta_E \approx 2.7 \cdot \delta_E$, and the same scaling factor must be applied to the continental boundary layer. Therefore, we chose a craton thickness of 20% mantle depth for all simulations presented so far, which corresponds to 580 km in the model and 215 km if rescaled to the Earth.

[17] To investigate the influence of continental thickness on the convective regime we compared three simulations with a craton thickness of 20%, 10% and 5% mantle depth. The oceanic yield stress is set to $\sigma_Y \approx 480$ MPa. Results are displayed in Figure 3b via the time series of surface mobility, i.e., the ratio of surface rms-velocity and volume rms-velocity [*Tackley*, 2000a]. A mobility of about 1 indicates plate-like behaviour, while very small mobilities are characteristic of a stagnant lid. It can easily be seen that an increase in the thickness ratio of cratonic and oceanic lithosphere leads to the evolution of plate tectonics. With the smallest thickness, the craton is completely embedded in the oceanic boundary layer, such that the underlying mantle does not feel the heterogeneity in the lithosphere; therefore it behaves similarly to cases with a homogeneous lithosphere and a stagnant lid develops. For the intermediate case the craton thickness slightly exceeds the oceanic boundary layer thickness and the heterogeneity starts to affect the underlying mantle. However, the effect is not strong enough to allow for plate boundaries continuously in time. Consequently, the episodic regime develops. Only for the thickest root is the thickness difference large enough for the continuous presence of plate tectonics.

5. Discussion and Conclusion

[18] In the present study a heterogeneous lithosphere containing oceanic and continental parts has been shown to have an important effect on plate-like behaviour. In general a lithosphere containing both shows a stronger tendency to develop plate tectonics than a homogeneous lithosphere. The presence of a craton increases the critical yield strength below which plate tectonics occurs by up to a factor of 2, which gives a better agreement with results from laboratory experiments on mantle rocks [*Kohlstedt et al.*, 1995], although still a little smaller. This might be explained by the fact that a Rayleigh number of 10^6 had to be used in this study due to computational reasons. Increasing Ra by an order of magnitude would result in more vigorous convection, which may increase convective stresses according to the scaling law $\sigma \propto Ra$ for a linearly depth-dependent yield stress, proposed by *Moresi and Solomatov* [1998]. Accordingly, it could be easier to reach the limit of plastic yielding at higher Ra.

[19] Moreover, it was found that the thickness ratio of cratonic and oceanic lithosphere affects the development of plate tectonics, with a sufficiently high ratio of about 2 favouring its continuous presence. This ratio is likely to match the settings on present-day Earth.

[20] Although the consideration of a simple continent in a 3D model with self-consistent plate tectonics is a next step towards a consistent geodynamical model of the Earth, our approach is still simplified. In future models multiple continents will have to be considered, which has so far only been done for cases without plate tectonics [e.g., *Zhang et al.*, 2009; *Phillips and Coltice*, 2010]. Increasing the number of continents increases the number of heterogeneities in the lithosphere. However, the question of whether several heterogeneities multiply the effect of a single one or if they share its potential is not trivial to answer. Does the lithosphere break more often then? Or does the amount of stress focussing per margin decrease with the number of margins, such that the focussing effect is too small?

[21] Another shortcoming of our simulations is the lack of active upwellings due to an insulating core-mantle boundary. Rising plumes might have important effects on the observations presented here, as they may induce additional stresses to the lithosphere. Furthermore, *Heron and Lowman* [2010] observed in their model with core heating and prescribed plates that plumes do not preferentially form below supercontinents, which they explain by a minor role of thermal insulation on the thermal field of the mantle. This aspect has to be addressed in future studies with a spherical, self-consistent model.

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