Available online at www.sciencedirect.com
ScienceDirect
Lithos 102 (2008) 12–24
LITHOS
www.elsevier.com/locate/lithos

Dynamics of cratons in an evolving mantle

C.J. O'Neill [a,*], A. Lenardic [b], W.L. Griffin [a], Suzanne Y. O'Reilly [a]

[a] GEMOC ARC National Key Centre, Department of Earth and Planetary Science, Macquarie, University, Sydney, Australia, NSW, 2109
[b] Department of Earth Science, MS-126, Rice University, Houston, Texas, USA, 77003

Received 17 October 2006; accepted 30 April 2007
Available online 10 May 2007

Abstract

The tectonic quiescence of cratons on a tectonically active planet has been attributed to their physical properties such as buoyancy, viscosity, and yield strength. Previous modelling has shown the conditions under which cratons may be stable for the present, but cast doubt on how they survived in a more energetic mantle of the past. Here we incorporate an endothermic phase change at 670 km, and a depth-dependent viscosity structure consistent with post-glacial rebound and geoid modelling, to simulate the dynamics of cratons in an "Earth-like" convecting system. We find that cratons are unconditionally stable in such systems for plausible ranges of viscosity ratios between the root and asthenosphere (50–150) and the root/oceanic lithosphere yield strength ratio (5–30). Realistic mantle viscosity structures have limited effect on the average background cratonic stress state, but do buffer cratons from extreme stress excursions. An endothermic phase change at 670 km introduces an additional time-dependence into the system, with slab breakthrough into the lower mantle associated with 2–3 fold stress increases at the surface. Under Precambrian mantle conditions, however, the dominant effect is not more violent mantle avalanches, or faster mantle/plate velocities, but rather the drastic viscosity drop which results from hotter mantle conditions in the past. This results in a large decrease in the cratonic stress field, and promotes craton survival under the evolving mantle conditions of the early Earth.

Keywords: Mantle convection; Cratons; Mantle lithosphere; Mantle avalanche

1. Introduction

Evidence for the thermal evolution of the mantle is primarily preserved on cratons, which are defined as regions of long-lived crust and lithosphere that have resisted or avoided significant deformation and recycling since at least the Precambrian, in some cases for billions of years (e.g. Lenardic and Moresi, 1999).

Their tectonic stability is fundamental to their long-term survival (Lenardic et al., 2003), particularly in the case of Archaean cratons, and many well-documented examples of their exceptional tenacity exist. For instance, the most dramatic ongoing orogeny – the Himalayas – primarily deformed the leading edge of the Indian continent which consists predominantly of Tethyan margin assemblages of Cambrian age (Myrow et al., 2003). A large portion of the southern Indian continent, consisting primarily of Archaean Indian craton, has avoided significant internal deformation. Similarly the creation of the Urals on the edge of Baltica deformed the Neoproterozoic arc assemblages but left the 'core' of Baltica – the Archaean craton – undeformed (Puchkov, 2002). The three major African Archaean cratons – the Kaapvaal, the Congo and the West African – likewise have undergone minimal

* Corresponding author.
E-mail address: coneill@els.mq.edu.au (C.J. O'Neill).

0024-4937/$ - see front matter © 2007 Elsevier B.V. All rights reserved.
doi:10.1016/j.lithos.2007.04.006

internal deformation despite being involved in major orogenic episodes such as the Pan–African orogeny (Stern, 2004). More recently, the breakup of Gondwana has led to the disaggregation of many continental shields, but the Archaean cratons embedded in such shields, particularly the Indian craton and the Gawler craton on the southern edges of India and Australia respectively, and both directly adjacent to rifting, have been obdurate (e.g. Ball et al., 2003).

The interaction between the cratonic lithosphere and their environment – both through plate-driving forces and more directly the convecting mantle – is fundamental to understanding the long-term stability of cratons, particularly under the hotter, more energetic conditions of the early Earth. Modelling the conditions of the early Earth is non-trivial, however, and a number of critical factors are necessary for addressing the range of mantle behaviour expected in the past. These include depth-dependent viscosities, an endothermic phase transition at 670 km, and higher mantle heat production. Depth dependent viscosities strongly affect plate-mantle coupling, and can produce divergent styles of convection in the upper and lower mantle (Lenardic et al., 2006). An endothermic phase transition promotes mantle layering and gives rise to a phenomenon known as mantle avalanches — violent mantle overturn events which have been postulated to have been more severe early in Earth's history (Stein and Hofmann, 1994). Higher mantle heat production in the past would increase the vigour of convection, but also decrease internal viscosities, affecting plate–mantle coupling in a non-trivial manner. The interaction of these parameters in a plate tectonic-like system has not previously been addressed, particularly from the context of cratonic stability. The scope of this work is to understand the tectonic stability of cratons, in systems that faithfully reproduce the plate–mantle interaction, for 'Earth-like' convecting mantles, under a range of conditions applicable to cratons today and in their past — i.e. conditions relevant to cratons throughout their geological history.

1.1. Survival issues and previous modelling

A number of previous modelling studies have addressed the problem of cratonic root survival in a convecting mantle. The key observation addressed by such models is not that cratons survive for billions of years – the surface of the Moon is older – but that cratons survive so long despite most of the planet's surface, the oceanic plates, being recycled on timescales of around a hundred million years (Lenardic and Moresi, 1999). The survival of the cratonic crust is potentially due to two factors: the survival of the deep root zone supporting the crust (life-raft model, Jordan, 1978; Sengor, 1999; O'Reilly et al., 2001), or the intrinsic properties of the crust itself, compared to surrounding crust (e.g. greater strength, the crumple-zone model, Lenardic et al., 2000. Cratonic crust may have a greater strength than younger crust for a number of reasons. Crustal strength is a strong function of temperature (Maggi et al., 2000) — and low mantle heat flow into cratonic crust, due to the presence of thick cool roots, would result in relatively cooler crustal temperatures and a greater integrated crustal strength. Similarly, the re-distribution of heat producing elements (HPEs) in the crustal column may have a similar effect (Sandiford and McLaren, 2002).

Most workers, however, have focussed on the physical properties of the mantle lithosphere itself. Seismic evidence shows the existence of high-velocity roots beneath cratons to at least 250 km depth (Polet and Anderson, 1995; Gung et al., 2003; Niu et al., 2004; Priestley et al., 2006). Numerous lines of evidence suggest these deep root zones are as old the cratonic crust itself (Boyd et al., 1985; Richardson et al., 1984; see also Spetsius et al., 2002; O'Neill and Moresi, 2003). The refractory nature of Archean subcontinental lithospheric mantle means it is significantly more buoyant than asthenospheric mantle at similar depths (~2.5%, Poudjom Djomani et al., 2001; Deen et al., 2006; Lee, 2006). Buoyancy is primarily controlled by the Mg#, and the proportion of olivine to clinopyroxene + garnet. For representative Archean mantle lithosphere, this results in an average density of 3310 kg/m^3, compared to 3340 kg/m^3 for Proterozoic lithosphere, and 3360 kg/m^3 for Phanerozoic (Poudjom Djomani et al., 2001). This positive buoyancy means the cratonic lithosphere is intrinsically less susceptible to recycling or delamination than oceanic lithosphere, and is likely to survive the delamination of the continental lithosphere in an orogenic setting. A number of modelling studies have explored the contribution of root buoyancy to cratonic survival (Doin et al., 1997; Shapiro et al., 1999; Lenardic and Moresi, 1999; Cooper et al., 2006). The preliminary result of most of these studies is that while buoyancy might be a factor in root stability, it cannot account for the stability of cratons: Buoyant roots either spread out under their own gravity, or re over-thickened and entrained, or even whole-sale recycled, into mantle. Such behaviour often results in the complete destruction of cratons, but, more importantly, the degree of internal deformation and ongoing tectonism means that such lithosphere cannot be classified as cratonic. Cratons are defined by their ability to resist tectonism for billions of years, the consensus from previous modelling highlights the role of highly viscous cratonic roots to stabilise cratonic blocks and enable craton survival over geological timescales (Pollack, 1986; Doin et al., 1997; Shapiro et al., 1999; Lenardic and Moresi, 1999).

The extreme depletion of cratonic roots suggests they are dehydrated. Hirth and Kohlstedt (1996) showed that water can reduce the viscosity of olivine aggregates by a factor of 140. In the extreme case of completely dehydrated roots, and water/volatile bearing asthenospheric mantle, the viscosity contrast can be over two orders of magnitude (Lenardic and Moresi, 1999), providing a physical basis for viscous cratonic roots.

Lenardic and Moresi (1999) showed that while craton viscosity may account for survival under present-day mantle conditions, it might not allow craton survival in the Archean. Further studies with more realistic rheologies (Lenardic et al., 2003) have shown that when brittle-plastic failure mechanisms are included, high root viscosities can enhance lithospheric stress resulting in the brittle failure of the cratonic lithosphere, and its recycling at subduction zones. However, the relative strength of cratons is a crucial factor. A number of studies have shown greater elastic lithosphere thicknesses for many Archean terranes (e.g. Poudjom Djomani et al., 2005). This is suggested to be due to the contribution of a strong elastic core in cratonic regions, in large part due to the upper mantle lithosphere. While a direct correlation with mantle lithospheric strength is difficult due to the complicated rheology of the continents, these observations are broadly consistent with a strong mantle lithosphere in cratonic regions compared to non-cratonic terranes. Lenardic et al. (2003) showed that for sensible values of root thickness, root viscosity, relative yield strength of the cratonic root and crust, and, importantly, weak non-cratonic continental regions which act as a "buffer-zone", cratonic stability can be realised for present-day mantle conditions.

1.2. Craton destruction events

A key observation in the discussion of craton stability is that not all cratons are indestructible, and a number of examples exist of the destruction of cratonic lithosphere. Clearly it is important to this work to understand the conditions under which craton integrity is compromised, and in what ways cratonic lithosphere is susceptible to destruction.

The Sino–Korean craton in eastern China contains diamondiferous Ordovician kimberlites, and xenolith, mineral concentrates and diamond inclusions all indicate a thick (>180 km) lithosphere at this time (Xu, 2001; Griffin et al., 1998b). Later Cenozoic tholeiitic–alkali basalts contain xenoliths indicating a thin, hot fertile lithosphere. The mechanism by which the lithospheric keel has been removed is unclear. A Triassic collision with the Yangtze block may have undermined craton's integrity (Xu, 2001). The mantle additionally may have been weakened by progressive refertilisation, from either subduction or magmatic sources, which left it susceptible to later ablative subduction or rifting.

The Archean Wyoming Province has no seismically discernible continental root at present (van der Lee and Nolet, 1997). However, the province contains numerous diamondiferous kimberlites of Devonian age (Griffin et al., 2004b). The root appears to have been strongly modified — possibilities include hydration due to the subduction of the Farallon slab (Humphreys et al., 2003), and it was probably destroyed in a later collisional orogeny (Eggler et al., 1988). The orogeny thrust wedges of Archean lithosphere in younger, more fertile lithosphere (Griffin et al., 2004b).

Diamondiferous kimberlites of mid-Paleozoic age are found in the Magan and Anabar provinces of the Siberian craton, indicating that these areas were underlain by a cold, refractory root 190–240 km thick at this time (Griffin et al., 1998a; Milashev, 1974). However in the Olenek province, to the northwest, Mesozoic kimberlite volcanism, which is devoid of diamond, indicates that the lithosphere is 50–60 km thinner (Griffin et al., 1998a). This is possibly due to Triassic volcanism and Devonian rifting in these regions.

Other notable thermo-magmatic events have not resulted in wholesale destruction of cratonic roots. The Bushveld igneous event in South Africa, had a marked local effect on the mantle lithosphere, resulting in a major P-wave anomaly (James and Fouch, 2002), resetting of the lithospheric age (Carlson et al., 2000), and probable destruction of pre-2 Ga diamonds in the mantle in vicinity of Premier (Griffin et al., 2003). However, numerous pipes in the Kaapvaal contain diamonds with Archean-age inclusions (Shirey et al., 2002; Griffin et al., 2004a), which indicates that, despite its magnitude, the Bushveld event only locally affected the lithosphere and did not result in wholesale lithospheric destruction. Similarly, the MacKenzie dyke swarm in northwest Canada did not affect diamonds in the Slave cratonic root (Helmstaedt and Gurney, 1995; Westerlund et al., 2003; Griffin et al., 1999). Nor did widespread Cambrian flood volcanism in northern Australia (Handley and Wingate, 2000) destroy diamonds in the lithospheric mantle beneath the Timber

Creek or Merlin kimberlites (assuming the diamonds predate this event).

To summarize, cratonic roots are rather robust to the effects of isolated tectono-thermal events. A key ingredient in most cases of cratonic lithosphere destruction is the prior refertilization of the lithosphere by metasomatic processes, either due to fluids from a subducting plate or from magmatic fluids. This makes cratons at plate boundaries particularly susceptible, due to a higher background fluid flux at such regions, and the propensity of weakened cratonic lithosphere to then respond to plate boundary forces, resulting in orogeny or rifting.

The picture emerging is that in order to destabilise a craton it must be weakened, and that cratons are resilient to distinct tectonic events. So we return to the original conundrum: plausible cratonic properties can provide stability for present-day conditions, but are insufficient to protect cratons under the more energetic mantle conditions of the Precambrian.

1.3. Precambrian mantle environment

In all previous studies, however, the question of Precambrian mantle dynamics has not been adequately addressed. Typically most studies have simulated Archean mantle conditions by increasing the system's Rayleigh number (a measure of convective vigour). However, heat production in the Archean mantle has been estimated to have been 3–4 times greater than today (see Turcotte and Schubert, 1982), which results not only in significantly more vigorous mantle convection, but also lower internal viscosities. Furthermore, the spinel → magnesiowustite + perovskite phase transition at 670 km is endothermic, and it has been shown that higher mantle temperatures may impede mass transfer across the 670 km discontinuity (Christensen and Yuen, 1985). This would promote layering between the upper and lower mantle, interspersed by mantle avalanche events, where cold slabs periodically breakthrough the 670 km discontinuity, and hot lower mantle material upwells in response (Christensen and Yuen, 1985). Episodic crustal formation in the Precambrian has been attributed to such events (Stein and Hofmann, 1994), but the effect of mantle avalanches on existing continental material has never been explored. The style of mantle avalanches, and indeed the coupling of the plates to the mantle, are strongly affected by the depth-dependent structure of the mantle. The low viscosity of the asthenosphere beneath many oceanic and continental regions is not fully understood (cw. Richards et al., 2001), but is thought to be due to the pressure and temperature dependence of mantle mineral phases, perhaps nfluenced by the presence of upper mantle melt (Schmeling and Bussod, 1996) or water (Hirth and Kohlstedt, 1996). Regardless of its cause, the effect of a low-viscosity layer on the surface dynamics of plates is

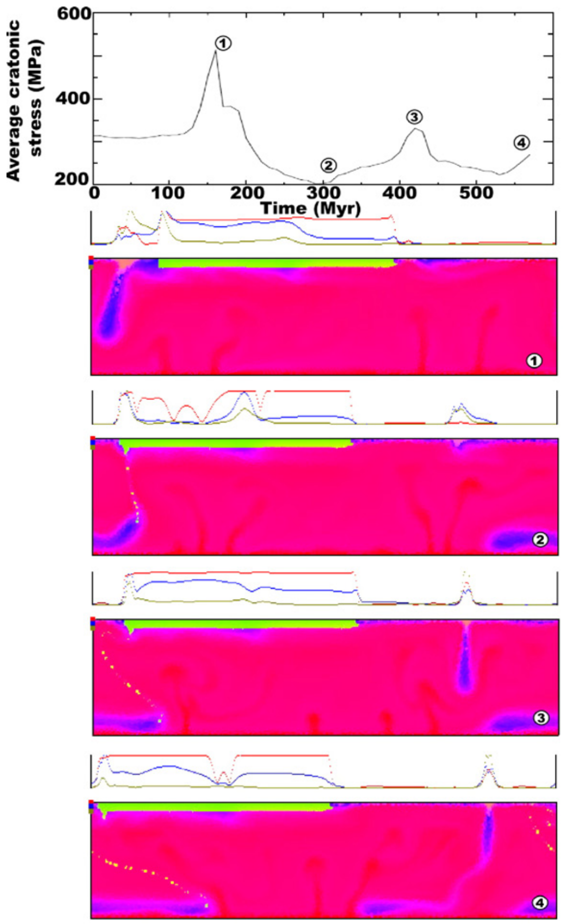

Fig. 1. Evolution of cratonic stress during a mantle breakthrough events. The simulations are for the whole mantle using appropriate nominal thermal parameters (Table 1), with a grid resolution 96 × 384, with additional grid compression (by a factor of 1.5) at the surface and at the 670 km discontinuity. Non-adiabatic temperature variations drive the flow, and the temperature field shown at times indicated by (1–4) reflects these non-adiabatic variations (the total non-adiabatic temperature contrast is 2555 K). The side boundary conditions are periodic, the viscosity is both temperature and depth dependent, and the highly viscous surface layers fail in a plastic manner (light coloured regions at the surface). The continent is modelled as a chemically distinct block, and the horizontal profiles are the normalised total deviatoric stresses (total magnitude, including both extensional and compressional components), at depths of 0, 115 and 230 km, positions shown by the coloured sampling points to the left of the temperature snapshots. The peaks in cratonic stress at points 1, 3 and 4 correlate with slab breakthrough of the 670 km discontinuity. The stress increase at these times is a factor of 2–3 higher than the average stress than at times without slab penetration of 670 km.

dramatic (Richards et al., 2001; Lenardic et al., 2006). Previous work has focussed on the oceanic regions, and has shown that 1) the partial decoupling of the plates from the mantle-generated stresses leads to far more stable plate configurations and long-lived subduction zones (Richards et al., 2001), and 2) the low viscous dissipation beneath plates leads to greatly enhanced plate dimensions, i.e. the generation of Pacific-type plates in mantle convection simulations (Lenardic et al., 2006). The decoupling effects of the asthenosphere is anticipated to affect continental regions in a similar way, i.e. decreased lithosphere stresses due to viscous decoupling with mantle stresses, and subsequent enhanced stability of continents. Thus the introduction of realistic mantle structures into simulation of cratonic stability results in two competing effects; either the more violent mantle dynamics, particularly in the past, adversely affect craton longevity, or the decoupling of the plates from mantle dynamics enhances their long-term stability.

Here we explore the stability and dynamics of cratons with plausible rheological properties within simulations of "Earth-like" mantle convection, incorporating surface plate, temperature and depth-dependent viscosities, and a mantle phase change at 670 km. In particular we assess how cratons survive under plausible range of dynamic regimes appropriate for the Precambrian mantle, based on our understanding factors of Earth-style convection today.

2. Methodology

The standard model configuration we employ is shown in Fig. 1. We model whole-mantle convection (i.e. 2890 km depth) in a 4×1 box, with periodic boundary conditions at the sides, and free-slip boundary conditions at the top and bottom. The default material properties for our standard model are given in Table 1.

The mantle in our model is a viscoplastic material, with a temperature-dependent viscosity varying over five orders of magnitude. The highly viscous surface plates deform primarily by brittle failure, where the plate fails in a plastic manner once a critical yield stress has been reached. The formulism can be found in Moresi and Solomatov (1998). We also include a strain-weakening mechanism. The mantle viscosity is also depth-dependent, with a low viscosity asthenosphere, and a high viscosity lower mantle. The exact values vary between models. Our default values are a decrease in asthenospheric viscosities by a factor of 0.1, and an increase in lower mantle viscosities from the temperature-dependent flow law of 10. No viscosity pressure-dependence was considered within each layer as for plausible parameters as this is a second order effect. We also simulate the spinel → perovskite + magnesiowustite phase change at ~670 km, in a similar fashion to Christensen (1995). The Clayperon slope of this transition strongly affects the dynamics of mass transfer between the upper and lower mantle, and we explore the effect of varying this parameter between reasonable bounds. The

Table 1

Symbol	Property	Value
ρ_m	Mantle density	3400 kg m^{-3}
g	Gravitational Acceleration	9.81 m s^{-2}
α	Thermal expansivity	3×10^{-5} K^{-1}
T_{CMB}	Temperature at the CMB	~4000+/−600 K
ΔT	Non-adiabatic T contrast[a]	2555 K
d_m	Depth if 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} (varies)
$\eta(T)$	T-dependent viscosity	10^{19}–10^{24} Pas^{-1} (FK approx[b])
–	Upper mantle visc contrast	$0.5*\eta(T)$ (default, varies)
–	Lower mantle visc contrast	$5*\eta(T)$ (default, varies)
dP/dT	Clayperon slope at 670 km	−2.5 MPa/K (default, varies)
$\Delta\rho/\rho$	Density change across transition	10%
μ	Friction coefficient	0.15 (varies for stronger cratons)

All values are the default unless explicitly stated otherwise. For references see Schubert et al. (2001) and references therein.

[a] This is the temperature contrast which drives convection, and is found by subtracting the adiabatic contribution (~1445 K at the CMB) from the CMB temperatures. Roughly equivalent to the temperature drop across the upper and lower thermal boundary layers (~1300 K for the oceanic lithospheric, presumably similar at the CMB).

[b] Frank–Kamenetski approximation viscosity profile — see Lenardic et al. (2003) for details.

values, or range of values, we adopt in this study are listed in Table 1. The basal Rayleigh number for the mantle, using the values listed in Table 1, and assuming a non-adiabatic temperature gradient of 2555K, and an average mantle viscosity of 10^{21} Pa s, is 6×10^{7}. The cratons are modelled are chemically distinct blocks with distinctive density and rheology compared to the convecting mantle. We also the effect varying the cratons intrinsic properties in the following sections. The main problems we address by varying these parameters are: 1) Are cratons stable in Earth-like convecting mantles for plausible rheological parameters? 2) What range of mantle dynamics and surface effects can be expected for the hotter conditions of the Precambrian? 3) What are the dominant effects back through time, particularly with regard to surface stress, and how do they affect craton stability in the past?

We use a particle-in-cell finite element code (Ellipsis, Moresi et al., 2003) to model the evolution of chemically distinct continents in a convecting mantle. The code solves the standard convection equations, i.e. Stokes equation for incompressible flow, and energy equation, subject to the incompressibility constraint (see Moresi et al., 2003 for details). The equations are solved in a discrete form on an Eulerian finite element grid. Ellipsis extends the traditional finite-element approach, however, by using Langrangian particles as integration points, allowing the simulation of chemically distinct phases and history-dependent variables. All simulations are run on a 98×384 grid, with additional grid compression (by a factor of 1.5) in regions of special interest (the surface and 670 km). The initial condition for all simulations is a thermal configuration from a prior low-resolution simulation run to a statistical steady state for the same parameters as the high-resolution simulations.

3. Results

Fig. 1 shows the evolution of stress in a stable craton through time. In this example the mantle undergoes an avalanche event, where cool dense "oceanic" style lithosphere punches through the 670 km discontinuity and recycled into the lower mantle. The surface stress shown is the total deviatoric stress, which is scalar and best represents the total stress field of the craton, regardless of whether the stress is extensional, compressional, or some combination. As previously found (Davies, 1995) in models incorporating realistic oceanic lithosphere, the 670 km discontinuity acts as only a minor impediment to mass transfer between the upper and lower mantle, and only shallowly-dipping slabs stagnate at the transition, and even then, only when their oceanic trench is exhibiting significant roll-back. This is consistent with previous modelling (Davies, 1995) which found models with strong oceanic plates can punch through the transition far more easily than in equivalent isoviscous models, in particular when the slabs are steeply dipping. For instance, the steeply-dipping oceanic subduction in Fig. 1, which is stable over 250 Myr, is repeatedly responsible for episodes of slab penetration over the course of the simulation. The phase transition does, however, introduce an additional non-linearity into the system, so even while the subduction zone exhibits long-term stability, slab breakthrough into the lower mantle is rather time-dependent. Slab-breakthough into the lower mantle has a marked effect on the cratonic stress field. The slabs affect the surface stress field not only through greater slab pull (stresses propagated to the surface through an intact slab); but also through what has been dubbed "slab-suction" (Conrad and Lithgow-Bertelloni, 2002). Here the viscous coupling between the down-going slab and the mantle flow field effectively induce large mantle tractions at the surface, resulting in large peaks in surface stress during mantle avalanche events. These peaks are important as they are the times when the cratonic lithosphere is most likely to fail as a result of the global tectonic configuration.

A craton can be destroyed in these examples in a number of ways, as illustrated in Fig. 2. Far field forces are experienced by the cratons in two ways: promulgation of stress through the surface plates, and the coupling of cratonic lithosphere with the subjacent mantle. The thermal state of the cratonic lithosphere tends to be cooler than adjacent oceanic terranes, and as a result a cool thermal boundary layer (TBL) tends to extend beneath the cratonic root. This TBL tends to protect the chemical lithosphere from subcratonic shear, as most deformation is accommodated in the lower portions of the TBL. In the event of hot asthenosphere in contact with the chemical layer, the low viscosity asthenosphere again tends to accommodate most significant deformation. Vertical flow may also affect cratonic stress; however, we find the buoyancy force associated with subcratonic 'drips' insufficient to produce deformation. Plumes, on the other hand, produce significant buoyancy forces, and the loci of rifting is often associated with plume activity, though the role of plumes versus surface promulgated stress is difficult to delineate.

The different craton-destruction mechanisms observed in Fig. 2 include 1) *Whole-sale recycling into the mantle*. The craton's density, viscosity and/or ield strength are insufficient to prevent their wholesale destruction. Such a scenario is unlikely for reasonable cratonic properties. However, if a craton is previously thinned and weakened,

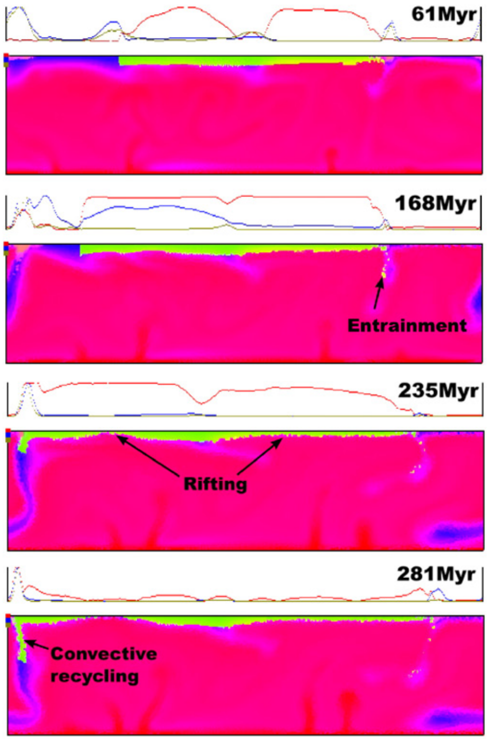

Fig. 2. Craton destruction mechanisms. Shown are four snapshots of a simulation with similar setup to Fig. 1, and the nominal values outlined in Table 1, except for the continent having the same viscosity relationship as the mantle (i.e. no viscosity contrast), and the same yield strength parameters as the oceanic lithosphere. Viscous thinning of the continent is followed by entrainment of portions of the continent adjacent to subduction zones (168 Myr), rifting of the continent in response to the surface stress field at 235 Myr, and finally convective recycling of some of the thinned continental lithosphere at 281 Myr.

its buoyancy – which depends on its density as well as the volume of material – may be insufficient to prevent convective recycling. By this stage, however, such material would not be labelled a craton. 2) *Entrainment into the mantle*. High stresses near subduction zones result in either brittle failure or viscous deformation of the edges of the cratons, resulting in their long-timescale recycling into the mantle. Entrainment is difficult to model, but in general depends on long-lived subduction zones affecting the craton's edges for very long periods of time; and even then it may not affect parts of the craton distal to the subduction zone. 3) *Rifting/orogenesis*. Cratons by definition avoid destruction by active tectonism for billions of year. If the physical properties of the continental lithosphere are insufficient to prevent a significant amount of strain, the continent is, for all-purposes, non-cratonic. Naturally this depends not only on the properties of the craton, but on the stresses applied, and we explore the effect of these factors in turn.

3.1. Cratonic properties

A number of previous studies (see Lenardic and Moresi, 1999, and references therein) have concluded that while buoyancy may be an important factor, by itself it is not sufficient to prevent cratonic destruction, and we do not revisit this theme here. The realistic mantle structure and tectonic regime in these models largely affects the surface stress field, and two cratonic properties which are contingent on this are the viscosity and yield strength of the mantle roots.

The effect of highly viscous roots on cratonic stress and deformation is shown in Fig. 3. In this example, the cratonic roots have the same temperature-dependence as the convecting mantle, but the actual viscosity at a given temperature is shifted by a constant factor (the viscosity contrast). For no viscosity contrast (i.e. the same viscosity function for both the oceanic lithosphere and the cratonic lithosphere), the cratons undergo significant deformation. At the end of the simulation (approximately 1 Gyr) the continent has exhibited 60% strain. For higher viscosity contrasts, the continent becomes progressively less susceptible, and the resultant strains are negligible. Though never zero, we consider the strain negligible at a nominal value of $<5\%$. For this value, the continent can be considered a “craton” exhibiting minimal tectonic strain, for viscosity contrasts of 50 or more. This is well below the theoretical maximum contrast of ~ 150 expected for dry cratonic roots versus wet oceanic asthenosphere, and root stability is expected for a plausible range of viscosity contrasts between the root and asthenospheric mantle. Interestingly, the stress in the root increases for higher viscosity contrasts — higher viscosity roots can sustain more stress before deforming. However, at such high stresses, the roots may yield and fail in a plastic manner, as noted by Lenardic et al. (2003).

Fig. 3c shows the effect of increasing the relative yield strength of the roots ith respect to the mantle. The viscosity functions of the cratonic root and ocean lithosphere are the same. For high relative cratonic strength, the cratons are capable of supporting higher stresses, however, the magnitude of stress increase (a few megapascals) is quite subdued. The reason, in this case, is that for non-existent viscosity contrast between the roots and asthenosphere, the roots are not capable of supporting significant stress before viscously deforming. For higher viscosity contrasts, roots with high relative yield strength are capable of supporting much higher stress increases (Lenardic et al., 2003). However, Fig. 3c highlights the point that without the viscosity increase, the effect of higher yield strength ratios is of second order importance for cratonic survival.

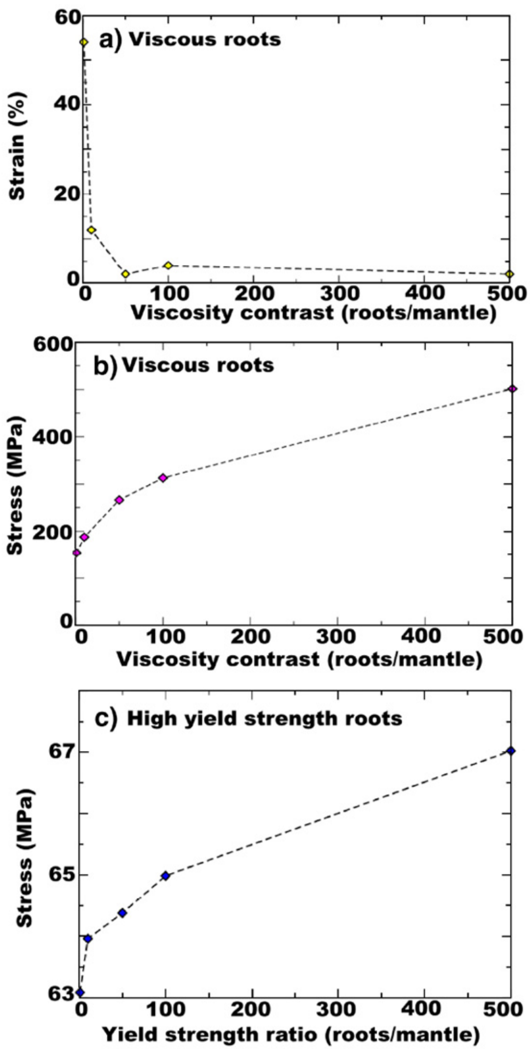

Fig. 3. Results from our simulations over 1 Gyr evolution, for different cratonic lithosphere properties. a) Final strain versus the viscosity contrast between the cratonic roots and the bulk mantle. Both cratons and the mantle are assumed to have the same temperature dependence, but the cratonic viscosities are multiplied by the viscosity contrast. For viscosity contrasts > 50, the strains are negligible are the cratonic lithosphere is stable. b) Average cratonic stress for the same models in (a). More viscous roots are able to support greater stresses without significant deformation. c) Cratonic stress versus the relative yield strength of the roots and mantle. These models assume no viscosity contrast between the roots and mantle. Stronger roots are able to support more stress without yielding, but the magnitude of stress increase (~4 MPa) is minimal.

The two factors, however, are not unrelated. Dry, depleted cratonic mantle lithosphere not only has a higher intrinsic viscosity, but is capable of supporting up to 800 MPa before yielding (cw. Kohlstedt et al., 1995). In contrast, the presence of minor water in oceanic asthenosphere not only lowers its viscosity, but affects the yield strength of the oceanic lithosphere. Pore water pressure in sub-oceanic faults can reduce their strength by an order of magnitude cw dry faults (Kohlstedt et al., 1995), and the localized serpentinization of such faults may further reduce their strength by a factor of 2–3 (Escartin et al., 1997). For a plausible range of viscosity contrasts (50–150) and yield strength ratios (5–30), cratonic stability is expected under present day mantle conditions.

3.2. Mantle properties

These models specifically address the question of cratonic survival in a realistic convecting mantle, and the two important facets to this realism compared with earlier models are the inclusion of the 670 km phase transition (Christensen, 1995), and a depth-dependent mantle viscosity structure (e.g. Richards et al., 2001).

The spinel → magnesiowustite + perovskite at ~670 km is an endothermic transition, and it has been shown to hinder mass transfer across this internal boundary. Earlier models (Christensen and Yuen, 1985) were able to achieve perfect layering between the upper

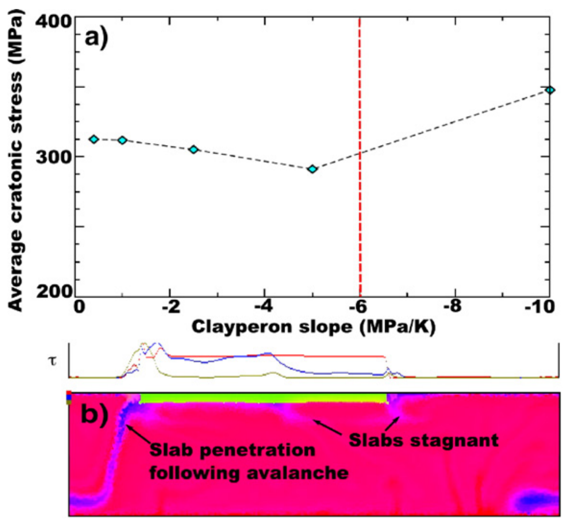

Fig. 4. Variation of the average cratonic stress with the Clayperon slope of the spinel → perovskite + magnesiowustite phase transition at c. 670 km. The theoretical maximum slope for before isoviscous convection becomes layered (Christensen and Yuen, 1985) is shown as a vertical line. Slab breakthrough into the lower mantle occurs for all values of the Clayperon slope explored here, due to the penetrative ability of strong, viscous slabs. b) A temperature snapshot of our nominal model, here with a Clayperon slope of −6 MPa/K. Time dependent slab-breakthrough into the lower mantle is observed even for this example, despite the unrealistically steep Clayperon slope. For certain geometric configurations, though, such as shallow-dipped subduction with trench rollback, slabs may stagnate for a period at the 670 km discontinuity before breaking through (shown).

and lower mantle when the Clayperon slope of the transition were less than −6 MPa/K. This is far lower than typically accepted values (−2.5 MPa/K, see Schubert et al., 2001, and references therein), and the Clayperon slope for this reaction may be even less. For such mild values, the resistance to slab breakthrough is not effective, and whole mantle convection is expected. The difficulties in obtaining mantle layering are exacerbated in models with strong, rigid plates, as is the case here, and we find significant, albeit time-dependent, mass transfer for all Clayperon slopes explored (Fig. 4a). Furthermore, since there is no systematic variation in the tectonics of the system, varying the Clayperon slope has no systematic effect on the average cratonic stress.

Fig. 5 illustrates the effect of a realistic mantle structure on cratonic stress. Two different variables are explored in the figure; an increase in lower mantle viscosities (with respect to the temperature dependent viscosity law), and the increase in upper mantle (asthenospheric) viscosities. When the viscosity of the lower mantle is increased, the convective vigour of the lower mantle decreases (i.e. lower lower mantle Rayleigh numbers), and thus convection switches from small-scale plumes to broad-scale upwellings. High viscosity lower mantles are capable of generating greater stresses, however, their flow velocities are diminished. Thus we see two effects in Fig. 5a. The average cratonic stress increases with increasing lower mantle viscosity, since the lower mantle is capable of generating large convective stresses. However, the maximum stress experienced by the cratons is diminished, since the severity of avalanche events is subdued by the lower speeds at which slabs can move in the highly viscous lower mantle.

With decreasing asthenospheric viscosities, the average cratonic stress is largely unaffected. However, the maximum stress experience by the cratons decreases with decreasing asthenospheric viscosities, since during slab breakthrough events, the viscous coupling of the "slab-suction" force associated with the subducting slab is mitigated by the low viscosity asthenosphere, and so buffers the cratons from the tectonic extremes experienced during mantle avalanche events.

To summarise, realistic mantle structures have a minimal effect on the average, background cratonic stress field. They do, however, have an important buffering effect during the tectonic stress extremes induced by mantle avalanche effects. High-viscosity lower mantle hinders the sinking speed of slabs, modulating stress excursions at the surface. Low-viscosity asthenosphere decouples surface plates from "slab-suction" during the penetration of slabs into the lower mantle. Both factors work to cushion cratons from the tectonic extremes they would otherwise be subject to during a mantle avalanche event.

3.3. Cratons in an evolving mantle

As previously noted cratons have survived for billions of years not under present-day mantle condition, but in a mantle that has evolved significantly in that

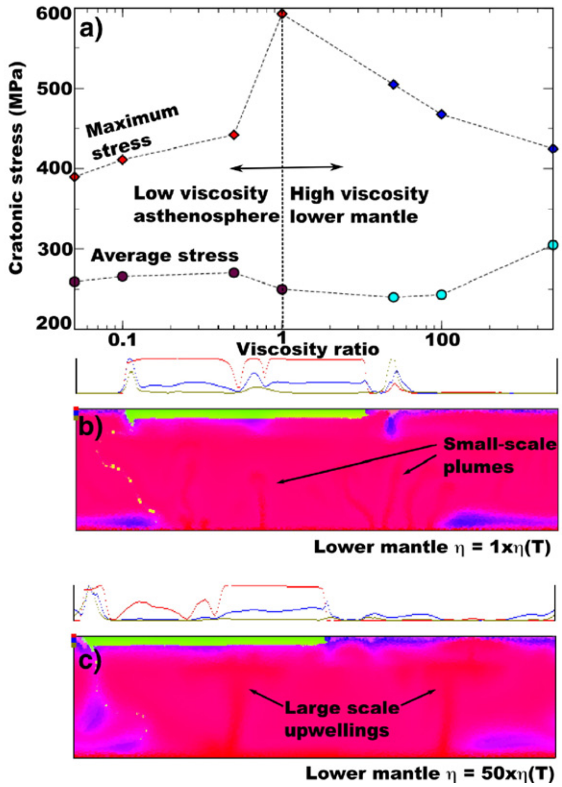

Fig. 5. The effect of variations in the viscosity structure of the mantle in average and maximum cratonic stress. Two sets of experiments are shown in this plot; the effect of varying solely the asthenospheric viscosity (i.e. for depths les than 640 km, and temperatures greater than 1300 °C), shown to the left of the vertical dashed line; and the effect of increasing the viscosity of the lower mantle (to the right of the vertical line). In both cases, the viscosity ratio is multiplied by the pre-existing default viscosity from the temperature-dependent law. Decreasing the asthenospheric viscosity has the effect of decoupling the plates somewhat from mantle dynamics, which, while not grossly affecting the average cratonic stress, does protect cratons from large stress increases during mantle avalanche events. Increasing the lower mantle viscosity has the same effect for a different reason, it slows slabs penetrating the 670 km discontinuity. High lower mantle viscosities also greatly affect the planform of convection. Shown in b) and c) are two examples of our nominal model, with either no additional increase in lower mantle viscosity (b) — resulting in numerous small scale plumes, or a 50-fold increase in the lower mantle viscosity from its nominal value (c) — resulting in the development of very broad, large-scale upwellings.

time. This includes not just thermal state – the average mantle heat flux was 3–4 times greater in the Archean – but also in tectonic regime. Higher mantle heat production results in a greater degree of convective vigour and generally higher mantle flow velocities (Turcotte and Schubert, 1982). It has also been suggested that a greater degree of mantle layering would be anticipated for higher mantle temperatures in the past due to the endothermic nature of the 670 km phase transition (Christensen, 1995). However, hotter mantle temperatures also reduce the internal viscosity of the mantle, due to its extreme temperature dependence (Karato and Wu, 1993). This has the potential to lower the viscous coupling between the cratons and induced mantle stresses, effectively lowering the stress state of the cratons.

It is this latter effect which is dominant, as seen in Fig. 6. For increasing heat production, hotter internal mantle temperatures result in lower mantle viscosities, moderating the viscously induced stresses felt by the cratons. The net result is that cratons experience lower average lithospheric stresses in Archean mantle, despite an overall increase in mantle velocities. We do not see a systematic increase in the degree of layering or the timing of mantle breakthrough events in our models. Mantle breakthrough events, in these models, are extremely time-dependent, and no discernible change in the frequency is noted at higher mantle heat production values. Mass transfer between the upper and lower mantle remains effective for the range of mantle heat production explored here, and slab penetration is expected to occur without difficultly in Archean mantle conditions. The primary effect of higher antle temperatures is, unexpectedly, to enhance craton stability for much of the Precambrian. This is at odds with previous studies (Lenardic and Moresi, 1999), and the reason for the difference is primarily the inclusion of strongly temperature-dependent viscosities in these simulations, which constitute the greatest effect on the cratonic stress field.

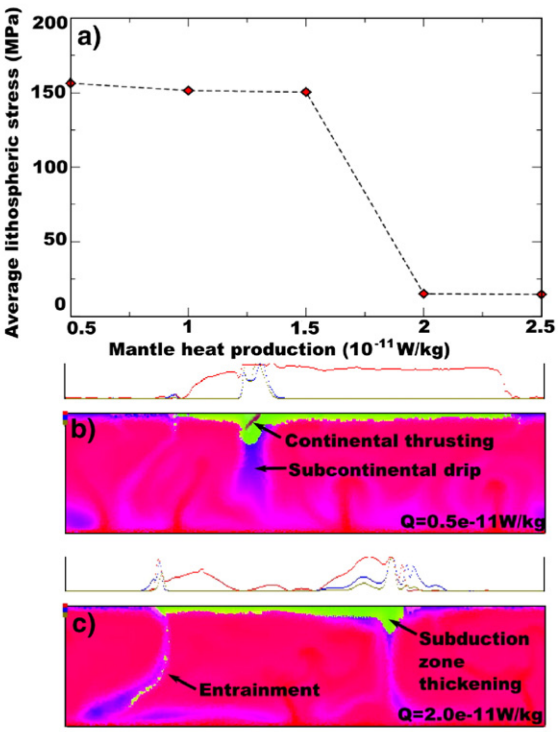

Fig. 6. a) Average cratonic stress versus increasing mantle heat production, from present-day conditions (~0.5e−11 W/kg) up to Archean mantle conditions (heat production 3–4 times greater than today). Increasing mantle temperatures results in a drastic drop in the average lithospheric stress. Hotter mantle conditions in the past result in lower mantle viscosities, decreasing the convectively induced cratonic stress. b) Temperature snapshot of simulation with "present-day" heat production and other nominal model parameters (Table 1), but with continental lithospheric viscosities a factor of 10 greater than the mantle. The continent deforms in response to the ambient mantle stresses, and mantle downwellings penetrate the 670 km discontinuity efficiently. c) Mantle convection under Archean mantle conditions (i.e. internal heat production Q of 2e−11 W/kg), all other factors similar to (b). Slab penetration into the lower mantle still occurs in this case despite, on average, higher mantle temperatures. The dominant difference between (b) and (c) is the lower mantle viscosities in the latter, resulting in much lower (around a factor of 4) average cratonic stress.

4. Discussion and model limitations

The main results of this study are: 1) within reasonable estimates of the viscosity contrast between cratonic roots and the asthenosphere (50–150) and the yield strength ratio between oceanic faults and cratonic mantle lithosphere (5–30), cratonic stability is expected under present day mantle conditions. 2) The surface tectonics of the Earth is time-dependent, and one of the sources of this time-dependence is the spinel → perovskite + magnesio wustite phase transition at ~670 km. Mantle avalanches through this transition can cause a 2–3 fold increase in cratonic stress. 3) Realistic mantle viscosity layering, consistent with post-glacial rebound and geoid studies (King, 1995), with a high viscosity lower mantle, and low-viscosity asthenosphere, tend to mitigate the stress excursion felt by cratons during mantle avalanche events. 4) The dominant effect of an evolving mantle is changing mantle viscosities; lower mantle viscosities for much of the Precambrian would tend to promote cratonic stability. Thus cratonic longevity is a function of the physical properties of the cratonic lithosphere, and their dynamics in a realistically layered evolving mantle.

This study has advanced on previous work by the inclusion of an additional layer of realism in the simulation of Earth-like convection. But what are the limitations of this work and in what way do they affect our primary results? From the point of view of mantle dynamics, there are three obvious limitations: the limited modelling domain, the fact the models are two-dimensional, and the restricted time (<~2 Gyr) of individual simulations. While the results of these simulations have been scaled to Earth values, the use of limited-size 4×1 boxes is an obvious restriction. However, the dynamics of system are affected little by the choice of geometry, and our fundamental results are invariant over a range of aspect ratios explored. The use of two-dimensional geometries naturally restricts us from commenting on global averages or energy budgets, and many important Earth-system dynamics (toroidal motion of plates, oblique subduction etc) are inherently three-dimensional effects. The complexities of three-dimensional effects are beyond the scope of this paper, the point of which was to explore the effects of a broad range of physical parameters — an exercise that is extremely time-consuming in three-dimensions. Finally, these systems are very time-dependent, and for similar parameter ranges the simulations can exhibit different behaviour if started from a different initial state. We have run all our models to a statistical steady state in order to calculate time-averaged values such as average cratonic stress, and the effect of this is to average-out the individual histories of each simulation into a globally resolvable parameter. Furthermore, it should be noted that the average continental stress field at depth is controversial, and the exact values noted here dependent on rheological factors such as viscosity. The absolute values presented are reasonable for highly viscous cratonic roots, but the important thing here is the relative changes due to the variation in different parameters.

One critical aspect of cratonic destruction which we do not address here is the effect of heterogeneities. By this we mean either lateral strength heterogeneities, such as strong cratons embedded in weaker mobile zones which effectively act as a buffer-zone (Lenardic et al., 2000), or time-dependent heterogeneities such as weakening by metasomatism. The former concept was explored by Lenardic et al. (2000), who noted that in this case it is not solely up to the physical properties of the cratons to resist deformation, but rather cratons can avoid deformation by virtue of their positioning, for example. We have not explored this concept further in this work, and the cratons in our simulations by definition survive solely by virtue of their physical properties. It should be noted, however, that the "crumple-zone" model has a parallel in the viscous decoupling of cratons from the mantle by a low-viscosity asthenosphere — this is, essentially, a vertical "crumple-zone" where the low-viscosity asthenosphere absorbs high strain rates that otherwise could propagate into cratonic regions.

A critical aspect of many documented episodes of cratonic destructions appears to be re-fertilization of the depleted lithosphere prior to the destructive tectonic event. The extent and degree of re-fertilization of cratonic lithosphere is not well-known (Griffin et al., 2003), but in some cases is extensive (Beyer et al., 2006). For the purposes of our models, refertilization results in a decrease of the viscosity contrast of the craton and asthenosphere, and/or the reduction of the relative yield strength of the cratonic mantle lithosphere. Both factors conspire to make cratons susceptible to further tectonic disturbance. Quantitatively, our results show that if this process reduces cratonic root viscosities to less than 50 times that of the asthenosphere, then de-cratonization should result. The processes and mechanisms of such re-fertilization are not well understood, however, and a better understanding of the geodynamic circumstances under which re-fertilization occurs is required before such mechanisms can be simulated in mantle convection models.

5. Conclusions

The longevity of cratons, when compared to rapidly recycled oceanic lithosphere or tectonically active continental terranes, is an important factor in the geodynamic evolution of Earth's surface. Under present-day mantle conditions, and using realistic mantle structures, craton stability is expected if the viscosity ratio between the cratonic lithosphere and the asthenosphere is 50–150, and if the yield-strength ratio between cratonic mantle lithosphere and hydrated oceanic faults exceeds 5–30, well within plausible bounds. Realistic mantle viscosity structures tend to mitigate extreme stress excursions within cratonic roots. The inclusion of more complex mantle physical behaviour, such as phase transitions, tend to introduce a time-dependence to convection, resulting in high stress events such as mantle avalanches, which have been suggested to have been dominant in the previous models for the Precambrian mantle. We find, however, that the dominant effect of a hotter Precambrian mantle is lower mantle viscosities, which result in lower cratonic stresses, promoting cratonic survival in the past. Craton longevity in an Earth-like style of mantle convection is not at all paradoxical for plausible cratonic properties, and realistic mantle structures.

Acknowledgements

This work is supported by a Macquarie University Research Fellowship (CO'N), and an ARC Discovery Grant (WLG & SYO'R). This is GEMOC ARC National Key Centre publication no. 479.

References

Ball, P., Ebinger, C., McClay, K., Brown, B., Totterdell, J., 2003. Constraints on continental break-up within the Great Australian Bight from gravity and magnetic data analyses. EGS – AGU – EUG Joint Assembly 5237.

Beyer, E.E., Griffin, W.L., O'Reilly, S.Y., 2006. Transformation of Archaean lithospheric mantle by refertilization: evidence from exposed peridotites in the Western Gneiss Region, Norway. J. Petrol. 47 (8), 1611–1636. doi:10.1093/petrology/eg1022.

Boyd, F.R., Gurney, J.J., Richardson, S.H., 1985. Evidence for a 150–200 km thick Archaean lithosphere from diamond inclusion thermobarometry. Nature 315, 387–389.

Carlson, R.W., Boyd, F.R., Shirey, S.B., Janney, P.E., 2000. Continental growth, preservation and modification in Southern Africa. GSA Today 10, 1–7.

Christensen, U., 1995. Effects of phase transitions on mantle convection. Annu. Rev. Earth Planet. Sci. 23, 65–87.

Christensen, U., Yuen, D.A., 1985. Layered convection induced by phase transitions. J. Geophys. Res. 90, 10291–10300.

Conrad, C.P., Lithgow-Bertelloni, C., 2002. How mantle slabs drive plate tectonics. Science 298, 207–209.

Cooper, C.M., Lenardic, A., Levander, A., Moresi, L., 2006. Creation and preservation of cratonic lithosphere: seismic constraints and geodynamic models. In: Benn, Keith, Mareschal, Jean- Claude, Condie, Kent C. (Eds.), American Geophysical Union. Archean Geodynamics and Environments, Geophysical Monograph Series, vol. 164, pp. 75–88.

Davies, G.F., 1995. Penetration of plates and plumes through the mantle transition zone. Earth Planet. Sci. Lett. 133, 507–516.

Deen, T.J., Griffin, W.L., Begg, G., O'Reilly, S.Y., Natapov, L., Hronsky, J., 2006. Thermal and compositional structure of the subcontinental lithospheric mantle: derivation from shear wave seismic tomography. Geochem. Geophys. Geosyst. 7 (7). doi:10.1029/2005GC001120.

Doin, M.-P., Fleitout, L., Christensen, U., 1997. Mantle convection and the stability of depleted and undepleted continental lithosphere. J. Geophys. Res. 102, 2771–2787.

Eggler, D.H., Meen, J.K., Welt, F., Dudas, F.O., Furlong, K.P., McCallum, M.E., Carlson, R.W., 1988. Tectonomagmatism of the Wyoming Province. Colorado School of Mines Quarterly, vol. 83, pp. 25–40.

Escartin, J., Hirth, G., Evans, B., 1997. Effects of serpentinization on the lithospheric strength and the style of normal faulting at slow-spreading ridges. Earth Planet. Sci. Lett. 151, 181–189.

Griffin, W.L., Kaminsky, F.V., Ryan, C.G., 1998a. Thermal state and composition of the lithospheric mantle beneath the Daldyn kimberlite field, Yakutia. Tectonophysics 262, 19–33.

Griffin, W.L., Zhang, A., O'Reilly, S.Y., Ryan, C.G., 1998b. Phanerozoic evolution of the lithosphere beneath the Sino–Korean craton. Mantle Dyanmics and Plate Interactions in East Asia. Geodynamic Series, vol. 27, pp. 107–126.

Griffin, W.L., Doyle, B.J., Ryan, C.G., Pearson, N.J., Davies, R., Kivi, K., Achterberg, E.V., Natapov, L.M., 1999. Layered mantle lithosphere in the Lac de Gras Area, Slave craton: composition, structure and origin. J. Petrol. 40, 705–727.

Griffin, W.L., O'Reilly, S.Y., Abe, N., Aulbach, S., Davies, R.M., Pearson, N.J., Doyle, B.J., Kivi, K., 2003. The origin and evolution of Archean lithospheric mantle. Precambrian Res. 127, 19–41.

Griffin, W.L., Graham, S., O'Reilly, S.Y., Pearson, N.J., 2004a. Lithosphere evolution beneath the Kaapvaal Craton. Re–Os systematics of sulfides in mantle-derived peridotites. Chem. Geol. 208, 89–118.

Griffin, W.L., O'Reilly, S.Y., Doyle, B.J., Pearson, N.J., Cooper-smith, H., Kivi, K., Malkovets, N., Pokhilenko, N., 2004b. Lithospheric mapping beneath the North American plate. Lithos 77, 873–922.

Gung, Y., Panning, M., Romanowicz, B., 2003. Global anisotropy and the thickness of continents. Nature 422, 707–711.

Handley, LM., Wingate, MTD., 2000. SHRIMP zircon age for an Early Cambrian dolerite dyke: an intrusive phase of the Antrim Plateau Volcanics of northern Australia. Aust. J. Earth Sci. 47 (6), 1029–1040.

Helmstaedt, H.H., Gurney, J.J., 1995. Geotectonic controls of primary diamond deposits: implications for area selection. J. Geochem. Explor. 53, 125–144.

Hirth, G., Kohlstedt, D.L., 1996. Water in the oceanic upper mantle: implications for rheology, melt extraction and the evolution of the lithosphere. Earth Planet. Sci. Lett. 144, 93–108.

Humphreys, E.D., Hessler, E., Dueker, K.G., Farmer, G.L., Erslev, E.A., Atwater, T.A., 2003. How Laramide-age hydration of North American lithosphere by the Farallon slab controlled subsequent activity in the western United States. Int. Geol. Rev. 45, 575–595.

James, D.E., Fouch, M.J., 2002. Formation and evolution of Archean cratons: insights from southern Africa. In: Flower, C.M.R., Ebinger, C.J., Hawkesworth, C.J. (Eds.), The Early Earth: Physical, Chemical and Biological Development, Special Publication—Geological Society of London, vol. 199. Geological Society of London, London, pp. 1–26.

Jordan, T.H., 1978. Composition and development of the continental tectosphere. Nature 274, 544–548.

Karato, S., Wu, P., 1993. Rheology of the upper mantle — a synthesis. Science 260, 771–778.

King, S.D., 1995. The viscosity structure of the mantle. Rev. Geophys. 33, 11–17.

Kohlstedt, D.L., Evans, B., Mackwell, S.J., 1995. Strength of the lithosphere: constraints imposed by laboratory experiments. J. Geophys. Res. 100, 17,587–17,602.

Lee, C.T., 2006. Geochemical/Petrological constraints on the origin of cratonic mantle. In: Benn, Keith, Mareschal, Jean-Claude, Condie, Kent C. (Eds.), Archean Geodynamics and Environments. Geophysical Monograph Series, vol. 164. American Geophysical Union, pp. 89–114.

Lenardic, A., Moresi, L., 1999. Some thoughts on the stability of cratonic lithosphere: effects of buoyancy and viscosity. J. Geophys. Res. 104, 12747–12758.

Lenardic, A., Moresi, L., Muhlhaus, H., 2000. The role of mobile belts for the longevity of deep cratonic lithosphere: the crumple zone model. Geophys. Res. Lett. 27 (8), 1235–1238.

Lenardic, A., Moresi, L., Muhlhaus, H.-B., 2003. Longevity and stability of cratonic lithosphere: insights from numerical simulations of coupled mantle convection and continental tectonics. J. Geophys. Res. 108. doi:10.1029/2002JB001859.

Lenardic, A., Richards, M.A., Busse, F.H., 2006. Depth-dependent rheology and the horizontal length scale of mantle convection. Geophys. res. 111. doi:10.1029/2005JB003639.

Maggi, A., Jackson, J.A., McKenzie, D., Priestley, K., 2000. Earthquake focal depths, effective elastic thickness, and the strength of the continental lithosphere. Geology 28, 495–498.

Milashev, V.A., 1974. The kimberlite provinces, Nedra, Leningrad, pp. 548–574.

Moresi, L., Solomatov, V., 1998. Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of Earth and Venus. Geophys. J. Int. 133, 669–682.

Moresi, L., Dufour, F., Muhlhaus, H.-B., 2003. A Langrangian integration point finite element method for large deformation modeling of viscoelastic geomaterials. J. Comput. Acoust. 184, 476–497.

Myrow, P.M., Hughes, N.C., Paulsen, T.S., Williams, I.S., Parcha, S.K., Thompson, K.R., Bowring, S.A., Peng, S-C., Ahluwalia, A.D., 2003. Integrated tectonostratigraphic analysis of the Himalaya and implications for its tectonic reconstruction. Earth Planet. Sci. Lett. 212, 433–441.

Niu, F., Levander, A., Cooper, C.M., Lee, C.-T.A., Lenardic, A., James, D.E., 2004. Constraints on the depth and composition of the mantle keel beneath the Kaapvaal craton. Earth Planet. Sci. Lett. 224, 337–346.

O'Neill, C., Moresi, L., 2003. How long can diamonds remain stable in the continental lithosphere? Earth Planet Sci. Lett. 213, 43–52.

O'Reilly, S.Y., Griffin, W.L., Poudjom Djomani, Y.H., Morgan, P., 2001. Are lithospheres forever? GSA Today April, 4-10.

Polet, J., Anderson, D.L., 1995. Depth extent of cratons as observed from tomographic studies. Geology 23, 205–208.

Pollack, H.N., 1986. Cratonization and thermal evolution of the mantle. Earth Planet. Sci. Lett. 80, 175–182.

Poudjom Djomani, Y.H., O'Reilly, S.Y., Griffin, W.L., Morgan, P., 2001. The density structure of subcontinental lithosphere through time. Earth Planet. Sci. Lett. 184, 605–621.

Poudjom Djomani, Y.H., Griffin, W.L., O'Reilly, S.Y., Doyle, B.J., 2005. Lithospheric domains and controls on kimberlite emplacement, Slave Province, Canada: evidence from elastic thickness and upper mantle composition. Geochem. Geophys. Geosyst. 6 (10). doi:10.1029/2005GC000978.

Priestley, K., McKenzie, D., Debayle, E., 2006. The state of the upper mantle beneath southern Africa. Tectonophysics 416, 101–112.

Puchkov, V., 2002. Paleozoic evolution of the east European continental margin involved in the Uralide orogeny. Mountain building in the Uralides: Pangea to the present. Geophysical Monograph, vol. 132. American Geophysical Union, pp. 9–31.

Richards, M.A., Yang, W.-S., Bunge, H.-P., 2001. Role of a low-viscosity zone in stabilizing plate tectonics: implications for comparative terrestrial planetology. Geochem. Geophys Geosyst. 2 (8). doi:10.1029/2000GC000115.

Richardson, S.H., Gurney, J.J., Erlank, A.J., Harris, J.W., 1984. Origin of diamonds in old enriched mantle. Nature 310, 198–202.

Sandiford, M., McLaren, S., 2002. Tectonic feedback and the ordering of heat producing elements within the continental lithosphere. Earth and Planet. Sci. Lett. 204, 133–150.

Schmeling, H., Bussod, G.Y., 1996. Variable viscosity convection and partial melting in the continental asthenosphere. J. Geophys. res. 101, 5411–5424.

Schubert, G., Turcotte, D., Olson, P., 2001. Mantle Convection in the Earth and Planets. Cambridge University Press, Cambridge, UK, p. 940.

Sengor, A.M.C., 1999. Continental interiors and cratons: any relation? Tectonophysics 305, 1–42.

Shapiro, S.S., Hager, B.H., Jordan, T.H., 1999. Stability and dynamics of the continental tectosphere. Lithos 48, 135–152.

Shirey, S.B., Harris, J.W., Richardson, S.H., Fouch, M.J., James, D.E., Cartigny, P., Deines, P., Viljoen, F., 2002. Diamonds genesis, seismic structure, and evolution of the Kaapvaal–Zimbabwe craton. Science 297, 1683–1686.

Spetsius, Z.V., Belousova, E.A., Griffin, W.L., O'Reilly, S.Y., Pearson, N.J., 2002. Archean sulfide inclusions in Paleozoic zircon megacrysts from the Mir kimberlite, Yakutia: implications for the dating of diamonds. Earth Planet. Sci. Lett. 199, 111–126.

Stein, M., Hofmann, A.W., 1994. Mantle plumes and episodic crustal growth. Nature 372, 63–68.

Stern, R.J., 2004. Arc assembly and continental collision in the Noeproterozoic east African orogen. Annu. Rev. Earth Planet. Sci. 22, 319–351.

Turcotte, D.L., Schubert, G., 1982. Geodynamics: Applications of Continuum Physics to Geological Problems. Wiley, New York, p. 450.

van der Lee, S., Nolet, G., 1997. Upper mantle dS velocity structure of North America. J. Geophys. Res. 102 (B10), 22815–22838.

Westerlund, K., Shirey, S.B., Richardson, S.H., Gurney, J.J., Harris, J.W., 2003. Re–Os isotope systematics of peridotitc diamond inclusion sulphides from the Panda kimberlite, Slave Craton. Extended Abstract, Eight International Kimberlite Conference, Victoria.

Xu, Y.-G., 2001. Thermo-tectonic destruction of the Archaean lithospheric keel beneath the Sino-Korean craton in china: evidence, timing and mechanism. Phys. Chem. Earth, Part A Solid Earth Geod. 26 (9–10), 747–757.