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Formation of cratonic lithosphere during the initiation of plate tectonics

A.P. Beall[1*], L. Moresi[1], and C.M. Cooper[2]

[1]*School of Earth Sciences, University of Melbourne, Parkville, Victoria 3010, Australia*

[2]*School of the Environment, Washington State University, PO Box 642812, Pullman, Washington 99164-2812, USA*

**Current address: School of Earth & Ocean Sciences, Cardiff University, Cardiff CF10 3AT, UK*

ABSTRACT

The Earth's oldest near-surface material, the cratonic crust, is typically underlain by thick lithosphere (>200 km) of Archean age. This cratonic lithosphere likely thickened in a high compressional stress environment, potentially linked to the onset of crustal shortening in the Neoarchean. Mantle convection in the hotter Archean Earth would have imparted relatively low stresses on the lithosphere, whether or not plate tectonics was operating, so a high stress signal from the early Earth is paradoxical. We propose that a rapid transition, from heat-pipe mode convection to the onset of plate tectonics, generated the high stresses required to thicken the cratonic lithosphere. Numerical calculations are used to demonstrate that an existing buoyant and strong layer, representing depleted continental lithosphere, can thicken and stabilize during a lid-breaking event. The peak compressional stress experienced by the lithosphere is 3–4x higher than for the stagnant lid or mobile lid regimes immediately before and after. It is plausible that the cratonic lithosphere has not been subjected to this high stress-state since, explaining its long-term

stability. The lid-breaking thickening event reproduces features observed in typical Neoarchean cratons, such as lithospheric seismological reflectors and the formation of thrust faults. Paleoarchean 'pre-tectonic' structures can also survive the lid-breaking event, acting as strong rafts, that are assembled during the compressive event. Together, the results indicate that the signature of a catastrophic switch, from a stagnant lid Earth to the initiation of plate tectonics, has been captured and preserved in the characteristics of cratonic crust and lithosphere.

INTRODUCTION

The surviving remnants of the Archean crust appear to have been formed under conditions of a low geothermal gradient (Burke and Kidd, 1978) best explained by a pre-plate tectonic, stagnant, heat-pipe mode of mantle convection characterized by vertical tectonics and low convective stresses (Moore and Webb, 2013; Rozel et al., 2017). This crust is underlain by the thickened remnants of lithosphere of similar age (Pearson et al., 1995; O'Reilly et al., 2001) that form the stable, buoyant continental tectosphere (Jordan, 1975). Long term stability of the cratonic lithosphere requires high strength in addition to buoyancy to survive erosion by mantle convection (Lenardic and Moresi, 1999). We propose that cratons formed above regions of lithospheric foundering, predicted to have occurred when the Earth switched from the heat-pipe mode to a mobile-lid, plate tectonic regime (Moresi and Solomatov, 1998). We argue that stresses of this magnitude have not been reproduced since the Archean, explaining the long-term mechanical stability of the cratons and lack of modern examples.

The initial stability of the cratonic lithosphere requires high compressive stresses (Cooper et al., 2006) to overcome its strength and buoyancy to thicken it to the now

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observed 200–300 km (Pasyanos et al., 2014), which cannot be generated by any steady-state Archean mantle regime. Mobile lid convection involves heat-loss though strong oceanic geothermal gradients, as a result of surface migration and recycling, characteristic of plate tectonics. In its absence, convection operates in the stagnant lid mode (Moresi and Solomatov, 1998), which is dominated by conduction through a globally uniform boundary layer (lid). The heat-pipe regime is a variant of this stagnant lid mode, in which upward melt transport into and through the lid is the dominant heat-loss mechanism and can generate continental crust and highly depleted lithosphere (Moore and Webb, 2013; Rozel et al., 2017), consistent with the observed cratonic composition (O'Reilly et al., 2001). However, this lithosphere would initially have a modest thickness, consistent with estimated melting depths of <100 km (Lee, 2014). An additional thickening mechanism is therefore required.

Cratonic mantle lithosphere thickening appears to have occurred through thrusting analogous to plate tectonic collisional zones (Bostock, 1998), also observed in Neoarchean crustal deformation (Percival et al. 2006). Once plate tectonics begins, the mobile lid regime is able to generate higher stresses than the heat-pipe regime, but subduction zone and convective stresses would still be significantly lower in the warmer Archean mantle compared to today (van Hunen and van den Berg, 2008; Sandu et al., 2011). Cratonic lithosphere formed as a result of thickening by steady-state mobile lid convection is potentially unstable to rising mantle stresses over time in a cooling Earth (Cooper et al. 2007). A transient, high-stress event occurring during the early mobile lid regime is therefore required to thicken and stabilize strong lithosphere that can sustain stability for billions of years.

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The last significant recorded cratonic deformation (Percival et al., 2006; Van Kranendonk et al., 2007) typically occurred at a similar time (within 300 Ma) to the proposed initiation of plate tectonics at ~3 Ga (based on data summarized by Hawkesworth et al., 2017). This switch from the heat-pipe regime to mobile lid (plate tectonics) would momentarily generate extremely high stresses, as the heat-pipe thermal boundary layer was catastrophically recycled into the mantle (Moore and Webb, 2013). This paper examines the manner in which proto-crust and depleted lithosphere thicken and stabilize during lid collapse and the transition to mobile-lid convection, forming cratonic nuclei of considerable strength.

DYNAMIC MODELING

We solve Stokes flow and temperature advection-diffusion using Underworld in a 2D, Cartesian domain 8700 × 2900 km in size (Moresi et al., 2007, see the GSA Data Repository for further details of the numerical modeling methodology). We modeled internally heated whole mantle convection (Fig. 2), with a Rayleigh number (Ra) of 5×10^8, a viscosity range of five orders of magnitude and a resolution of 12 km in the horizontal direction. In the vertical direction, the upper mantle grid spacing is refined to 6km.

Our models begin with a 72.5 km harzburgite layer with a density of 3269 kg/m^{-3} (relative to an assumed mantle density of $3310\ kg/m^3$), consistent with xenoliths (O'Reilly et al., 2001). Our focus is the dependence of craton stability on the deep lithospheric buoyancy and strength. The influence of the continental crust, which could have formed previously through melting and recycling through the heat-pipe lid, is generally ignored. The lithosphere has a finite plastic strength modeled by a depth-dependent Von Mises criterion, representing the increased strength of olivine at high pressure (Karato, 2010) and is assumed to be dry and melt-depleted compared to the mantle which is 8x weaker. The stagnant lid can only be recycled if its strength is limited by a low yield stress of 10-50 MPa (as in Moresi and Solomatov, 1998), which can be generated through dynamic grain-size reduction and a subsequent switch to diffusion creep (Rozel, 2012). A yield strength of 0.6 MPa/km is set, assuming that this weakening has occurred. A short damage-dependent model demonstrates that the lid-breaking event and craton stabilization occur in a similar way without this simplification (data repository).

Melting buffers the mantle temperature, which is approximated by capping the geotherm at the depth-dependent solidus. This produces an intermediate geotherm which lies between the cool, extrusive heat-pipe end member (Moore and Webb, 2013) and the warmer, intrusive end-member (Rozel et al., 2017). When the mantle cools down sufficiently for melting to switch off (following Moore and Webb, 2013), the lithospheric thickness is no longer controlled by the solidus and a large wavelength lithosphere-asthenosphere boundary (LAB) slope can form, typical of the stagnant lid regime (Moresi and Solomatov, 1998). This LAB slope is associated with high stresses which trigger an overturn event, subjecting the harzburgitic layer to an episode of high compressive stress (Figs. 1 and 2). The continental layer is shortened by <80% to form <300 km thick 'cratonic nuclei' during the foundering event.

STRESS HISTORY AND STABILITY OF THE CRATONIC LITHOSPHERE

The lid-breaking event is associated with high horizontal compressional stresses of ~150MPa in the buoyant lithosphere above the zones of mantle downwelling (Fig. 1).

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These maximum stresses are 3–3.5x higher than the heat-pipe and stagnant lid regimes (~30 MPa) prevailing immediately before the initiation of lid-breaking and the mobile lid (plate tectonics) regime that follows.

The lid is recycled (except the continental layer) in fragments initially 1500km - 2000 km wide. Each produces a local stress pulse which lasts 10-50Ma and contributes to the formation of localized shear zones and thrust stacking within the cratonic lithosphere. Thrusting occurs locally until the cratonic lithosphere stabilizes, incrementally assembling a complete craton with up to three pulses over 100 - 350 Ma (central craton in Figure 2B; one pulse has generated a single nucleus by 193 Ma and two more by 609 Ma), predicting in a wide range of craton formation ages globally. Due to the lithospheric strength at high pressure and its cold pre-tectonic geotherm, stabilization occurs as soon as the lithosphere is thickened to a critical depth (as in Cooper et al. 2006). No further cooling or annealing mechanisms are required for stability, which would require longer lived compressional regimes than typical for the hotter mantle.

Once all of the cold, dense lower lid has been recycled, no subsequent high stress pulses occur, as a thick boundary layer can no longer form in the mobile lid (plate tectonics) regime. Models are run for ~500 Ma after the lid-breaking event, in which the modeled craton stress remains below ~30% of the thickening stress. The cratons remain stable within the convecting mobile lid regime (e.g., craton stability after 609 Ma, Fig. 2B), despite an initial period of craton warming and mantle cooling.

The stress state evolution from the modeled Archean mobile lid convection to the modern Earth is estimated by scaling convective stress as a function of mantle temperature, which is estimated to have cooled by ~200 °C (Herzberg et al., 2010).

Mobile lid convective stress varies in proportion to the mantle viscosity (the Frank-Kamenetskii approximation of temperature-dependence is assumed, with $E = 12$) and the convective boundary layer thickness which scales with $Ra^{-\frac{1}{3}}$ (Turcotte and Oxburgh, 1972; Moresi and Solomatov, 1998). This scaling predicts modeled stresses which are 2.5x higher (~75 MPa) in the modern Earth than in the Archean (Fig. 1). The modern cratonic lithospheric stress-state would therefore have only reached ~60%–80% of the stress experienced during its Archean thickening event.

The maximum lithospheric stress during Earth's evolution over the last 3 Ga should also be reflected in the evolution of maximum orogenic crustal thickness. The maximum thickness of orogenic crust is limited to isostatic equilibrium with its local compressional stress state. The largest crustal thickness is proportional to the highest tectonic stress and provides an upper bound for the cratonic lithosphere stress state. Dhuime et al. (2015) calculated the billion year evolution of juvenile upper-plate continental crust thickness at subduction zones, based on geochemical isotopic compilations, which we use as a proxy for tectonic stress. After ~3Ga, stresses increased by 2.2x by ~1.5 Ga, which is initially more rapid than the scaling estimate (Fig. 1). However, the subduction zone stress peaked at this time, such that it is still only ~70% of the stress experienced during craton formation.

The simplified stress evolution is also consistent with the mantle traction stresses and lithospheric stress due to topography (<50 MPa) and net lithospheric stresses near subduction zones (~100 Mpa, Lithgow-Bertolloni and Guynn, 2004). This estimate depends on the lithosphere-mantle stress balance, without which subduction traction could theoretically generate <200 Mpa.

These stress estimates support our hypothesis that the high lithospheric stresses which built the cratons have not been reproduced since their formation. The thermal boundary layer may have thickened, but the cratonic nuclei are predicted to be undeformed unless weakened. The importance of weakening cratons through slab-derived fluid influx is debatable (Lee et al. 2011). The calculated stress evolution indicates that weakening was critical in the past, though subduction zone traction stresses may have risen to magnitudes for which this may be becoming less necessary in anomalously high stress regions.

PRESERVATION OF PRE-PLATE TECTONICS CRUST

The simplified numerical model is an end-member in which the buoyancy and strength of continental crust is negligible. A strong compressional stress-state consistent with both shallow and deep thrusting is generated (Fig. 2 and Cooper et al., 2006). Thrusting is recorded in Neoarchean crustal structures (e.g., Yilgarn and Superior cratons; van der Velden et al., 2006; Percival et al., 2006), but its absence is a key characteristic of Paleoarchean crust. Cratonic nuclei which formed >3 Ga (Pilbara and Kaapvaal cratons) are instead dominated by buoyancy-driven domes (Van Kranendonk et al., 2007), with no evidence of craton thickening. This can be explained by assuming that lid breaking events generated the earliest thrusting and only occurred post-3Ga, which agrees with the predicted onset of tectonics (Hawkesworth et al., 2017) and allows Paleoarchean crust sufficient time to have 'stabilized'. Additional models (Fig. 2d) test this by including an embedded 580 x 36.25 km crustal fragment of Paleoarchean crust, stronger and more buoyant than the continental lithosphere with a plastic strength of 1.5 MPa/km and a density of 2877 kg/m^{-3}.

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The buoyancy of the crustal block, applicable to felsic crust of any age, prevents its deep burial. It is transported onto thick cratonic lithosphere (Fig. 2d) with no internal plastic deformation and relatively modest viscous thickening (which would be further reduced if strain-localization were modeled). Neoarchean crust was forming and therefore still weak enough for the lid-breaking event to generate thrusts (preserved as a synmagmatic combination of horizontal and vertical tectonics in the Yilgarn, Zibra et al., 2014). The structural and geochemical contrasts, for example between Yilgarn and Pilbara crust, may then be the consequence of the large-scale convective regime they formed in, whereas their lithospheric roots formed through similar thickening processes during the initiation of tectonics.

EVIDENCE IN THE GEOLOGICAL RECORD

While our cratonization models are primarily designed to reconcile mantle regime dynamics with the mechanical properties of the cratons, they also reproduce observed features of cratons. Thrusting of the lithosphere and displacement of internal layering on a large scale occurs in the models and can reproduce observed dipping lithospheric reflectors (Bostock, 1998), as well as mid-lithospheric discontinuities (Calò et al., 2016). Rapid Neoarchean events involving the amalgamation of multiple cratonic nuclei (Percival et al., 2006) can also be generated by the lid breaking episode, which provides a mechanism for suturing terranes in a relatively low stress tectonic environment. In the models, individual mantle downwellings generate cratonic nuclei and suture them to form larger assembled cratons. This has the combined effect of transporting crust laterally by <4000 km, suturing terranes which would have contrasting geological evolution due to their initial isolation, as observed in the Superior Province (Percival et al., 2006). Finally, Neoarchean crust preserves horizontal tectonic features which can be explained by the lid-breaking mechanism. The thrust sheets would reflect the compressional stress-state (described earlier). Late-stage felsic magmatism (Percival et al., 2006) could be generated as hydrated mafic crust was rapidly buried to >100 km (Fig. 1, Bédard, 2006), the displacement also assisting with its recycling through the lithosphere.

The strongest evidence for craton formation during a catastrophic mantle transition is the anomalous thickness, strength and billion year stability of the preserved cratonic lithosphere, which places constraints on the evolving mantle dynamics since the Archean. These features are reconcilable with craton formation during the proposed lid-breaking event, which would have generated a pulse of high stress, anomalous compared to lithospheric stresses generated both during the Archean and today. The lid-breaking event involves strong coupling between the asthenosphere, mantle lithosphere and crust, providing opportunities to support our hypothesis with Archean crustal observations. The simplified models demonstrate that our craton formation model is plausible, while there is now the opportunity to more thoroughly explore the role of dynamic rheology and crustal deformation. The lid-breaking model provides a plausible framework for interpreting unique Neoarchean lithospheric and crustal features, which are otherwise difficult to explain purely with secular adjustment to modern-style subduction and plate tectonic models.

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FIGURES

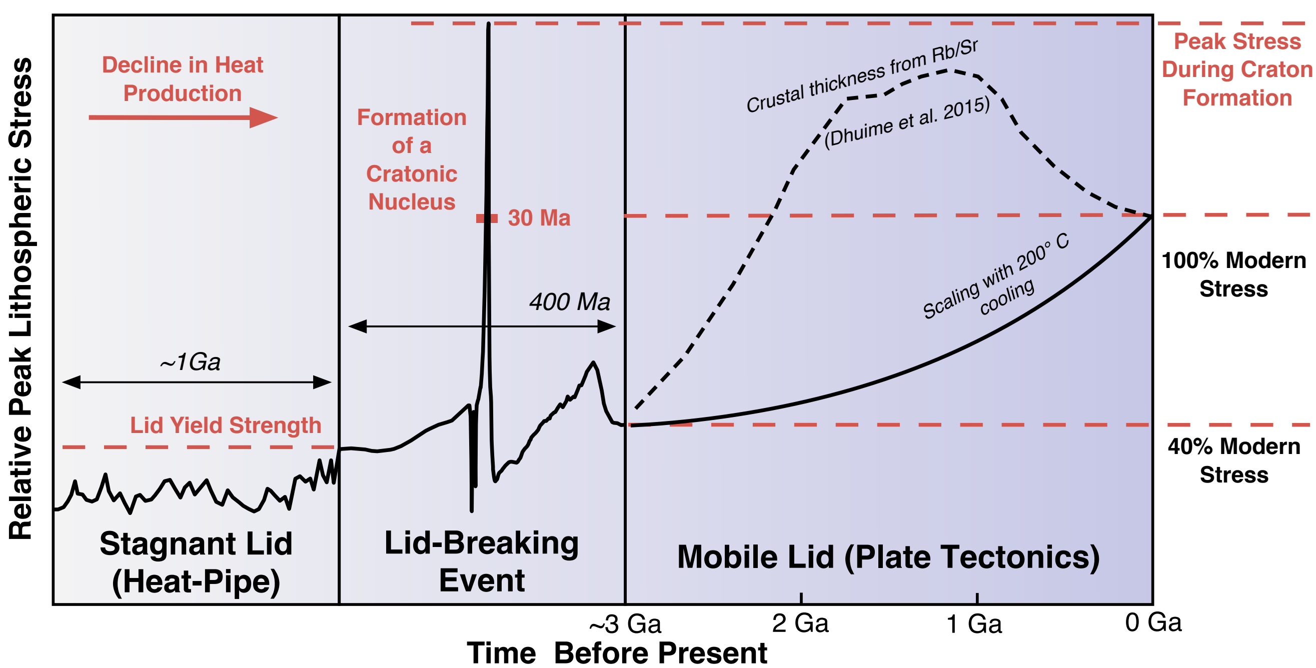

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Figure 1. Earth's evolving lithospheric stress state. The maximum stress during the modeled stagnant lid regime is relatively low and increases gradually due to declining heat-production. When widespread melting switches off, the stagnant lid can develop a boundary layer slope and stresses increase rapidly, reaching the stagnant lid yield stress and triggering the lid-breaking event and the initiation of plate tectonics, at ~3Ga (Hawkesworth et al., 2017). Archean tectonic stresses are likely to be ~60% smaller than today, based on the variation of crustal thickness over time (Dhuime et al. 2015) and mantle convection scaling with $200^{o}C$ of cooling. The modeled lid-breaking event generates pulses of anomalously high stress (ranging <300 Ma after lid-breaking) which form strong cratonic nuclei. These anomalous stresses have not been reproduced since. The lid-breaking event explains cratonic lithosphere formation in an otherwise low-stress environment and the following stability at the billion year time-scale.

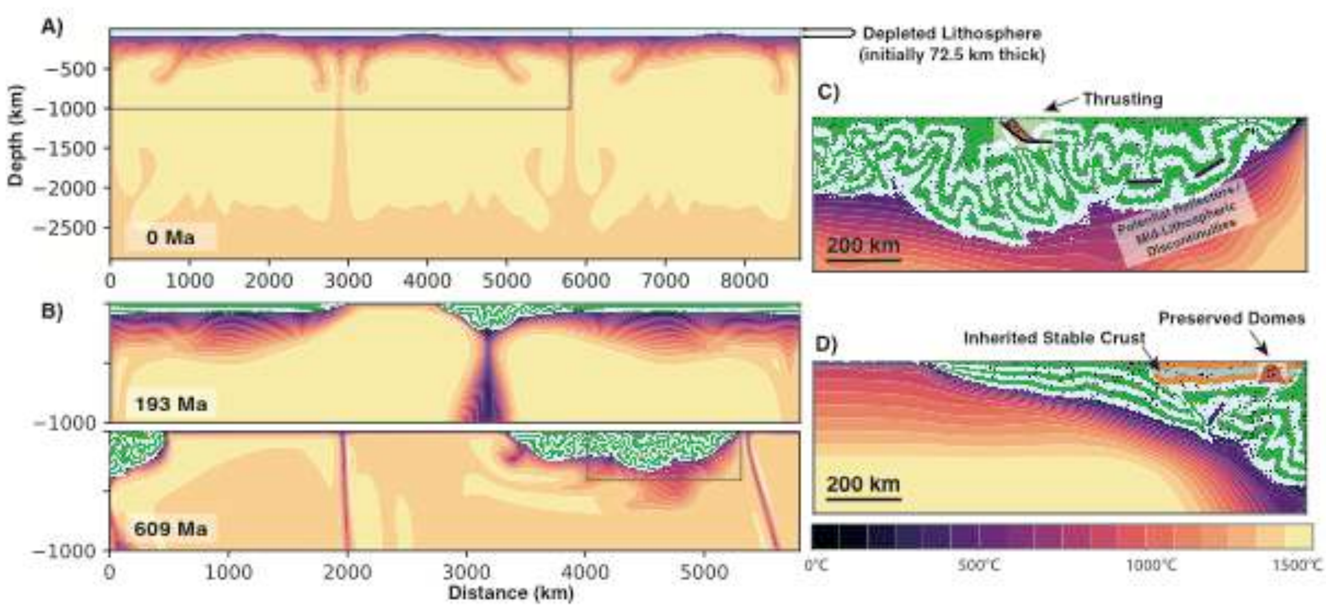

Figure 2. Typical model evolution from the stagnant lid regime (A), to the initiation of the lid-breaking event and stabilization of thick cratonic lithosphere (B, note scale

change). The maximum stress evolution of the left craton is shown in Figure 1 (central panel) and is representative. The data repository contains a movie of this model and the stress evolution for the other two cratons. Individual thickening events take <30 Ma, while multiple cratonic nuclei are amalgamated into the large central craton within 100 Ma. An enlarged frame of the thickened lithosphere (C) demonstrates the thrust structures preserved after thickening. (D) An embedded Palaeoarchean crustal fragment, assumed to have stabilized before the lid-breaking event, records relatively small shortening. This explains the preservation of vertical tectonic structures, whereas significant thrusting occurs in the weaker Neoarchean crust (shown schematically).

1GSA Data Repository item 2018xxx, further details of numerical modeling methodology and movie of typical model evolution, is available online at http://www.geosociety.org/datarepository/2018/ or on request from editing@geosociety.org.