How craton margins are preserved: Insights from geodynamic models

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Abstract. Lateral variations in lithosphere thickness are observed in many continental regions, especially at the boundary between the ancient cratonic core and the adjacent more juvenile lithosphere. In some places, such as the North America craton margin in western Canada and the Sorgenfrei-Tornquist Zone in northern Europe, the transition in lithosphere thickness has a steep gradient ($>45^\circ$) and it appears to be a long-lived feature (at least 50 Ma). We use thermal-mechanical numerical models to address the dynamics of lithospheric thickness changes on timescales of 100 Ma. Models start with the juxtaposition of 60 km thick lithosphere (“mobile belt”) and 160 km thick lithosphere (“craton”). In the reference model, all mantle materials have a damp olivine rheology and a density comparable to primitive mantle. With this configuration, edge-driven mantle convection occurs at the craton boundary, resulting in a lateral smoothing of the thickness transition. The density and rheology of the craton mantle lithosphere are then varied to approximate changes in composition and water content. For all densities, a steep transition is maintained only if the craton strength is 5-50 times stronger than the reference damp olivine. If dry olivine is an upper limit on strength, only cratonic mantle with moderate compositional buoyancy (20-40 kg/m³ less dense than primitive mantle) remains stable. At higher densities, the thick lithosphere is eroded through downwellings, and the craton margin migrates inboard. Conversely, a compositionally buoyant craton destabilises through lateral spreading below the mobile belt.

Keywords: cratons; continental lithosphere; numerical models; gravitational instability; edge-driven convection

42 **Highlights**

- 43 • lateral changes in continental lithosphere thickness induce edge-driven convection
- 44 • thick cratonic lithosphere is readily eroded if it is weak due to partial hydration
- 45 • steep cratonic margins are preserved for at least 100 Ma if the lithosphere is dry
- 46 • moderate compositional buoyancy enhances margin stability

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1. Introduction

Since Dana (1875), it has been recognised that within Earth's continental lithosphere, a number of Archaean and Proterozoic-aged blocks have been preserved for sometimes billions of years, surviving multiple plate tectonic cycles (Figure 1). The Archaean craton fragments in North America and central Eurasia are surrounded by early Proterozoic lithosphere, while in other locations (Africa, South America, western North America, western Europe), Archaean cratons may be juxtaposed against lithosphere that varies in age from Paleozoic to Cenozoic. Archaean and Proterozoic cratons that are surrounded by younger mobile belts are believed to survive plate tectonic cycles because the mobile belts are weaker than the cratonic lithospheric blocks, and the mobile belts are therefore preferred locations for continental rupture during the supercontinent cycle (Audet and Bürgman, 2011). This is not the only inferred role of mobile belts in craton preservation: they may also protect cratons from mantle-derived stresses (Lenardic et al., 2000), shielding them from erosional effects of the mantle wind associated with mantle convection and plate motions. Removal of the mobile belt would then facilitate craton recycling (Lenardic et al., 2000).

While protected by mobile belts, cratons are further shielded from destruction by their strength and buoyancy. Doin et al. (1997) and Wang et al. (2014) find that preservation of cratons is promoted by rheologically strong roots. Cratons are also inferred to possess a lithospheric root with a density lower than the asthenospheric density. The thermal thickness (depth to the 1300°C isotherm) of Archaean and Proterozoic lithosphere varies between about 200 and 350 km (Artemieva and Mooney, 2001), where thicker lithosphere corresponds to older (Archaean) lithosphere. Combining average elevation of Archaean and Proterozoic lithosphere with their thermal gradients, and assuming isostatic equilibrium, suggests that mantle lithosphere

of Proterozoic and Archaean ages must have a density that is ~ 0.5 - 1.5% lower than the density of the underlying asthenosphere (Artemieva and Mooney, 2001). This is consistent with xenolith studies that show that Proterozoic and Archaean mantle lithosphere can be less dense than primitive mantle (e.g., Poudjom Djomani et al., 2001; Lee et al., 2011). The strong and buoyant nature of the cratonic mantle is thought to generally prevent its destruction by delamination or Rayleigh-Taylor-type instabilities, although there is abundant evidence that craton destruction has occurred in geologic history (Li et al., 2015 for an overview).

Several mechanisms have been proposed to explain craton destruction from below. For example, Gorczyk et al. (2012) show in a geodynamic modeling study that lateral variations in mantle lithosphere rheology may trigger Rayleigh-Taylor instabilities in intra-cratonic regions, resulting in craton mantle lithosphere removal. This may have contributed to destruction of the central Australia craton (Gorczyk et al., 2012). Refertilisation of cratonic mantle lithosphere (Tang et al., 2013) will reduce its viscosity while increasing its density (Hu et al., 2000; Li et al., 2008), promoting delamination. The Colorado Plateau (Levander et al., 2011), Wyoming craton (Boyd et al., 2004) and North China craton (Gao et al., 2004; Wu et al., 2005; Zhao and Zheng, 2007) are thought to have been thinned by this mechanism. Thermo-chemical destruction by mantle flow may be effective after the thermal boundary layer of the lithosphere has been eroded by asthenospheric shear (Cooper and Conrad, 2009).

Asthenosphere erosion may also affect the marginal regions of cratons, resulting in craton destruction from its side. The North China and Yangtze cratons were both weakened along their margins before destruction (Li et al., 2015). Miocene erosion of Colorado Plateau lithosphere by small-scale upper mantle convection (Karlstrom et al., 2008; van Wijk et al., 2010) may have removed mantle lithosphere material from the edge of this Proterozoic block. Also, the western

side of the North American craton may have been eroded by edge-driven convection (Hardebol et al., 2012; Bao et al., 2014).

Because cratonic lithosphere is thicker than surrounding younger lithosphere, a step in lithospheric thickness occurs across the margins of cratons. This has been imaged tomographically in several locations; Figure 2 illustrates three examples. Cratonic lithosphere is generally colder than younger lithosphere, and in tomographic images a craton boundary typically is manifested as a sharp seismic velocity gradient with faster velocities below the craton and lower velocities adjacent to the craton boundary. These tomographic images can be used to understand the geometry of a craton margin. The three examples in Figure 2 (the European Sorgenfrei-Tornquist Zone, the western margin of the Colorado Plateau, and western Canada) show sharp, steep lithospheric steps over short lateral distances, resulting in craton boundaries that vary between sub-vertical and dips larger than 45 degrees. The vertical size of a step may be 100 km or more. For western Canada, this geometry is inferred to have been in place for at least 50-100 Ma (Lowe and Ranalli, 1993; Bao et al., 2014); for the Colorado Plateau for 10-20 Ma (van Wijk et al., 2010), and for northern Europe, it may have been in place for 200-300 Ma (Hieronymus et al., 2007). As a result of the steep lithospheric step, there are large lateral thermal gradients across the margins of cratons. In western Canada, the thermal contrast between the mobile belt and craton has long been recognized through surface heat flow and other geophysical observations (Lowe and Ranalli, 1993; Hyndman and Lewis, 1999).

Such large thermal gradients will induce edge-driven convection (e.g., King and Anderson, 1998; Shahnas and Pysklywec, 2004; King, 2005; van Wijk et al., 2010; Hardebol et al., 2012). Edge-driven convection is small-scale (upper) mantle convection induced by the lateral variations in temperature (and therefore density). It creates downgoing flow below the

craton and upwelling flow below adjacent thinner lithosphere. The downgoing flow may destabilise cratonic mantle lithosphere when this is weak, dense, and has low viscosity (King and Anderson, 1998; van Wijk et al., 2010), thereby eroding the craton margin. This form of craton destruction has received less scrutiny than craton destruction from below and is therefore the focus of this study.

Here, we use simplified numerical models of the lithosphere and upper mantle to address the stability of an abrupt change in lithosphere thickness. Such a step in lithosphere thickness and the accompanying large thermal gradient may suddenly arise when plate tectonic processes affect and deform the mobile belts and younger lithosphere that surround cratonic lithosphere. Once a step is created, craton margins may either undergo erosion or remain stable and resistant to erosion. Our goals are to assess the conditions under which such a step in thickness can be maintained on timescales of 50 to 100 Ma and marginal erosion is limited. This timescale is motivated by the observations from western Canada, which suggest that the Cordillera-Craton transition was created by at least the Eocene (Bao et al., 2014) and perhaps prior to the Late Cretaceous (Lowe and Ranalli, 1993). The numerical models provide insight into the physical conditions needed to explain the seismic images of an abrupt contrast in lithosphere thickness (Figure 2) and provide constraints on the mechanisms of craton destruction through edge-driven instabilities. We further compare and contrast results from our models with earlier studies by Hardebol et al. (2012), Hieronymus et al. (2007) and Wang et al. (2014). The combined insight from these studies and the present study provides a framework for understanding the rheological and density conditions that protect craton margins from erosion while surrounding mobile belts are undergoing deformation.

2. Numerical models of craton margins

2.1 Model geometry and governing equations

The initial geometry of the numerical models is shown in Figure 3. The two-dimensional models start with the juxtaposition of continental lithospheres with thicknesses of 60 km and 160 km, and the thermal-mechanical evolution is monitored to determine how the transition in thickness changes with time. The initial transition corresponds to a vertical step in thickness between cratonic lithosphere and younger lithosphere of a mobile belt. This represents an end-member situation, for example following a tectonic event in which one area of continental lithosphere experiences abrupt thinning (e.g., extension of the Basin and Range Province creating a step in lithosphere thickness with the Colorado Plateau (van Wijk et al., 2010) or delamination of Canadian Cordillera lithosphere causing abrupt lithosphere thinning adjacent to the North American craton (Bao et al., 2014)), or through transcurrent plate movement which places a thin lithosphere next to thicker lithosphere (e.g., Stern et al., 2013).

The models are 2400 km wide and 600 km deep, and the lithospheric step is located in the middle of the model domain. The initial continental crust thickness is 40 km, consisting of a 25 km upper-mid crust and 15 km lower crust. The thickness and properties of the two crustal layers are uniform across the model. On the left side of the model, the mantle lithosphere is 20 km thick, giving a total lithosphere thickness of 60 km. This side represents younger lithosphere. On the right side, the mantle lithosphere is 120 km thick and the total lithosphere thickness is 160 km; this side represents older lithosphere. For simplicity, we also refer to the thinner lithosphere as the “mobile belt” and the thicker lithosphere as the “craton” when discussing the models.

The finite element code SOPALE is used to model the coupled thermal-mechanical

evolution of the system, subject to the conditions of plane strain and incompressibility (Fullsack, 1995). Material deformation is governed by the equations of conservation of mass (volume):

$$\frac{\partial v_j}{\partial x_j} = 0 \quad (1)$$

and conservation of force balance:

$$\frac{\partial \sigma_{ij}}{\partial x_i} + \rho g = 0 \quad (2)$$

where x_{ij} are spatial coordinates, v_{ij} are components of velocity, ρ is density, and g is (vertical) gravitational acceleration. Repeated indices imply summation and $i, j = 1, 2$ in the two-dimensional models. The components of the stress tensor are:

$$\sigma_{ij} = -P\delta_{ij} + \sigma'_{ij} = -P\delta_{ij} + 2\eta_{\text{eff}}\dot{\epsilon}_{ij} \quad (3)$$

where P is the total pressure, σ'_{ij} is the deviatoric stress tensor, δ_{ij} is the Kronecker delta (1 if $i=j$; 0 if $i \neq j$), η_{eff} is the effective viscosity, and the strain rate tensor is:

$$\dot{\epsilon}_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right) \quad (4)$$

The energy balance equation is used to calculate the temperature field:

$$\rho c_p \left(\frac{\partial T_K}{\partial t} + v_i \frac{\partial T_K}{\partial x_i} \right) = k \frac{\partial}{\partial x_i} \left(\frac{\partial T_K}{\partial x_i} \right) + A_T + \sigma'_{ij} \dot{\epsilon}_{ij} + v_2 \alpha g T_K \rho \quad (5)$$

where c_p is specific heat, T_K is the absolute temperature, t is time, k is thermal conductivity, A_T is volumetric radioactive heat production, and α is the volumetric thermal expansion coefficient.

The third term on the right side of equation (5) corresponds to strain heating, in which all dissipated mechanical energy is converted to heat, and the last term is the temperature correction for adiabatic heating for vertical velocity v_2 . In our models, the thermal and mechanical fields are coupled through the temperature-dependent density and viscous rheology of model materials, and the strain heating term in equation (5).

The SOPALE code uses an Arbitrary Eulerian-Lagrangian approach to solve the above

equations, subject to the boundary conditions described below (Fullsack, 1995). The governing equations are solved on an Eulerian finite element mesh that is fixed in space, except that it can stretch vertically to conform to the top model boundary as topography develops. Materials are tracked on a Lagrangian mesh and additional Lagrangian tracer particles. At each time-step, the Lagrangian particles are advected by the model velocity field, and the new distribution is used to update the material properties in the Eulerian mesh. The Eulerian mesh uses element widths of 10 km for the entire domain. Element heights are 2.5 km in the upper 50 km, 5 km between 50 and 400 km depth, and 10 km in the lower 200 km of the model domain. Previous benchmarking tests (e.g., Pysklywec and Beaumont, 2004; Currie et al., 2015) show that this mesh can resolve the growth rates of lithospheric gravitational instabilities to within a 6% of the expected values.

2.2 Material properties

Table 1 lists the material parameters that are used in the reference model; in subsequent models, the rheology and the density of the craton mantle lithosphere are varied. All materials have a viscous-plastic rheology. Materials follow a pressure-dependent Drucker-Prager yield criterion at shallow depths and high stresses:

$$J'_2 = c_0 \cos \phi_{\text{eff}} + P \sin \phi_{\text{eff}} \quad (6)$$

where c_0 is the cohesion, ϕ_{eff} is the effective internal angle of friction, and J'_2 is the square root of the second invariant of the deviatoric stress tensor ($J'^2_2 = 0.5 \sigma'_{ij} \sigma'_{ij}$). Frictional plastic deformation is modelled using an effective viscosity that places the state of stress on yield (Fullsack, 1995; Willett, 1999). At larger depths and lower stresses, deformation is viscous, with an effective viscosity given by:

$$\eta_{\text{eff}} = f(B^*) (\dot{\epsilon}_2)^{(1-n)/n} \exp \left(\frac{E^* + PV^*}{nRT_K} \right) \quad (7)$$

where $\dot{I}_2$ is the square root of the second invariant of the strain rate tensor ($\dot{I}_2^2 = 0.5\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}$), f is a scaling factor, B^* is the pre-exponential factor, n is the stress exponent, E^* is the activation energy, V^* is the activation volume and R is the universal gas constant. The four rheological parameters (B^* , n , E^* , and V^*) are obtained from laboratory studies of deformation of crustal and mantle rocks. The scaling factor (f) is used to linearly scale the effective viscosity up/down relative to the reference laboratory materials. This allows us to vary the material strength to account for variations in material composition relative to the laboratory materials and/or uncertainties in the rheological parameters (see Beaumont et al., 2006 for a discussion). There is significant uncertainty in the application of laboratory-derived deformation parameters to deformation at geological time and space scales. For example, it is necessary to extrapolate the laboratory strain rates over many orders of magnitudes. Bürgmann and Dresen (2008) however show that laboratory-based rheologies are in good agreement with geodetic and field observations of deformation, suggesting that laboratory-derived parameters provide a good description of geological deformation.

In our models, we base the material rheologies on several well-established flow laws. The crustal rheology parameters follow those used by Beaumont et al. (2006). In all models, the upper-mid crust rheology is based on a wet quartzite rheology (Gleason and Tullis, 1995), with a scaling factor of $f=5$ assuming that the crust is more mafic and less hydrated than wet quartzite. The lower crust uses the dry Maryland diabase rheology of Mackwell et al. (1998), with $f=0.1$ to approximate a more hydrated material. The rheology parameters of the mantle lithosphere and sublithospheric mantle are based on the wet olivine dislocation creep flow law of Hirth and Kohlstedt (2003). In this flow law, the pre-exponential factor (B^*) in Equation 7 includes a term for water content:

$$B^* = \frac{1}{2} \left(\frac{3^{(n+1)/2}}{2} \right)^{-1/n} (A_{\text{uni}} C_{\text{OH}}^r)^{-1/n} \quad (8)$$

where A_{uni} is the experimentally-derived constant, C_{OH} is the water content in ppm H/Si, and r is the water content exponent (Hirth and Kohlstedt, 2003). The first term in brackets converts the rheological parameters from uniaxial laboratory experiments to the tensor invariant state of the numerical models. As a reference, we use a water content of 1000 ppm H/Si, which is comparable to the average mantle water content inferred from mid-ocean ridge studies (Hirth and Kohlstedt, 1996). In the reference model, this rheology is assigned to the sublithospheric mantle and the mantle lithosphere of both continental regions.

All materials have a temperature-dependent density:

$$\rho(T) = \rho_0 (1 - \alpha(T - T_0)) \quad (9)$$

where α is the volumetric thermal expansion coefficient and ρ_0 is the reference density at temperature T_0 . The reference density and temperature for each model material are consistent with expected densities for crust and mantle rocks (e.g., Poudjom Djomani et al., 2001; Beaumont et al., 2006). In the reference model, all mantle materials (mantle lithosphere, sublithospheric mantle) have the same reference density, and there is no compositional density difference between different mantle materials.

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2.3 Initial and boundary conditions

Models are initialised with the material distribution shown in Figure 3. The initial temperature structure is given by one-dimensional conductive geotherms based on the lithosphere thickness at each point in the model (Figure 4a). The base of the lithosphere corresponds to the intersection of the geotherm and a mantle adiabat with potential temperature 1300°C and adiabatic gradient of 0.4°C/km. The geotherms give a Moho temperature of 989°C

and 450°C for the thin and thick lithosphere, respectively. These are consistent with the Moho temperatures for the Cordillera and craton regions of western Canada (e.g., Lowe and Ranalli, 1993; Hyndman and Lewis, 1999; Currie and Hyndman, 2006; Hyndman et al., 2009). The sublithospheric mantle initially has an adiabatic thermal structure. We do not explicitly model convection in the sublithospheric mantle prior to the start of the models, i.e., to generate the adiabatic temperatures. However, as the models run, lateral density variations induce thermal convection in the sublithospheric mantle.

Boundary conditions are shown in Figure 3. The upper boundary of the model is a stress-free surface with $T=0^{\circ}\text{C}$, and thus, topography can develop dynamically during model evolution. Topography is not part of our discussion below; we focus here on the stability of the boundary between cratonic and younger lithosphere. The side boundaries are insulating (no horizontal heat flow) and no-slip. The bottom boundary is a no-slip boundary with $T=1540^{\circ}\text{C}$. Tests show that the use of free-slip conditions on the sides and bottom does not affect the overall model behaviour, as these boundaries are far from the area of interest (the lithosphere step).

The model runs consist of two stages. The models are run for a short time to bring the two lithospheres into isostatic equilibrium. In this stage, the left-hand boundary of the model has a no-slip boundary condition and the right-hand boundary has a free-slip condition. This allows the craton lithosphere to move vertically and in most cases, the craton subsides relative to the thinner lithosphere. After isostatic equilibrium is established, both side boundaries are set to be no-slip boundaries, and the models are run to 100 Ma to assess the behaviour of the lithosphere step.

3. Reference model

In the reference model, the only difference between the thick lithosphere (craton) and the adjacent thin lithosphere (mobile belt) is the initial temperature structure (Figure 4a); the craton mantle lithosphere is an average of $\sim 250^{\circ}\text{C}$ cooler than the mobile belt. Both mantle lithospheres and the sublithospheric mantle have the same rheology and density properties (Table 1). Owing to the cooler temperatures within the craton lithosphere, its density is greater than that of the sublithospheric mantle. The thermal density contrast increases from 0 kg/m^3 at the base of the craton lithosphere to over 90 kg/m^3 at the Moho (Figure 4b).

The evolution of this model is shown in Figure 5. At the start of the model run, the step change in lithosphere thickness induces edge-driven convection at the margin of the craton lithosphere. The craton lithosphere has a downward velocity owing to its cool temperature and high density; this is coupled with an upwelling below the adjacent thin lithosphere. By 10 Ma, the corner of the craton is gravitationally removed as a narrow ($\sim 20\text{ km}$) drip. Over time, additional gravitational instabilities are induced at the base of both the thin and craton lithosphere, resulting in time-dependent small-scale downwellings of the cool lower lithosphere. The craton material then circulates within the sublithospheric mantle (blue material in Figure 5). After 100 Ma, convection has removed over 40 km of the deep craton lithosphere. In addition, the edge of the craton has been eroded, creating a transition from thin to thick lithosphere over a horizontal distance of 200-300 km.

4. Effect of craton properties on craton margin stability

4.1 Origin of density and rheology variations of cratons

The main objective of our study is to determine the conditions under which a steep (i.e., $>45^{\circ}$) lateral change in lithosphere thickness at the craton boundary can be maintained over

timescales of at least 50-100 Ma following tectonic rejuvenation (thinning) of adjacent lithosphere; this would point to a stable craton margin. The sharp boundary between cratonic and younger lithosphere (Figure 2) would be erased by edge-driven convection if cratonic mantle lithosphere is easily removed by this process. Gravitational stability of mantle lithosphere depends on both its rheology and density structure (e.g., Houseman and Molnar, 1997; Conrad and Molnar, 1999). In the reference model (Figure 5), the cooler craton lithosphere temperatures result in a mantle lithosphere that is more dense than the sublithospheric mantle (Figure 4a) and is gravitationally unstable, while its rheology is weak enough so that the craton corner may deform and drip. This leads to a smoothing of the lithosphere step (Figure 5). This is an example of an unstable craton margin where the craton is eroded from its sides. Below, the rheology and density of the craton mantle lithosphere are varied to understand the conditions that keep the craton margin intact.

An important control on the strength of the mantle is its water content. Here, we focus on the effect of water on olivine strength as olivine is the most abundant mineral in the upper mantle and as it is believed to be the weakest, its rheology should control the overall strength of mantle rocks (Karato, 2008). As water is added to olivine, it becomes weaker (e.g., Karato and Wu, 1993; Hirth and Kohlstedt, 1996; 2003; Karato, 2008). Figure 4c shows the variation in olivine viscosity for changes in water content (equations 7 and 8), using the craton geotherm from our models. Relative to our reference value of $C_{OH}=1000 \text{ H}/10^6 \text{ Si}$ (damp olivine), water contents of 100 and 10 $\text{H}/10^6 \text{ Si}$ result in a viscosity increase by a factor of 2.2 and 4.8, respectively (Figure 4d). Also shown in Figure 4c is the calculated viscosity for a water-saturated olivine based on the Hirth and Kohlstedt (2003) wet olivine flow law and the olivine water-saturation curve of Hirschmann et al. (2005) (Figure 4a). This predicts that the viscosity of water-saturated olivine

is 0.3-0.6 that of the reference damp olivine (Figure 4d). Hirth and Kohlstedt (2003) provide rheological parameters for dry olivine ($A_{\text{uni}}=1.1 \times 10^{-16} \text{ Pa}^{-3.5} \text{ s}^{-1}$; $n=3.5$; $E^*=530 \text{ kJ/mol}$, $V^*=15 \text{ cm}^3/\text{mol}$). A comparison of the viscosity from this flow law and that of the reference damp olivine (Figure 4c) shows that the dry olivine is 6 to 19 times stronger (Figure 4d). In the models below, we vary the rheology of the craton mantle lithosphere by scaling its effective viscosity relative to the reference damp olivine rheology. We test scaling factors of $f=0.5$ (i.e., half the strength) to $f=50$, which covers the predicted strengths for all water contents from dry to water-saturated.

One hypothesis for the longevity and thickness of cratonic mantle is that it is chemically depleted and therefore its density is lower than that of the underlying mantle (e.g., King, 2005 and Lee et al., 2011 for overviews). Poudjom Djomani et al. (2001) suggest that Archaean lithosphere may be as much as 80 kg/m^3 less dense than primitive mantle (at the same temperature) and that Proterozoic mantle may be up to 50 kg/m^3 less dense. We assume a constant reference density throughout the cratonic mantle lithosphere and we investigate densities that are up to 80 kg/m^3 less than that of the surrounding mantle (at the same temperature). This difference in density is also called the compositional density difference ($\Delta\rho$). We also consider models in which the reference density of craton mantle lithosphere is 20 and 40 kg/m^3 greater than that of the surrounding mantle. It is suggested that some craton lithosphere may have formed through stacking of subducting oceanic plates (e.g., Lee et al., 2011 and references therein). In this case, the craton lithosphere may contain small regions of eclogitised crust, which can have a density greater than that of mantle. Craton mantle with an excess density of 20 kg/m^3 or 40 kg/m^3 corresponds to an eclogite content of $\sim 7\%$ or $\sim 13\%$, respectively, based on an eclogite density of 3550 kg/m^3 (Austrheim et al., 1997).

4.2 Model results

Figure 6 shows a summary of the model results for different combinations of craton compositional density difference and rheology. Each model was run for 100 Ma and the behaviour of the lithospheric step was monitored. Models are grouped according to the dynamics of the step and its geometry at 50 and 100 Ma. To quantitatively describe the geometry, we measure the dip angle of both the 1200°C contour and the material boundary between lithosphere and sublithospheric mantle at depths of 80 to 140 km. This is based on the idea that seismic tomography velocities (e.g., Figure 2) are most sensitive to temperature in the mantle, with composition being a secondary effect, except in cases where the craton is strongly chemically distinct (e.g., Goes et al., 2000). For all models, these two measures yield similar angles for the transition from thin to thick lithosphere over the duration of the model run. Below, we describe the results of selected models that demonstrate the range of behaviours.

Figure 7 shows the effect of variations in craton compositional density for two models in which the craton mantle lithosphere is 10 times stronger than the reference model ($f=10$). If the craton lithosphere has the same reference density as the underlying mantle (Figure 7a), the step in lithosphere thickness is gravitationally unstable, but the instability is smaller and slower than in the reference model. By 50 Ma, the lithosphere step still has a relatively steep slope ($>45^\circ$). After 100 Ma, ongoing erosion gradually reduces the slope to less than 45° , but the craton inboard of the margin has not been thinned. If the craton compositional density is 40 kg/m^3 greater than that of mantle (Figure 7b), the craton margin is more unstable, which leads to a more rapid erosion of the lithosphere step. These models demonstrate that a high craton viscosity promotes stabilization of the lithosphere step. However, this is not sufficient to maintain a

lithosphere step angle $>45^\circ$ after 100 Ma, as observed for example in the Sorgenfrei-Tornquist tomographic transect (Figure 2a).

An alternate way to stabilise the craton is to reduce its mantle lithosphere density. Models with a compositionally buoyant craton (40 kg/m^3 less dense than sublithospheric mantle) are presented in Figure 8. With this compositional buoyancy, the lower $\sim 55 \text{ km}$ of the craton mantle lithosphere is less dense than the sublithospheric mantle, even though it is cooler. Figure 8a shows a model in which the craton lithosphere has the reference rheology (damp olivine). The step in lithosphere thickness is gravitationally unstable and the craton lithosphere is weak enough to undergo viscous flow. Instead of downwelling, lowermost craton mantle lithosphere viscously spreads and flows laterally, owing to its positive buoyancy. It then upwells and underplates the adjacent thin lithosphere. Over 100 Ma, the craton lithosphere has flowed more than 200 km outboard, leading to thinning of the craton near its original margin and a more gradual transition ($<30^\circ$ dip) between thin and thick lithosphere. The outward flow of buoyant craton lithosphere is reduced by increasing the lithosphere strength. If the strength is increased by an order of magnitude (Figure 8b), there is very little removal of the edge of the craton lithosphere and the lithosphere step remains steep over the 100 Ma model run.

Figure 9 shows the surface heat flow for the two models in Figure 8, as well as the reference model (Figure 5). These three models exhibit the range of behaviours observed in our models. In all models, the mobile belt area has relatively high heat flow ($60\text{--}70 \text{ mW/m}^2$) and the craton block heat flow is $40\text{--}50 \text{ mW/m}^2$. However, the transition from high to low heat flow varies depending on the dynamics of the step. With a stable step (model in Figure 8b), the heat flow transition occurs over a lateral distance of $\sim 150 \text{ km}$ and does not change significantly during the model run. If the step destabilizes through downwelling (model in Figure 5), the

transition migrates into the craton and becomes slightly broader over time. If the step that destabilizes through lateral spreading (model in Figure 8a), the transition migrates into the mobile belt and is even broader (>300 km); the effects become most pronounced late in the model run, as the mobile belt crust is cooled by the underplated cratonic lithosphere. Thus, the location of the heat flow transition relative to the geological boundary of the craton at the surface may be diagnostic of the deeper dynamics of the craton margin.

4.3 Regime diagram

The behaviour of the craton margin for the range of craton strengths and densities that we have tested is summarised in the regime diagram in Figure 6. In models where the craton compositional density is more than ~ 30 kg/m³ less than that of underlying mantle, the edge of the craton undergoes destabilization through lateral flow (as in Figure 8a); at higher densities, the lithosphere step is eroded through downwellings. The lateral flow is driven by the compositional buoyancy of the mantle lithosphere, which makes it less dense than the sublithospheric mantle despite its lower temperatures. We have tested models in which the material densities do not depend on temperature (a thermal expansion coefficient of 0 K⁻¹). In models where the cratonic mantle density is even slightly less than that of the sublithospheric mantle, the cratonic mantle spreads laterally, as it is compositionally buoyant. With no density difference, the mantle materials are stationary and cratonic mantle is stable.

The behaviour of the edge of the craton lithosphere depends on: 1) the gravitational buoyancy forces which act to destabilise the step, and 2) the viscous strength of the cratonic mantle, which acts to resist destabilisation. To maintain a steep ($>45^\circ$) step for over 100 Ma, the craton lithosphere must be relatively strong. For a craton density that is equal or greater than that

of mantle or that is strongly depleted ($>60 \text{ kg/m}^3$ less dense than mantle), the craton viscosity must be more than an order of magnitude higher than that of damp olivine (at the same temperature); this is greater than the strength of dry olivine. The required strength is reduced if the craton density is $20\text{-}60 \text{ kg/m}^3$ less than that of mantle. The compositional positive buoyancy is partially offset by the negative buoyancy from the cool craton temperatures, and therefore the overall gravitational instability of the cratonic mantle lithosphere is reduced and a weaker rheology can remain stable. For a craton with a compositional density that is 40 kg/m^3 less dense than mantle, the craton margin can be preserved if its strength is a factor of 3-4 times greater than that of damp olivine.

4.4 Effect of a dry olivine (strong) rheology for craton mantle

We have also tested models with the rheological parameters for dry olivine from Hirth and Kohlstedt (2003) (Figure 4c). At depths greater than $\sim 100 \text{ km}$ in craton lithosphere, dry olivine is 6-8 times stronger than the reference damp olivine (Figure 4d). At shallower depths, dry olivine is even stronger. This is primarily because its activation energy (E^*) is larger than that of damp olivine, resulting in a greater temperature-dependence. Figure 10 shows the results of models in which the craton compositional density is varied. When the craton has the same compositional density as the underlying mantle (Figure 10a), the lithospheric step is gravitationally unstable. It is gradually eroded through small-scale downwellings, and after 50 Ma, the step has an angle less than 45° . If the craton composition density is 40 kg/m^3 less than mantle (Figure 10b), the lower part of the craton is buoyant and tends to upwell. However, the dry rheology prevents significant deformation of the craton margin, and a steep lithosphere step is maintained over 100 Ma. After an initial period of lateral upwelling (first 50 Ma), the step

appears to become steeper as counter-clockwise edge-driven convection adjacent to the step “pushes” the craton lithosphere back toward the margin. A craton with a lower compositional buoyancy (60 kg/m^3 less dense than mantle) is more unstable (Figure 10c). It flows outward up to $\sim 100 \text{ km}$ from the original craton margin and buoyantly underplates the mobile belt lithosphere. This creates a low-angle transition from thin to thick lithosphere.

The results of these three models are consistent with the behaviour of our models in which the damp olivine rheology was scaled upward by a factor of 5-10 (Figure 6). For this strength, a steep lithosphere step can only be maintained over 100 Ma if the craton lithosphere has a moderately low density (compositional density $20\text{-}40 \text{ kg/m}^3$ less than mantle). At higher densities, the craton margin is eroded through downwellings; at lower densities, the margin spreads outward and the cratonic mantle underplates the adjacent lithosphere.

4.5 Effect of mantle wind on craton margin stability

Our models are designed to examine the stability of a craton margin within 100 Ma of tectonic deformation of a mobile belt, and therefore we do not explicitly include a convecting sublithospheric mantle (c.f., Wang et al., 2014). However, a large-scale horizontal mantle flow (mantle wind) may affect the predicted conditions required to maintain a steep step (Figure 6). To test this, we have run several models with a 2 cm/yr horizontal flow in the sublithospheric mantle. Figure 11 shows models in which the craton lithosphere has a compositional density 40 kg/m^3 less than the mantle and a rheology that is 5 times stronger than the reference damp olivine. With no mantle wind, the craton margin maintains an angle $>45^\circ$ over the 100 Ma model evolution (Figure 11b).

In Figure 11a, the 2 cm/yr mantle wind flows to the left, from the craton toward the mobile belt. This causes a slight leftward viscous shearing of the deep craton lithosphere. As the mantle flow passes below the corner of the craton, a low pressure is created in the shallow sublithospheric mantle next to the craton and a counterclockwise flow is induced in this area. This results in a slight steepening of the craton margin, compared to the model with no mantle wind. However, the difference between the models are relatively small and we conclude that a 2 cm/yr mantle wind directed toward the mobile belt has little effect on the boundary between the “steep” and “shallow” margins in Figure 6.

In contrast, a craton-ward mantle wind results in a noticeable decrease in the angle of the step (Figure 11c). The deep craton lithosphere near the margin is viscously sheared by mantle flow and carried toward the craton. This results in a thickening of the craton lithosphere by ~20 km near the boundary of the craton, as the buoyant lithosphere accumulates here. In models where the mantle lithosphere is more dense than mantle (not shown), the lithosphere is gravitationally removed, causing a thinning of the craton edge and a further decrease in the angle of the lithosphere step. For a mantle wind in this direction, the critical strength needed to maintain a steep step may shift to a slightly stronger rheology (Figure 6), especially for cases where the craton mantle lithosphere has a compositional density greater than that of the underlying mantle.

4.6 Effect of a thicker craton lithosphere

The above models address the stability of the edge of a 160-km thick craton. We have tested additional models that have a craton thickness of 260 km, resulting in a 200 km step in thickness at the craton margin. Figure 12 shows two models with a craton mantle lithosphere

that follows a dry olivine flow law (Hirth and Kohlstedt, 2003). If there is no compositional density difference between the craton and sublithospheric mantles (Figure 12a), edge driven convection results in some removal of the craton mantle lithosphere. This primarily affects the lithosphere below ~180 km depth. The shallower lithosphere remains stable, owing to its cool temperatures and thus high viscosity. Thus, a steeper thickness transition at the craton margin can be maintained over 100 Ma than in a comparable model with a thinner lithosphere (Figure 10a).

Figure 12b shows a model in which the craton mantle lithosphere is 40 kg/m³ less dense than the sublithospheric mantle. The lower 20-30 km of the craton flows laterally and underplates the mobile belt; more craton material is displaced than in a similar model with a thinner lithosphere (Figure 10b). The difference in behaviour is due to the differing thermal structures. With a thicker lithosphere, a greater amount of the deep lithosphere is at temperatures greater than 1000-1100°C, and thus it has both a low density and relatively low strength. As a result, the craton step is smoothed more quickly when the lithosphere thickness is greater.

Overall, a thick craton lithosphere with a relatively low compositional buoyancy (0-20 kg/m³ less dense than sublithospheric mantle) is more stable due to cooler temperatures and higher viscosities. Therefore, the strengths in Figure 6 are maximum values. A greater compositional density difference between the craton and sublithospheric mantle decreases the stability of the craton, and therefore it may be more difficult to maintain a steep margin.

5. Discussion: what controls craton margin stability?

Previous geodynamic modeling studies (e.g., Doin et al., 1997; Shapiro et al., 1999; King, 2005 and references therein; Lenardic et al., 2003; Sleep, 2003; Wang et al., 2014) primarily

addressed the preservation of wide areas of thick (>200 km) cratonic lithosphere above a convecting mantle. They commonly conclude that thick, cool craton lithosphere is prone to gravitational thinning if it has the same composition as the underlying mantle; the high viscosities associated with cool craton temperatures are insufficient to stabilise the lithosphere. The main factor that can allow thick craton lithosphere to be maintained is a rheologically strong lithosphere. In successful models of craton preservation, the cratonic mantle is 10 to >100 times more viscous than the convecting mantle (at the same temperature, pressure and strain rate conditions) (e.g., Lenardic et al., 2003; Sleep, 2003; Wang et al., 2014). Moderate chemical buoyancy of cratonic mantle lithosphere aids in its preservation, but on its own, buoyancy does not ensure craton stability (e.g., Lenardic et al., 2003; King, 2005). Wang et al. (2014) show that even with a buoyant mantle lithosphere, cratons must be a factor of 3 more viscous than convecting mantle (at the same conditions) in order to remain intact. A strong and buoyant cratonic mantle lithosphere is typically attributed to dry and chemically depleted conditions. This composition is consistent with xenolith studies of cratonic composition (e.g., Lee et al., 2001; Poudjom Djomani et al., 2001; Lee et al., 2011).

How are the margins of cratonic lithosphere preserved? The sparse seismic studies that cross craton margins show that these boundaries correspond to a transition from relatively high shallow mantle velocities below the craton to lower velocities below the adjacent lithosphere, indicating a change in lithosphere thickness (Figure 2). In several places, such as northern Europe (e.g., Goes et al., 2000; Hieronymus et al., 2007) and western Canada (e.g., Goes and van der Lee, 2002; Bao et al., 2014; Y.J. Gu, personal communication, 2015), the craton margin corresponds to a relatively steep ($>45^\circ$) gradient in lithosphere thickness; we used this gradient to discriminate between preserved and unstable craton margins in our model analyses. The main

result from our models is that the thicker craton lithosphere must have a distinct rheology (and possibly density) relative to the surrounding mantle in order to maintain a steep transition over 50-100 Ma. Without a difference in properties, the cool craton lithosphere is eroded by edge-induced convection and its own gravitational instability. For all densities that we have tested, the craton mantle lithosphere viscosity must be significantly stronger than the reference rheology of damp olivine (Figure 6). Strengthening by more than a factor of 10 is required if the cratonic mantle is either compositionally dense ($\Delta\rho \geq 0 \text{ kg/m}^3$) or highly buoyant ($\Delta\rho < -40 \text{ kg/m}^3$). For a moderate compositional buoyancy, the required strength is slightly lower, as the compositional buoyancy offsets some of the negative thermal buoyancy of the cool, thick region. If a dry olivine rheology is an upper limit on the strength of cratonic mantle, our models predict that a steep gradient in lithosphere thickness can be preserved only if the craton mantle is approximately 20 to 40 kg/m^3 less dense than typical mantle.

The overall behaviour of our models is similar to that seen in previous modeling studies that include a lateral change in lithosphere thickness. In Figures 4e and 4f, we show for comparison the viscosity flow laws in three studies (Hieronymus et al., 2007; Hardebol et al., 2012; and Wang et al., 2014) plotted for the conditions in our models. We note however, that the original studies have different model properties (e.g., lithosphere thicknesses, thermal gradients, etc.), as mentioned hereafter. Hieronymus et al. (2007) address the stability of a lithosphere step at the Sorgenfrei-Tornquist Zone in northern Europe (Figure 2a). Their reference (weak) rheology is somewhat weaker than our reference, and they conclude that strengthening by a factor of 50 is required in order to maintain a lithosphere step that agrees with seismological observations; they do not consider variations in density. This magnitude of strengthening is consistent with that predicted by our models for cratons with $\Delta\rho = 0 \text{ kg/m}^3$ (Figure 6).

Hardebol et al. (2012) developed numerical models to address the mantle dynamics at the transition in lithosphere thickness in western Canada (Figure 2c). They find that the thick craton lithosphere is readily eroded with a weak wet olivine rheology, but it can be maintained for 100 Ma if the craton has a dry olivine rheology (Hirth and Kohlstedt, 1996); they also do not consider variations in craton density. Hardebol et al. (2012) do not include an activation volume in the rheological formulation (Equation 7) and thus their dry olivine rheology is only slightly stronger than our reference damp olivine rheology (Figure 4f). In our models, we find that a much stronger rheology is needed to maintain the change in lithosphere thickness. A possible reason for the different results between their and our models is that our models use a vertical step in lithosphere thickness whereas Hardebol et al. (2012) implement a 60 km or 100 km change in thickness over a horizontal distance of 250 km. With the more gradual lateral gradient, the vigor of edge-driven convection is reduced, which may allow for a greater range of stability than we find. Further, the craton lithosphere in their models is somewhat cooler than in our models, which results in a higher viscosity that may help stabilise the craton.

Finally, Figure 4f shows the reference rheology used by Wang et al. (2014) in their study of craton stability over 2 Gyr, which is a factor of 3-4 stronger than our reference damp olivine rheology. Their models show that cratonic lithosphere that is 3-10 times stronger than this (i.e., 12-40 times stronger than our reference rheology) can remain stable. To preserve a stable craton, the craton strength does not have to be this high if the craton has some compositional buoyancy; Wang et al. (2014) consider a two-layer craton lithosphere with a boundary at 144 km depth and chemical buoyancy of 70 kg/m³ and 42 kg/m³ in the upper and lower layers, respectively. Thus, their overall conclusions are similar to what we find in our study. This suggests that the factors that promote the longevity of thick cratonic lithosphere also enable its margins to be preserved as

areas of steep gradients in thickness. In other words, a thick, strong and buoyant craton should be characterised by steep changes in lithosphere thickness at its margins.

Our models also show that a lithospheric step is readily eroded if the craton mantle is weaker than dry olivine for most of the densities we have tested (Figure 6). If the craton density is similar or greater than that of the surrounding mantle, destabilization occurs through downwellings that progressively erode the craton inward, resulting in the gradual decrease in the relief on the base of the lithosphere at the craton margin. This result is consistent with seismic observations and magmatic data, and surface topography at places such as the Rio Grande Rift (van Wijk et al., 2008), Colorado Plateau (van Wijk et al., 2010; Figure 2b), and the North Island of New Zealand (Stern et al., 2013). In these places, it is inferred that continental lithosphere is actively thinned through convective instabilities induced by the lateral variation in lithosphere thickness.

If the craton mantle lithosphere is both weak and highly buoyant ($\Delta\rho < -40 \text{ kg/m}^3$), we find that a step in lithospheric thickness destabilises through lateral gravitational spreading, such that buoyant craton lithosphere underplates the adjacent mobile belt lithosphere. This behaviour is also seen in the numerical models of Wang et al. (2014), as well as models of lithosphere dynamics at passive margins (e.g., Huisman and Beaumont, 2011; Beaumont and Ings, 2012; Armitage et al., 2012). As noted by Beaumont and Ings (2012), cratonic lithosphere is generally inferred to be both chemically buoyant and anhydrous (and therefore strong), owing to partial melting and chemical depletion during craton formation. In order for lithosphere to be weak enough to flow laterally, either the initial depletion event must not have fully dehydrated the lithosphere, or the lithosphere experienced later refertilisation or metasomatism that resulted in weakening but no significant density increase. At passive margins, lateral flow of buoyant craton

lithosphere has been invoked to explain exhumation of continental mantle lithosphere, magma and xenolith geochemistry and temporal changes in topography (Huisman and Beaumont, 2011; Beaumont and Ings, 2012; Armitage et al., 2012). We are not aware of observational evidence of this process at the boundary of cratonic lithosphere in continental interiors (K. Condie, personal communication, 2015), but a more thorough examination of geophysical and geological data is needed. Kaban et al. (2015) propose that the deep root beneath the Superior craton (Figure 1) has shifted laterally, away from its original position, by basal drag resulting from mantle flow. Underplating of buoyant craton mantle lithosphere as predicted by some of our models would be an alternative mechanism for root displacement.

In our study, we have used simplified numerical models to demonstrate the dynamics that may occur where there is a step in lithosphere thickness at a craton margin. Through a systematic variation in the rheology and density of the cratonic lithosphere, our models constrain the conditions that may permit a steep gradient in lithosphere thickness to be maintained (Figure 6). In doing this, we have made several assumptions. For example, the majority of our models consider a 100 km vertical step between the mobile belt and craton lithosphere, with a 160 km thickness for the craton. A thicker craton may be more stable, unless its compositional density is significantly different than that of the sublithospheric mantle (Section 4.6). Similarly, as discussed above, a more gradual step in lithosphere thickness may lead to a greater degree of craton stability. Furthermore, our models do not include mantle convection or plate tectonic processes. A horizontal mantle flow of 2 cm/yr does not significantly affect the model behaviour over a timescale of 100 Ma, especially if the mantle flow is directed toward the mobile belt (Figure 11). However, larger flow velocities or interactions between the craton margins and plate margin processes (e.g., a subducting plate) may have a greater effect. We have also not

considered the effect of vertical layering within the craton mantle lithosphere or variations in crustal strength. These may introduce weak interfaces to the lithosphere, which may promote destabilisation and should be examined in future studies. Finally, our models address stability on a timescale of 100 Ma. At longer timescales, the effects of conductive heating and continued shearing from mantle flow may start to erode the step. In this case, the preservation of the step may also depend on the structure of the adjacent lithosphere and the degree to which it shields the craton boundary from mantle stresses (e.g., Lenardic et al., 2000).

6. Conclusions

Seismic tomography data show that the boundary between cratonic and younger continental lithosphere corresponds with a sharp gradient in upper mantle seismic velocities (e.g. Figure 2). In several places, craton margins corresponds to a steep ($>45^\circ$) gradient in lithosphere thickness from >100 km below the craton to <100 km below the adjacent continent. We have used numerical models to understand how variations in the rheology and density of the craton lithosphere affect the geometry of the craton margin. We conclude the following:

- in accordance with earlier modeling studies of craton longevity, the rheology of the cratonic mantle lithosphere is the primary control on its stability. Cratonic mantle must have a stronger intrinsic rheology than the underlying mantle. The compositional buoyancy of cratonic lithosphere provides a secondary control, but even a buoyant craton requires some rheological strengthening to remain stable.
- if the craton mantle rheology follows existing olivine flow laws, a steep gradient in lithosphere thickness at the craton margin can only be maintained if the craton is dry

($C_{OH} < 10$ ppm H/Si) and has a moderate compositional buoyancy (20-40 kg/m³ less dense than underlying mantle).

- if the lithosphere is denser than this, a lithospheric step is eroded through downwellings that remove the edge of the craton margin. Conversely, if the lithosphere is more buoyant, it destabilises through laterally spreading, resulting in underplating of cratonic lithosphere below the adjacent lithosphere.

There are only a few tools available to study the dynamics of cratons over millions to billions of years. Xenoliths, magmatism and surface geology provide a detailed, but incomplete, picture of select regions. Most geophysical observations (e.g. seismic tomography) provide snapshots of present-day structure and therefore must be carefully interpreted to gain insight into dynamic processes. Numerical models, such as those presented here, provide an experimental framework for studying the temporal evolution of the lithosphere-upper mantle system. However, the models are necessarily simplified relative to the real Earth and use material properties that are often not well-constrained. Nevertheless, numerical models are useful because they provide insight into how different parts of the system may dynamically interact. They also provide testable predictions for how these dynamics may be recognised in surface geological and geophysical observations. For example, the models in this study predict that regions with a steep lithosphere gradient at the craton margin (e.g., western Canada, Sorgenfrei-Tornquist zone) have a dry and chemically depleted craton mantle lithosphere. This prediction should quantitatively be tested against geophysical observations such as seismic tomography (e.g., Figure 2), gravity and magnetotellurics, as well as geochemical studies of craton mantle lithosphere composition.

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Table 1. Material properties used in the reference model.

	Continental Upper-Mid Crust	Continental Lower Crust	Mantle Lithosphere	Sublithospheric mantle
<i>Plastic rheology</i>				
C ₀ (MPa)	20	0	0	0
φ _{eff} ^a	15-2°	15-2°	15-2°	15-2°
<i>Viscous rheology</i>				
f	5	0.1	1	1
A _{uni} (Pa ⁻ⁿ s ⁻¹)	1.10×10 ⁻²⁸	5.05×10 ⁻²⁸	9.0×10 ⁻²⁰	9.0×10 ⁻²⁰
C _{OH} (H/10 ⁶ Si)	n/a	n/a	1000	1000
r	0	0	1.2	1.2
B*(Pa ⁻ⁿ s ⁻¹ H/10 ⁶ Si) ^b	2.92×10 ⁶	1.91×10 ⁵	7.79×10 ³	7.79×10 ³
n	4.0	4.7	3.5	3.5
Q (kJ mol ⁻¹)	223	485	480	480
V* (cm ³ mol ⁻¹)	0	0	11	11
<i>Thermal parameters</i>				
k (W m ⁻¹ K ⁻¹)	2.5	2.5	2.5	2.5
A (μW m ⁻³)	1.1	0.4	0	0
c _p (J kg ⁻¹ K ⁻¹)	1250	1250	1250	1250
<i>Density</i>				
ρ ₀ (kg m ⁻³)	2800	3000	3250	3250
T ₀ (°C)	627	627	1327	1327
α (K ⁻¹)	3.0×10 ⁻⁵	3.0×10 ⁻⁵	3.0×10 ⁻⁵	3.0×10 ⁻⁵

^a Strain-softening is incorporated through a decrease in φ_{eff} over accumulated strain of 0.5 to 1.5 (e.g., Huisman and Beaumont, 2003).

^b $B^* = \frac{1}{2} \left(\frac{3^{(n+1)/2}}{2} \right)^{-1/n} (A_{uni} C_{OH}^r)^{-1/n}$

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FIGURES

Figure 1. Age distribution of continental lithosphere (after Artemieva and Mooney, 2001). Archean cratons and Colorado Plateau are numbered (1-15, 16). 1: Slave craton, 2: Wyoming craton, 3: Superior, 4: Greenland, 5: Fennoscandian, 6: Siberian, 7: North China craton, 8: west Australian, 9: Indian Shield, 10: Tarim, 11: Tanzanian craton, 12: Zimbabwe and Kaapvaal cratons, 13: Congo, 14: West African, 15: Amazonian, 16: Colorado Plateau. Locations of tomographic transects that are shown in Figure 2 are indicated: STZ is Sorgenfrei-Tornquist zone, CP is Colorado Plateau, and CA is western Canada.

Figure 2. Seismic tomography images across: a) the Sorgenfrei-Tornquist Zone in northern Europe (after Hieronymus et al., 2007), b) the western Colorado Plateau (after Sine et al., 2008), and c) western Canada (after Bao et al., 2014). Profile locations are shown in Figure 1. LAB is lithosphere-asthenosphere boundary, HVB is high velocity block, BRP is Basin and Range Province, CP is Colorado Plateau, RGR is Rio Grande rift, JdF plate is Juan de Fuca plate. Interpretations of LAB from Hieronymus et al. (2007) and from Bao et al. (2014).

Figure 3. Initial geometry of the numerical models showing the distribution of materials and the thermal-mechanical boundary conditions.

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Figure 4. a) Initial geotherm applied to the 60 km thin lithosphere (red line) and 160 km thick lithosphere (black line). For each region, the lithosphere-asthenosphere boundary (LAB) corresponds to the intersection of the conductive geotherm with the mantle adiabat (grey line). The dashed blue line shows the maximum water content in olivine as a function of depth (Hirschmann et al., 2005). b) Initial density profile for the thin (red line) and thick (black line) lithosphere based on their temperature. The inset shows the density contrast relative to the sublithospheric mantle; this density contrast arises from differences in temperature only. c) Effective viscosity as a function of depth for the thin lithosphere (red) assuming damp olivine ($C_{OH}=1000$ ppm H/Si) rheology and a strain rate of 10^{-15} s^{-1} . Blue, black, grey and green lines show the effective viscosity for the thick lithosphere for water content ranging from dry to water-saturated. Rheologies are from Hirth and Kohlstedt (2003). d) The relative viscosity of the thick lithosphere for varying water content; the reference rheology is damp olivine ($C_{OH}=1000$ ppm H/Si). e) Effective viscosity profiles for the weak (solid lines) and strong (dotted lines) rheologies used in selected previous modeling studies. These are calculated using the 160 km thick lithosphere geotherm and a strain rate of 10^{-15} s^{-1} . f) Relative viscosity of the rheologies in (e) compared to the reference damp olivine in (c).

Fig. 4

Figure 5. Evolution of the reference model, in which the craton mantle lithosphere has the same reference density as the underlying mantle ($\Delta\rho = 0 \text{ kg/m}^3$) and a damp olivine rheology. Yellow is upper crust, red is lower crust, and blue is mantle lithosphere material. Sub-lithospheric mantle is white (Figure 3). Arrows indicate flow direction and magnitude.

Figure 6. Regime diagram showing the behaviour of the lithosphere step for models in which the craton strength and compositional density are varied. Destabilization of the step occurs through either downwelling (downward triangles) or lateral spreading and underplating (upward triangles); circles denote models with little lithosphere destabilization. The dashed grey line and thick black line show the conditions that give a 45° slope at the craton margin after 50 Ma and 100 Ma, respectively. The craton strength was varied through a linear scaling of a damp olivine rheology ($C_{OH}=1000$ ppm H/Si). Shaded grey region and shaded red region show the approximate strength of water-saturated and dry olivine, respectively (see Figure 4).

Figure 7. Evolution of models in which the craton lithosphere is 10 times stronger than the reference damp olivine rheology. a) Craton mantle lithosphere has the same compositional density as the underlying mantle. b) Craton mantle lithosphere compositional density is 20 kg/m³ greater than that of the underlying mantle.

Figure 8. Evolution of models in which the craton lithosphere is compositionally buoyant (40 kg/m^3 less dense than the underlying mantle). a) Craton mantle lithosphere has the reference damp olivine rheology. b) Craton mantle lithosphere is 10 times stronger than the reference rheology.

Figure 9. Surface heat flow at 50 and 100 Ma for models that show the key three behaviours observed in our study: stable craton margin (solid line), craton margin destabilisation through downwelling (black dashed line) and destabilisation through underplating (grey dashed line). The vertical grey line is the original location of the craton margin. The legend gives the strength and compositional density contrast ($\Delta\rho$, in kg/m³) of the craton mantle for each model. The weak craton has the reference damp olivine rheology; the strong craton is 10 times stronger.

Figure 10. Evolution of models in which the craton lithosphere has a dry olivine rheology, and a compositional density that is: a) 0 kg/m³, b) 40 kg/m³, and c) 60 kg/m³ less than that of the underlying mantle.

Figure 11. The effect of a mantle wind (horizontal mantle flow) on the behaviour of the lithospheric step: a) 2 cm/yr to the left, b) no mantle wind, c) 2 cm/yr to the right. In all models, the craton mantle lithosphere is 5 times stronger than the reference damp olivine and its compositional density is 40 kg/m^3 less than that of the underlying mantle.

Figure 12. Evolution of models in which the craton lithosphere is 260 km thick with a dry olivine rheology, and a compositional density that is: a) 0 kg/m^3 and b) 40 kg/m^3 less than that of the underlying mantle.