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Longevity and stability of cratonic lithosphere: Insights from numerical simulations of coupled mantle convection and continental tectonics

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[1] The physical conditions required to provide for the tectonic stability of cratonic crust and for the relative longevity of deep cratonic lithosphere within a dynamic, convecting mantle are explored through a suite of numerical simulations. The simulations allow chemically distinct continents to reside within the upper thermal boundary layer of a thermally convecting mantle layer. A rheologic formulation, which models both brittle and ductile behavior, is incorporated to allow for plate-like behavior and the associated subduction of oceanic lithosphere. Several mechanisms that may stabilize cratons are considered. The two most often invoked mechanisms, chemical buoyancy and/or high viscosity of cratonic root material, are found to be relatively ineffective if cratons come into contact with subduction zones. High root viscosity can provide for stability and longevity but only within a thick root limit in which the thickness of chemically distinct, high-viscosity cratonic lithosphere exceeds the thickness of old oceanic lithosphere by at least a factor of 2. This end-member implies a very thick mechanical lithosphere for cratons. A high brittle yield stress for cratonic lithosphere as a whole, relative to oceanic lithosphere, is found to be an effective and robust means for providing stability and lithospheric longevity. This mode does not require exceedingly deep strength within cratons. A high yield stress for only the crustal or mantle component of the cratonic lithosphere is found to be less effective as detachment zones can then form at the crust-mantle interface which decreases the longevity potential of cratonic roots. The degree of yield stress variations between cratonic and oceanic lithosphere required for stability and longevity can be decreased if cratons are bordered by continental lithosphere that has a relatively low yield stress, i.e., mobile belts. Simulations that combine all the mechanisms can lead to crustal stability and deep root longevity for model cratons over several mantle overturn times, but the dominant stabilizing factor remains a relatively high brittle yield stress for cratonic lithosphere. *INDEX TERMS:* 8110 Tectonophysics: Continental tectonics—general (0905); 8121 Tectonophysics: Dynamics, convection currents and mantle plumes; 8159 Tectonophysics: Rheology—crust and lithosphere; *KEYWORDS:* craton, continental tectonics, mantle convection, tectosphere

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1. Introduction

[2] Oceanic lithosphere is relatively short-lived. The operation of plate tectonics efficiently recycles the entire oceanic lithosphere back into the Earth's mantle on a timescale of 10^8 years, much shorter than the geologic age of the Earth. The age of the continental crust makes it clear that continental lithosphere, unlike oceanic lithosphere, is not efficiently recycled as a whole. However, the preservation of continental crust at the Earth's surface says nothing about the longevity of the deeper lithosphere. Evidence for the longevity of deep continental lithosphere comes instead from kimberlite pipes that have been erupted within continental cratons.

[3] The depth of origin for kimberlites, combined with their rapid eruption rates, has allowed them to remove bits of the mantle lithosphere on their ascent to the Earth's surface. This, in turn, has provided us with direct samples of deep cratonic lithosphere. Geothermobarometry, combined with dating of garnet inclusions from diamond xenocrysts within kimberlites, provided the first evidence that the

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lithosphere below the Kaapvaal craton of Africa was 200 km thick over 3 Gyr ago [*Richardson et al.*, 1984; *Boyd et al.*, 1985]. The ancient crystallization age of diamonds, combined with the fact that they were not erupted to the surface until some 100 Myr ago, further suggested that deep cratonic lithosphere served as a long-term storage reservoir that was isolated from recycling into the convecting mantle for billions of years [*Boyd et al.*, 1985]. Since these first Kaapvaal studies, similar studies have supported the view that ancient cratonic lithosphere has remained thick and isolated from mantle recycling below several other cratons including the Superior craton of North America [*Bell and Blenkinsop*, 1987] and the Siberian craton of Asia [*Pearson et al.*, 1995; *Richardson and Harris*, 1997].

[4] The inferred longevity of cratonic lithosphere is only interesting in context. If we lived on a single plate planet, such as present-day Mars or Venus, the longevity of any portion of the lithosphere would hardly be a surprise as the bulk of the lithosphere is permanent on a single plate planet for so long as a single plate state prevails [*Solomatov and Moresi*, 1995]. The longevity of cratonic lithosphere becomes intriguing when viewed relative to the more fundamental observation that the entire oceanic lithosphere is relatively short lived. The evidence that large portions of noncratonic, continental mantle lithosphere avoid recycling is sparse relative to the evidence from cratons and several studies suggest that the mantle lithosphere below noncratonic continental regions can be recycled on a relatively rapid timescale [e.g., *Bird*, 1979; *Tao and O'Connell*, 1992; *Willet et al.*, 1993; *Houseman and Molnar*, 1997; *Rowland and Davies*, 1999; *Pysklywec et al.*, 2000]. Thus the question is not why is deep cratonic lithosphere long-lived but how can it remain long-lived when the majority of the Earth's lithosphere is efficiently recycled.

[5] Most studies that have addressed the question above owe a debt to *Jordon* [1975, 1978], who reasoned, on the basis of a variety of observations, that the subcrustal lithosphere below cratons must be chemically distinct from the convecting mantle. This chemically distinct, deep cratonic lithosphere has come to be termed a "cratonic root", also referred to as the tectosphere, and the majority of ideas related to the longevity of cratonic lithosphere attribute it to the chemical buoyancy and/or the high viscosity of root material [e.g., *Jordon*, 1975, 1978; *Pollack*, 1986; *Shapiro*, 1995; *Doin et al.*, 1997; *Shapiro et al.*, 1999; *Sengor*, 1999].

[6] There is evidence that root material is chemically lighter than reference mantle [e.g., *Boyd*, 1989]. It is also likely that it has a high viscosity due to cool conditions and a dehydrated nature [*Pollack*, 1986]. However, it would be incorrect to conclude from this evidence that the compositional density anomaly associated with continental roots is sufficient to ensure their longevity; oceanic crust is chemically buoyant relative to reference mantle but it is recycled because its positive chemical buoyancy is overcome by the negative thermal buoyancy of subducting slabs. Similarly, young oceanic lithosphere as a whole is buoyant relative to reference asthenosphere yet it can be recycled if attached to an older section of subducting lithosphere. The same caution holds for connecting root longevity to high root viscosity; oceanic lithosphere has a very high viscosity relative to the bulk mantle, but this does not stop it from being recycled at subduction zones.

[7] Taken together, these observations suggest that we cannot fully explore the idea that the buoyancy and/or viscosity of cratonic roots is indeed what gives them longevity, without considering how buoyant and/or how viscous roots must be in order to resist being pulled into the mantle at a subduction zone. It is also possible that cratons avoid regions of mantle downflow. If this is the case, then the longevity of cratonic lithosphere is not principally tied to its material properties. We will return to explore this idea later; we begin by exploring the more prevalent idea that the material properties of cratonic lithosphere are somehow the fundamental keys to root longevity.

[8] To explore the specific material properties that can allow for cratonic root longevity, we have undertaken a suite of numerical simulations that send a model continent, with a cratonic root, into a model of a subduction zone environment. As well as deep lithospheric longevity, the models are also used to address the related question of what provides for the tectonic stability of cratonic crust. The modeling approach and numerical solution techniques are first described. Simulation results that vary key parameters such as root buoyancy and viscosity are then discussed.

2. Mathematical Model, Simulation Setup, and Numerical Method

[9] The principal methodology we will use involves a suite of numerical simulations based on the solution of model equations describing thermal-chemical convection in the infinite Prandtl number limit. The mathematical model consists of conservation equations for mass, momentum, energy, and composition together with an equation of state and constitutive laws for each chemical component. The nondimensionalized conservation equations together with a linearized equations of state are given by [e.g., *Lenardic and Kaula*, 1996]

$$\partial_i u_i = 0 \tag{1}$$

$$\partial_j\left[2\mu(T, C, \tau_{\text{yield}}, D)\epsilon_{ij}\right] = \partial_i p + Ra(T - B_r C)\hat{k} \tag{2}$$

$$\partial_t T + u_i \partial_i T = \partial_i^2 T + H \tag{3}$$

$$\partial_t C + u_i \partial_i C = 0 \tag{4}$$

$$\rho = [1 - \alpha(T - T_0) + \beta(C - C_0)] \tag{5}$$

where

$$Ra = \frac{\rho_0 g \alpha \Delta T d^3}{\mu_0 \kappa}$$

$$B_r = \frac{\beta \Delta C}{\alpha \Delta T}$$

$$H = \frac{\rho_0 Q D^2}{\Delta T K}$$

$$\alpha = -\frac{1}{\rho}\left(\frac{\partial\rho}{\partial T}\right)_{p,C}$$

$$\beta = \frac{1}{\rho}\left(\frac{\partial\rho}{\partial C}\right)_{p,T}$$

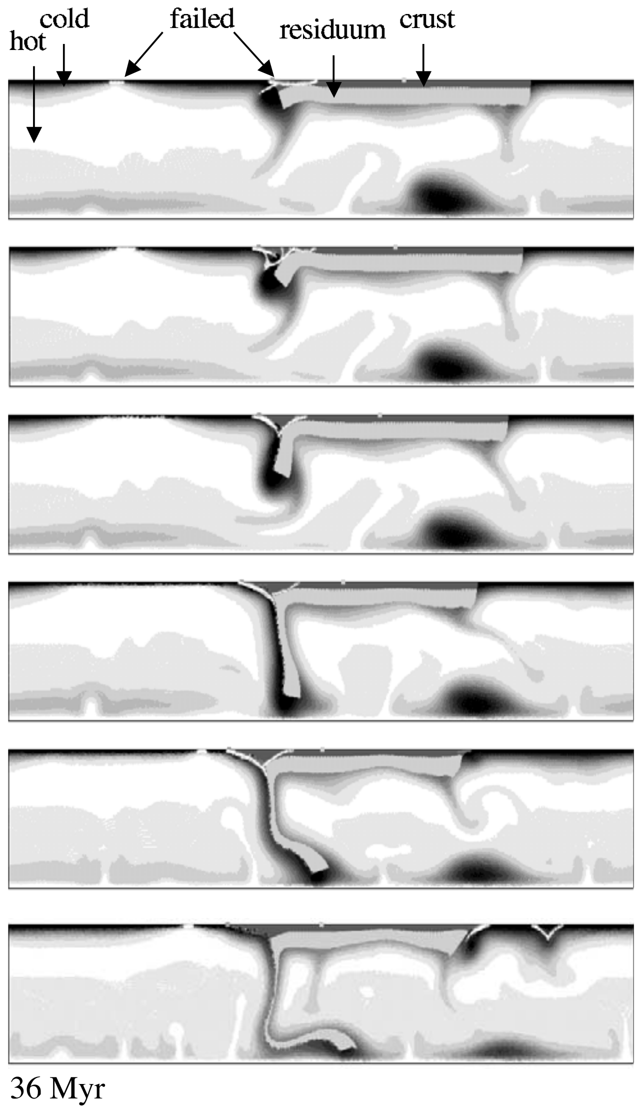

Figure 1. Reference model evolution. The grey scale image plots display composition, the mantle temperature field, and regions where stresses have exceeded the failure limit causing concentrated failure zones to form. Chemically distinct crust and residuum (cratonic root material) are displayed as unique grey scale tones, as labeled in the top panel. The mantle thermal field is displayed as 10 equally spaced grey shade isotherms with black being the coldest temperature and white the hottest. Failure zones appear as narrow white zones. Brittle failure requires cool temperature conditions which allows failure zones to always be visible within the image plots as these zones appear within otherwise darkly colored regions, i.e., cold regions.

and u_i is the velocity vector, $\mu(T, C, \tau_{\text{yield}}, D)$ is a viscosity function, T is temperature, C is a composition marker function used to distinguish chemically distinct components, τ_{yield} is a material yield stress, D is the second invariant of the strain rate tensor, ε_{ij} is the strain rate tensor, p is the pressure, Ra is the thermal Rayleigh number for bottom heating, B is the compositional to thermal buoyancy ratio, $\hat{k}$ is the vertical unit vector, H is the ratio of the thermal Rayleigh number defined for internal heating to that defined for bottom heating, ρ is the density, α is the coefficient of thermal expansion, β is its compositional equivalent, g is gravitational acceleration, ΔT is the temperature drop across the full system, d is the full system depth, μ_0 is the reference viscosity defined at the system base, ΔC is the maximum compositional variation, κ is thermal diffusivity, Q is the rate of internal heat generation per unit mass, and K is thermal conductivity.

[10] At this stage it is useful to describe the general setup of the numerical simulations we will explore as this will motivate our choice of model constitutive equations and will provide a framework for discussing the numerical method employed. This is best done by considering a prototype simulation near its initial start time. The top frame of Figure 1 serves this purpose. A chemically distinct continent resides within the upper thermal boundary layer of a convecting mantle layer. Three chemically distinct materials are present: Continental crust, subcrustal continental mantle lithosphere, and bulk mantle. All materials have unique reference densities with different values implying different degrees of chemical buoyancy. Local buoyancy ratios can be defined for the crustal and the continental mantle lithosphere component relative to the bulk mantle. These are defined as

$$B_c = \frac{\rho_m - \rho_c}{\rho_m \alpha \Delta T} \tag{6a}$$

and

$$B_{scml} = \frac{\rho_m - \rho_{scml}}{\rho_m \alpha \Delta T} \tag{6b}$$

respectively, where ρ_m, ρ_c, and ρ_{scml} are the reference densities of the bulk mantle, the crustal, and the subcrustal continental mantle lithosphere components, respectively. Physically, these ratios are the chemical density difference between a reference element of bulk mantle and a reference element of continental crust or subcrustal continental lithosphere divided by the maximum thermally induced density anomaly for mantle material. They thus provide a measure of chemical relative to thermal buoyancy forces. The vigor of thermal convection depends on the bottom heating Rayleigh number, Ra, and the heat ratio, H, which parameterizes the ratio of internal to bottom heating. Boundary conditions are free slip and isothermal for upper and lower surfaces. A wraparound boundary condition is used for vertical sidewalls to remove artificial edge boundary effects. The initial thermal field is obtained by running a simulation with a nondeformable continent for several convective overturn times. This leads to a thicker thermal boundary layer in continental versus oceanic regions and a continental thermal boundary layer that is locally thicker than the continental chemical boundary layer across a continent's extent, as must be the case when the system is at or near thermal equilibrium.

[11] The presence of a chemically distinct continent in Figure 1 is only one of the minimum set of requirements for the problem we wish to address. Our assertion that the stability of continental roots is only of interest in the context of plate tectonics requires the model oceanic lithosphere to participate in the convective circulation despite having a

higher-than-average viscosity through its cool temperature. This motivates our choice of material constitutive equations, i.e., our choice of rheologic assumptions.

[12] The self-consistent incorporation of lithospheric subduction and plate-like behavior into mantle convection models has experienced a recent surge [e.g., *Moresi and Solomatov*, 1998; *Tackley*, 1998; *Trompert and Hansen*, 1998]. These recent modeling studies are all based on a simple idea: Localized lithospheric failure at a critical stress level leads to the formation of weak faults or shear zones that allow otherwise cold and strong lithosphere to participate in convective mantle overturn. We follow this approach by incorporating a rheologic formulation akin to that used by *Moresi and Solomatov* [1998]. The rheology law for any chemical component remains on a temperature-dependent viscous branch for stresses below a specified yield stress, τ_{yield}, which can vary from component to component. Along this viscous branch, diffusion creep is considered to be the deformation mechanism and the viscosity function is given by

$$\mu_{\text{creep}} = A(C)\exp[-E(C)T] \tag{7}$$

where A and E are a preexponential constant and an activation energy, respectively (the dependence on C indicates that these material values can vary between components). For stresses above a yield stress, for any given component, the flow law switches to a depth-dependent plastic branch. The yield criteria and the form of the plastic flow law are based on a continuum representation of Byerlee's frictional law [*Byerlee*, 1968] and, in this sense, the plastic branch parameterizes brittle behavior within the lithosphere. Specifically, the yield criterion for any component is defined by

$$\tau_{\text{yield}} = \tau_0 + \tau_1 z \tag{8}$$

where τ_0 is the yield stress at zero hydrostatic pressure, τ_1 is the slope of the linear yield curve, and z is depth. The nonlinear, effective viscosity along the plastic deformation branch is given by

$$\mu_{\text{plastic}} = \frac{\tau_{\text{yield}}(C)}{D} \tag{9}$$

where D is the second strain rate invariant and the dependence on C indicates that different components can have different yield stress values.

[13] The rheologic formulation above, for any distinct chemical component, has followed *Moresi and Solomatov* [1998] and further details can be found therein. The models of this paper do extend the approach of *Moresi and Solomatov* [1998] in one regard. Along the plastic branch, a strain-dependent component of added weakening is introduced. This type of added weakening has been found to be important for generating plate-like behavior in mantle convection models [*Bercovici*, 1996; *Tackley*, 1998]. After a material fails, its yield stress decreases as a linear function of accumulated strain. The degree of added weakening is a variable, τ_w, that is expressed as a percentage (e.g., a value of 0.5 indicates that postyield weakening can lower a materials yield strength by, at most, 50% of its original value).

[14] Material yielding in Figure 1 is indicated by bright white zones that are marked as "failed" just above the top frame. The location of yielding in the top frame of Figure 1 reflects a key aspect of our modeling approach. After the nondeformable continent simulation had reached a statistically steady state, we monitored the simulation and stopped it at a point when the continent was about to move into an incipient subduction zone. At this point the assumption of a rigid continent was relaxed, continental material was given new material parameters, and the simulation was allowed to continue.

[15] The numerical solution strategy is based on a Lagrangian integration point finite element methodology that allows for the tracking of an arbitrary number of chemically distinct materials with unique physical properties [*Sulsky et al.*, 1995; *Sulsky and Schreyer*, 1996]. The approach has been extended to model localized shear band formation and to be able to deal with the large strains associated with mantle convection [*Moresi et al.*, 2001, 2002, 2003]. A large number of material points are embedded in the standard finite element mesh of the CITCOM finite element code [*Moresi and Solomatov*, 1995]. The material points form a Lagrangian reference frame which remains attached to the fluid as it moves. This allows composition, C, to be tracked as different particles can be tagged for specific chemical components (e.g., crust and mantle). The finite element mesh remains undeformed. The link between the Eulerian reference frame of the mesh and that of the particles is through the finite element integration scheme: The particles in a given element serve as the integration points of the element integration scheme. That is, they replace the standard fixed Gauss points generally used in finite element formulations.

[16] We have tested the finite element code against standard thermal mantle convection benchmark problems [*Blankenbach et al.*, 1989] and against thermal-chemical benchmark problems [*van Keken et al.*, 1997] with good results. We also performed convergence tests for our specific cratonic stability problem. Three different mesh densities where used for convergence testing: a 32 $\times$ 160 element mesh, a 48 $\times$ 240 mesh, and a 64 $\times$ 320 mesh (the domain aspect ratio being fixed at five). On the basis of tracking the Nusselt number, the RMS velocity, and the extent of a cratonic root (see below), we could determine that mesh errors on the intermediate mesh are not in excess of 5–7% for a particle density of at least nine material points per element.

3. Numerical Simulations

[17] There is a very large parameter space available to our model system so we have focused on the effects of continental buoyancy parameters and on the effects of rheologic parameters. Specific parameters that effect continental buoyancy are the reference densities of chemically distinct continental components and the relative amount of the components present (i.e., the lateral extent and depth of each chemical layer). Specific rheologic parameters for each chemically distinct component are (1) the cohesion term, τ_0, which sets the surface value of the material yield curve, (2) the effective friction coefficient, τ_1, which sets the depth-dependent slope of the yield curve, (3) the maximum

amount of postyield weakening, τ_w, (4) the preexponential term, *A*, in the temperature-dependent viscosity law, and (5) the activation term, *E*, of the temperature-dependent viscosity law.

[18] We have fixed the parameters that define the vigor of thermal convection in the mantle to constant values. For all simulations, $H = 1$ and $Ra = 2 \times 10^7$, based on the viscosity at the system base. For a convecting layer depth of 670 km, a driving temperature drop of 2000 K, and standard mantle thermal properties [*Turcotte and Schubert*, 1982], this implies an average viscosity of $\approx 10^{20}$ Pa s within the bulk interior of the mantle. The choice of upper mantle convection and a relatively low internal mantle viscosity can be justified for the problem at hand as the longevity of cratonic roots means we must consider not only present-day conditions but also past conditions when convection was more vigorous. Under such conditions the potential of mantle layering increases [*Christensen and Yuen*, 1985] and the interior viscosity of the mantle decreases [*Tozer*, 1972].

[19] The modeling domain size (dimensional values 670 km depth by 3350 km width), the reference density of mantle material (3300 kg/m^3), the initial extent of a continent (1340 km), the boundary conditions, and the initial conditions also remain constant for all simulations discussed. Finally, the temperature-dependent viscosity law used for all components allows for a factor of 10^5 viscosity variation from the maximum to the minimum system temperature. (e.g., the dimensional mantle viscosity at the surface is $\sim 10^{24}$ Pa s). The effects of intrinsic viscosity variations between chemical components are explored by varying the preexponential term in the viscosity laws for each material. This allows us to set the viscosity of root material, for example, to be higher than mantle material at equivalent temperatures.

[20] All modeling studies make simplifying assumptions that can effect quantitative results, and this should be kept in mind. Several of our main assumptions are geometric. Although our assumption of upper mantle convection may be justified for the problem at hand, it remains an assumption which may or may not be valid for the Earth over its full history. Further, we assume a single continent within a two-dimensional Cartesian domain of restricted extent. The fixed extent of the modeling domain as a whole also means that the lateral extent of our model continent is small relative to the average extent of the Earth's present-day continents. The fact that the uncertainties associated with the key rheologic and buoyancy parameters of our problem makes for a large modeling search space on its own is a key reason we have chosen simplified geometric assumptions. We will seek to isolate specific combinations of buoyancy and rheologic parameters that can allow for craton stability and root longevity within our geometrically simplified models with the knowledge that future studies must be undertaken to see how relaxing our geometric assumptions effects our conclusions.

[21] We performed over 150 numerical simulations to map a significant region of parameter space. A reference model is discussed first and parameter variation studies are then reported on. To condense results, we define a measure of cratonic longevity and stability as the normalized lateral extent of a cratonic root. The initial lateral extent of the undeformed root is used as the normalization factor. In all simulations, a point within the center of the craton at a depth just below the crust is tracked and the horizontal distance to the left and the right of it over which cratonic root material extends coherently is monitored. If root material is recycled into the mantle, this extent decreases. If a root is deformed to the point that it splits in two, the extent also decreases. Plotting this measure versus various model parameters is one method we use to present our results. Image plots, showing the evolution of representative simulations from the various parameter sets explored, are also used.

3.1. Reference Model

[22] The launching point for all the parameter variation subsets is a prototype simulation with nominal parameter values. For this simulation, the reference densities of the crust and subcrustal cratonic lithosphere are set to 2800 and 3200 kg/m^3, respectively. The initial thicknesses of the chemically distinct cratonic lithosphere and of the cratonic crust are 120 and 40 km, respectively. The yielding behavior of the mantle is determined by setting the cohesion term in the plastic flow law to 10 MPa, the effective friction coefficient to 0.1, and the postyield weakening factor to 0.5. These reference choices are motivated by the fact that we wish to allow for subduction of oceanic lithosphere within our simulations [*Moresi and Solomatov*, 1998]. The rheology of cratonic root material is considered equivalent to that of the mantle. The same is true for cratonic crust except that its viscosity is an order of magnitude lower than mantle at equivalent temperature [e.g., *Kirby*, 1985; *Kohlstedt et al.*, 1995]. The yield properties of the crust remain the same as those of the mantle. The choice of these values is simply to provide for a reference case. The effects of varying each parameter value will be explored in turn as will the effects of varying several parameters in unison.

[23] Figure 1 shows the dynamic evolution of the reference model. As the craton approaches the model subduction zone, slab induced stresses are sufficient to cause failure zones to form within the cratonic lithosphere. The second evolution frame shows that this failure generates detachment surfaces near the base of the crust which allow the subcrustal cratonic lithosphere to become decoupled from the crust above. A key effect is that the high chemical buoyancy of the crust can not contribute to the preservation of subcrustal lithosphere. Instead the subcrustal cratonic lithosphere is subducted in a manner reminiscent of the A-type subduction discussed by *Bally and Snelson* [1980] and subsequently taken up in models of doubly vergent orogen formation [e.g., *Willet et al.*, 1993].

3.2. Buoyancy Parameter Study

[24] The buoyancy structure of cratonic lithosphere can be varied by varying the crustal density, the subcrustal cratonic lithosphere density, the crustal thickness, and/or the thickness of the chemically distinct lithosphere.

[25] Figure 2 shows the effects of increasing the chemical buoyancy of cratonic root material. The reference density of root material in the simulation of Figure 2 is 3100 kg/m^3. Raising the root buoyancy does lead to significant changes in the exact manner by which root material is recycled into the convecting mantle, but it does not change the fact that it is recycled. Root recycling is no longer as coherent as it was in the simulation of Figure 1. It now involves the root being

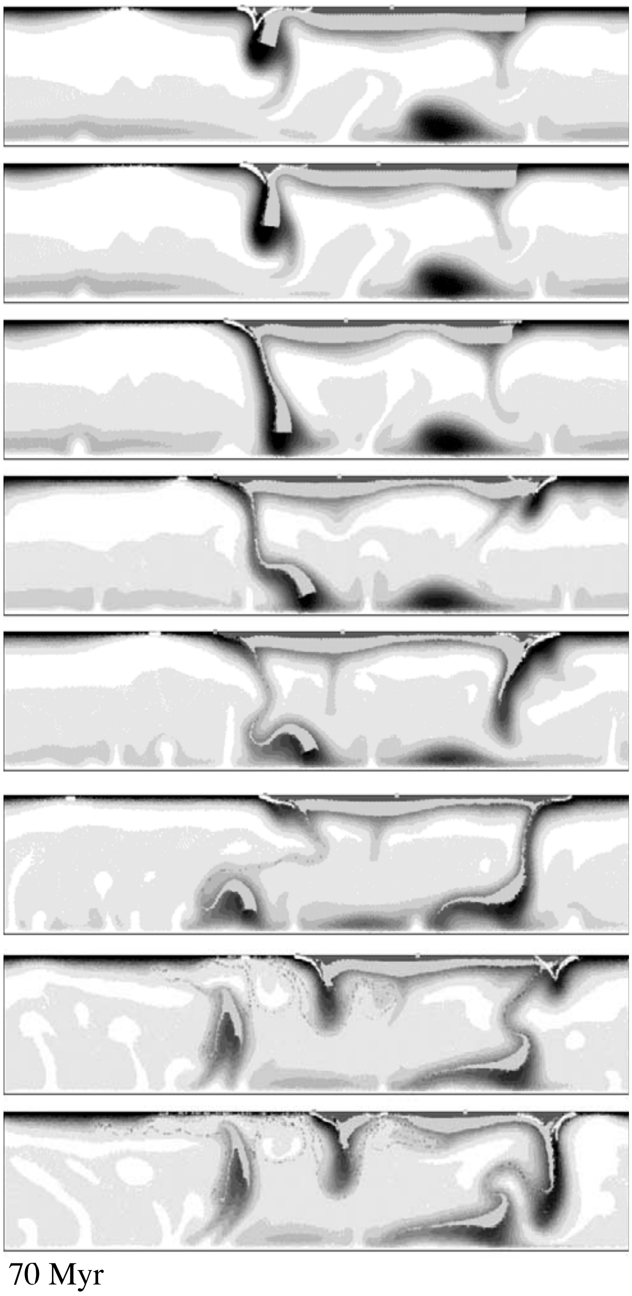

Figure 2. High root buoyancy model evolution.

stretched and split into small pieces that can be pulled into the mantle by cold subducting slabs. The total relative buoyancy of a detached parcel of root material depends not only on the buoyancy ratio but also on the size of the parcel. The simulation shows that the total buoyancy of a parcel of root material relative to the total negative buoyancy of cold subducting mantle is the key factor in determining if a parcel will be recycled. This in accord with theoretical considerations of root recycling in zones of dynamic mantle downflow [*Lenardic and Moresi*, 1999]. Thus, if a large root can be separated into smaller pieces, it can be progressively recycled in a piece by piece manner.

[26] In Figure 3a we plot the normalized root extent at different times from several simulations with variable reference densities of root material and variable root depths. Results are plotted in terms of the local buoyancy ratio between the subcrustal mantle lithosphere and the bulk mantle (equation (6b)). In Figure 3a a buoyancy ratio of 0.5 corresponds to a dimensional reference density of 3200 kg/m^3 for root material while a ratio of 1.0 corresponds to a root density of 3100 kg/m^3. This last value is greater than geochemically based upper bound estimates of root density [e.g., *Poudjom Djomani et al.*, 2001]. Figure 3b explores the effects of allowing for thicker chemical roots. A very thick root with a high degree of intrinsic chemical buoyancy tends to initially extend as buoyancy forces exceed the strength of the lithosphere. Subsequent to this initial phase, root recycling proceeds in a mode like that of Figure 2.

[27] The simulation results, contained within Figures 3a and 3b, suggest that the chemical buoyancy of root material is insufficient to provide for the stability of cratonic crust and/or the longevity of cratonic roots. This is not to say that root material is not chemically buoyant relative to reference

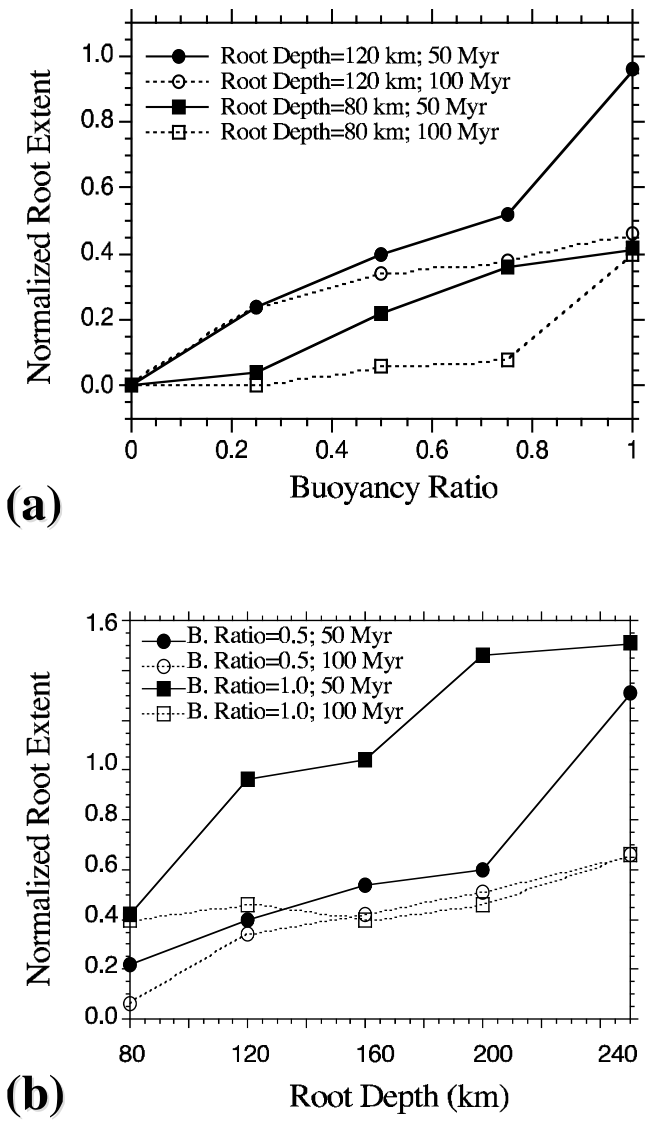

Figure 3. (a) Normalized root extent versus root buoyancy ratio. (b) Normalized root extent versus root depth.

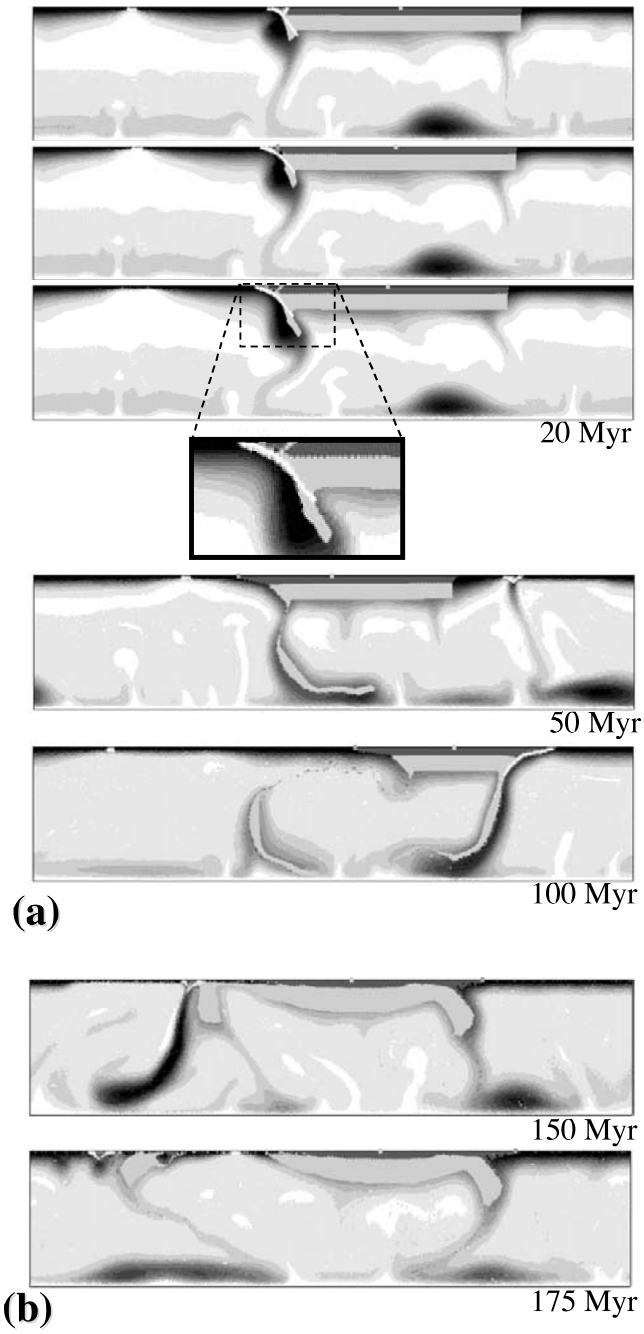

Figure 4. (a) High root viscosity model evolution. (b) high-viscosity, thick root model evolution.

mantle. Nor is it to say that root buoyancy may not partially contribute to stability and longevity. The simulations simply suggest that it is likely not the principal physical factor at work.

3.3. Viscosity Parameter Study

[28] Figures 4 and 5show the effects of varying intrinsic crustal and root viscosity parameters from the nominal simulation values. The most obvious viscosity parameter that could provide for root longevity is the intrinsic high viscosity of root material relative to the convecting mantle. Figure 4a shows the evolution of a simulation that is equivalent to the reference case except that the viscosity of root material is now set to be 1000 times that of the mantle at equivalent temperature. This is at the upper end of the strength contrast predicted from rheological experiments for dehydrated root material relative to bulk upper mantle [*Hirth and Kohlstedt*, 1996].

[29] Increasing the intrinsic viscosity of root material means that brittle behavior can occur at greater depths within the root. This is because the rheology, brittle versus ductile, of material at any depth is determined by the condition that stress be minimized. At the start of the simulation of Figure 4a, large stresses near the left continent margin cause failure within the oceanic lithosphere and within the upper portion of the cratonic lithosphere. As the simulation evolves, a failure zone propagates through the entire cratonic lithosphere near its left margin. This failure zone slices off a narrow portion of the subcrustal lithosphere. The sliver of cratonic lithosphere is entrained into the deeper mantle by the subducting slab. As the model continent moves leftward toward the subduction zone, large stresses continue to be generated near the craton margin due

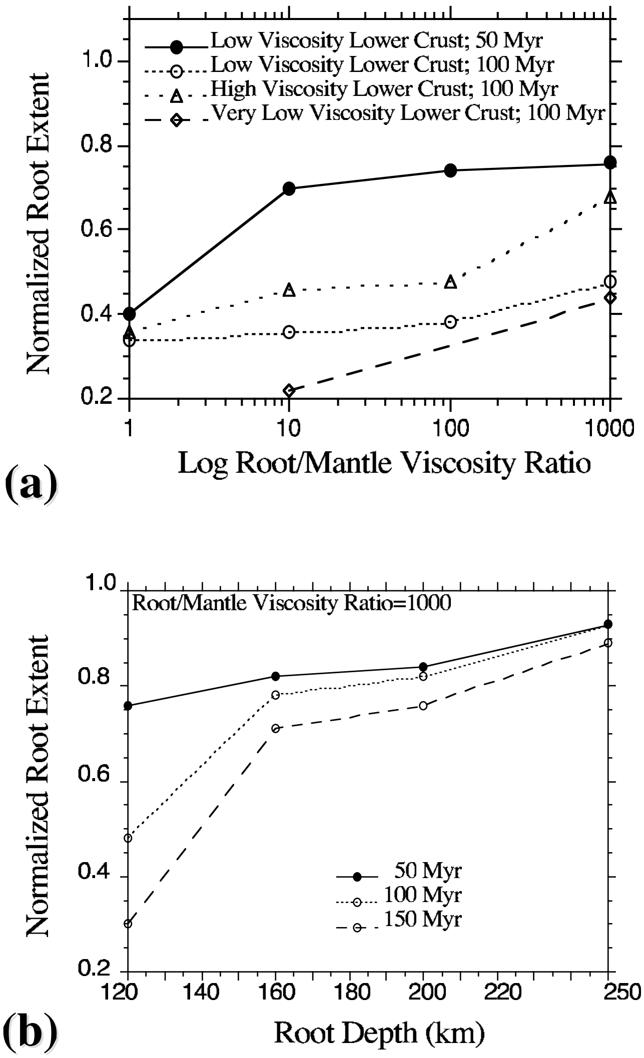

Figure 5. (a) Normalized root extent versus root to mantle viscosity ratio. (b) Normalized root extent versus root depth for models with a high root to mantle viscosity ratio.

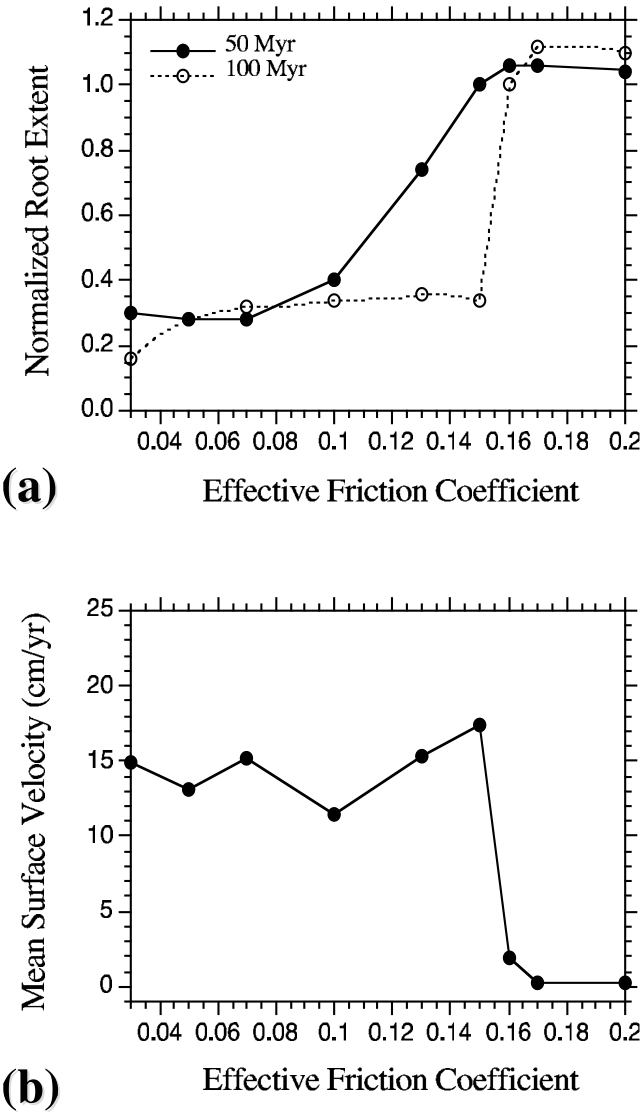

Figure 6. (a) Normalized root extent versus global friction coefficient value. (b) Root-mean-square velocity versus global friction coefficient value.

to the subducting slab underneath. This causes further failure of the cratonic lithosphere and allows added portions of it to be recycled into the deeper mantle. Over time, subduction ceases at the left continental margin and initiates to the right of the continent in the model oceanic region. The mantle flow that results eventually pulls the continent toward the subduction zone and cratonic recycling commences at the right continental margin as it enters into the region of subduction.

[30] The simulation of Figure 4a indicates that if the mineral assemblages that make up the lithosphere can fail at high stresses, then the nominal viscosity of deep root material alone is not what determines its strength. Increasing root viscosity can amplify stress levels within the lithosphere, and if the lithosphere can only maintain a fixed stress level before it fails, then recycling can still occur. Figure 5a further quantifies this. It also shows that the viscosity of the lower crust can effect root longevity. A low viscosity lower crust promotes detachment at the crust-mantle interface which removes the potential effects of crustal buoyancy on lithospheric preservation. The main message of these simulations, however, is the one previously noted: for the rheologic formulation employed, lithospheric strength does not equate solely to the ductile properties of the deep lithosphere. That is, the brittle yielding properties of the crust and mantle lithosphere must also be considered. The simulations of section 3.4 explore this issue.

[31] The slice and dice removal mode of Figure 4a relies on brittle failure extending across the majority of a cratonic root's depth extent. The stresses driving this brittle failure are associated with the subducting oceanic lithosphere that comes into contact with the cratonic root. This suggests that the slice and dice mode may have a geometric restriction. As a root becomes much thicker than the oceanic lithosphere, the depth extent over which it feels the stresses associated with the oceanic lithosphere can become small relative to its full depth extent and, as a result, brittle failure may not be able to extend across it. The simulations in Figure 5b explore this possibility and show that a thick viscous root limit does exist under which stability and longevity can be achieved. Before this limit is reached, there is also a regime under which roots can be long-lived but not tectonically stable (Figure 4b). If the thick viscous stability and longevity mode holds in nature, then simulations predict that the effective mechanical thickness of cratonic lithosphere should approach 100 km or more.

3.4. Yield Strength Parameter Study

[32] Figure 6a shows the results of varying the effective friction coefficient for all materials in unison. Once the value exceeds 0.15 a cratonic root can become long-lived. However, this is deceptive. In this case, the entire system has moved to a stagnant lid mode of convection [*Moresi and Solomatov*, 1998]. This can be seen in Figure 6b, which shows that once the effective friction coefficient globally exceeds 0.15 the surface velocity of the entire system drops almost to zero. That is, the simulations no longer allow for subduction of oceanic lithosphere but rather come to mimic a single plate planet. This is consistent with a previous study that explored plate generation in a convecting mantle layer without continents [*Moresi and Solomatov*, 1998]. Moresi and Solomatov found that an effective friction coefficient between 0.03 and 0.13 allowed for plate-like behavior and associated subduction of oceanic lithosphere.

[33] Figure 6b makes it clear that we must maintain the nominal friction coefficient value for bulk mantle if plate-like behavior and subduction of oceanic lithosphere is to be allowed for. We thus do not explore variations of bulk mantle yield parameters any further. We instead increase the effective friction coefficient of cratonic components while leaving the mantle value at a level that allows for subduction of oceanic lithosphere. Figures 7 and 8a show the results from several of these simulations. The ratio of the effective friction coefficient of cratonic relative to mantle components is termed a yield ratio. The friction coefficient determines the slope of the brittle yield curve and thus the maximum stress level the lithosphere sustains through viscous deformation before yielding. Thus, by varying the yield ratio we are assuming that cratonic

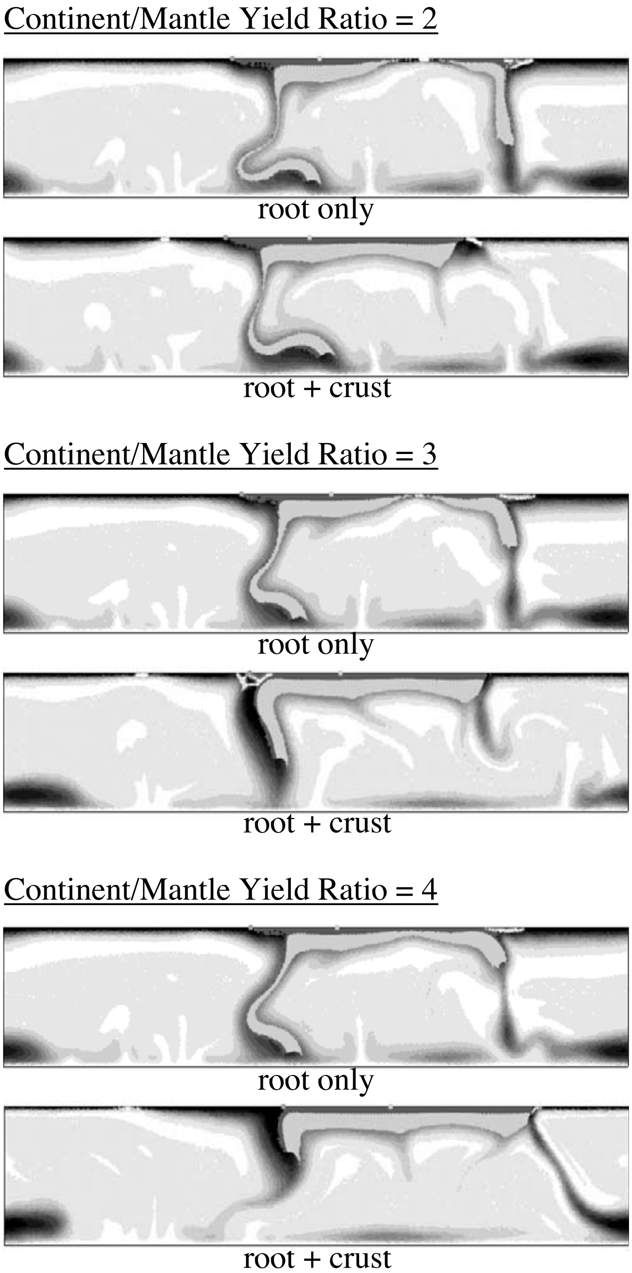

Figure 7. High continent yield models after 50 Myr evolution. Two situations are explored. For the first, an enhanced continental yield stress applies for the cratonic root and the continental crust, while for the second, an enhanced yield stress applies only for the root material.

components can maintain a greater stress level than reference oceanic lithosphere.

[34] The image plots of Figure 7 show the importance of having a high effective friction coefficient for the crustal, as well as the mantle, component of cratonic lithosphere. Cases in which the crust does not have a relatively high friction coefficient allow detachment surfaces to form within the lower crust. This decouples the buoyant crust from subcrustal lithosphere which favors the continued recycling of deep cratonic roots. Cases in which the crust does not have a high effective friction coefficient also favor rifting. The thickness of a high strength layer within the lithosphere will be greatest if both crust and mantle components can withstand high degrees of stress, i.e., if both have a relatively high effective friction coefficient. A lower friction coefficient within the crust tends to promote stress focusing within the high strength layer of the mantle and this, in turn, favors rifting [e.g., *Kusznir and Bott*, 1977]. The considerations above also hold true if only the crustal component of a craton is assigned a higher friction coefficient (this was confirmed in other simulations). An enhanced rifting potential does not favor tectonic stability, and it also means that cratonic roots can be broken into smaller pieces and recycled into the mantle. Thus a relatively high friction coefficient for both the mantle and crust does appear to be required for tectonic stability and root longevity (Figure 8a).

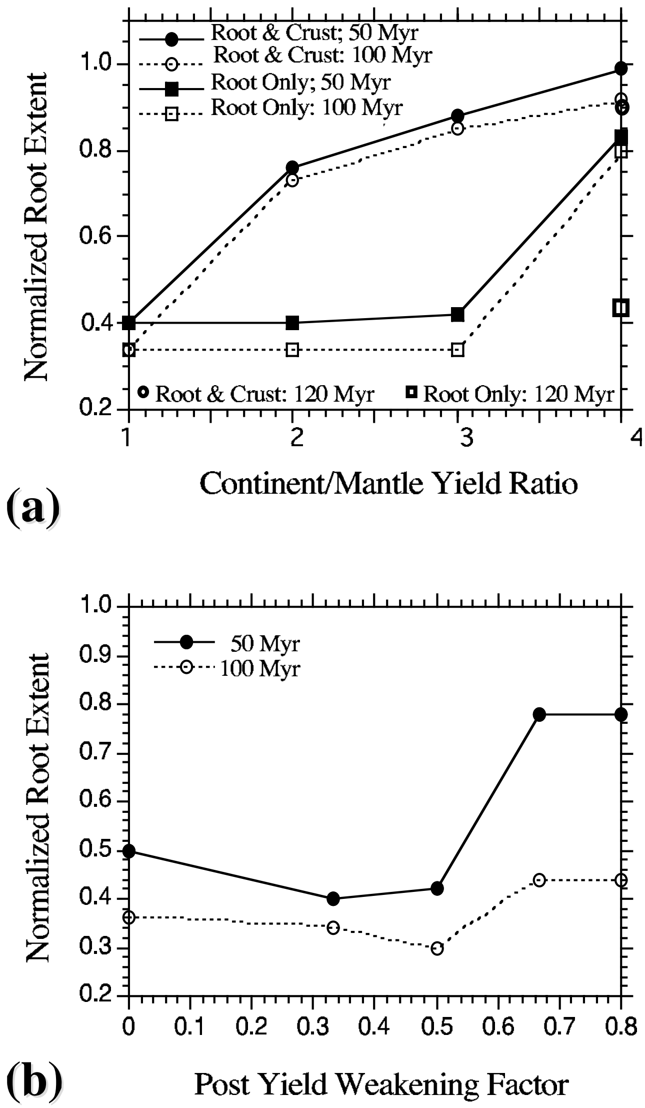

Figure 8. (a) Normalized root extent versus continent to mantle yield ratio. Two simulation sets are considered. For one set both crust and root material have enhanced yield while for the other only the root has an enhanced yield stress. (b) Normalized root extent versus postyield weakening factor.

[35] Increasing the cohesion term of the yield strength curve had a qualitatively similar effect to increasing the friction coefficient by increasing the maximum stress level a chemical component could withstand. Quantitatively, the effect was not as great. That is, whereas increasing the friction coefficient of cratonic components by only a factor of 2 led to pronounced changes for craton stability (Figure 8a), a greater increase in the cohesion term was required. Since the friction coefficient sets the slope of the yielding curve, increasing it by some increment has a greater effect on the maximum yield stress at depth.

[36] The effects of varying the postyield weakening factor are shown in Figure 8b. For the simulations represented the weakening factor was adjusted for all chemical components in unison. An enhanced postyield weakening factor produces greater spatial focusing of shear bands which also become more stable over time. More specifically, once failed regions formed at the edge of a craton they tended to be longer lived at the edge as opposed to migrating deeper into the craton. The stabilization effect was not large and we are limited as to how much postyield weakening we can allow for as values beyond those used would focus failure zones to the point that we could not resolve them numerically. Figure 8b does, however, suggest the possibility that a more pronounced history-dependent effect, which would tend to keep once failed regions weak for a longer time, might act to enhance craton stability. This could provide a means of isolating cratonic interiors from major regions of mantle downwelling which, as noted in section 1, is a potential means of stabilization that is not solely tied to the properties of cratonic lithosphere itself. That is, if preexisting fault zones remained weak to the point that they could be preferentially reactivated, then subduction zones would tend to form and remain locked onto these zones of long-lived weakness as opposed to forming within, or migrating into, lithosphere free of such nucleation points.

[37] The final statement above is speculative at this stage and further exploration requires that a true history dependence be built into our simulations so that failed regions remain weak even after deformation ceases. We have not done this at present. We can, however, explore a related effect that can also keep cratons buffered from regions of subduction. We do so in section 3.5 which considers the potential role of preexisting broad zones of chemically determined weakness within continents.

3.5. Mobile Belt Parameter Study

[38] Geologists have long known that long-lived lateral strength heterogeneities within continents can have a profound influence on patterns of tectonic deformation [e.g., *Smith and Mosley*, 1993]. Cratonic lithosphere is seen by most workers as anonymously strong relative to reference mantle which is why most proposed mechanisms for craton stability have focused on the properties of cratonic lithosphere itself. What has been less appreciated, in discussions of craton stability and root longevity, is the fact that the pervasively deformed continental regions that border many cratons may well be anonymously weak [e.g., *Ring*, 1994]. These mobile belt regions have experienced episodes of tectonic activation and reactivation over their lifetimes suggesting that their inherited weakness may also be relatively long-lived. The influence of long-lived continental weak zones on rifting has been addressed [*Vink et al.*, 1984; *Dunbar and Sawyer*, 1989; *Vauchez et al.*, 1997], and the role of long-lived weak zones for issues of global mantle convection is coming to be appreciated [e.g., *Gurnis et al.*, 2000]. The role of long-lived weak zones for issues of craton stabilization and root longevity has, however, received little if any attention. This motivates the simulations of this section.

[39] To model the effects of peripheral weak zones on craton stability, we must introduce added chemical components to our simulations. These new components will represent mobile belt crust and subcrustal mantle lithosphere. To limit parameter space, we will consider these components to be equivalent to their cratonic counterparts in terms of density and initial thickness but to differ in terms of rheologic parameters. The previous sections showed that of the available rheologic parameters, variations of the effective friction coefficient had the greatest relative effect on craton stability and root longevity. Given this, we will focus on the effects of varying the effective friction coefficient for mobile belt, relative to cratonic and reference mantle, components.

[40] In the mobile belt simulations, the friction coefficient of bulk mantle and cratonic components is set to the nominal value of 0.1. The effective friction coefficient of mobile belt components is varied between simulations to isolate the effects of weak mobile belts. An analogy exists between the weak mobile belts of our simulations and the crumple zones of an automobile which buffer its cab and passengers from collisional forces. Thus, for brevity, we will refer to the ratio of the friction coefficients between mobile belt and cratonic components as a crumple/craton yield ratio. A value of less than one will, in effect, provide cratons with peripheral crumple zones that can fail at relatively low stress.

[41] In the simulation shown in Figure 9, lithospheric failure again allows subduction to initiates at the left continental margin. The positive chemical buoyancy of mobile belt lithosphere cannot overcome the negative thermal buoyancy of the subducting slab and portions of the continental lithosphere are recycled at the subduction site just as they were for the nominal parameter case simulation (Figure 1). The mobile belt simulation differs from the nominal case in that continental recycling, driven by subduction, does not proceed as deeply into the continent. Rather, recycling proceeds only until the locus of subduction comes into contact with the mobile belt/craton boundary (Figure 9). Stresses are then concentrated at the right continental margin, and as the simulation was allowed to proceed, the right margin became the site of a new subduction zone.

[42] The driving force in the simulations of Figures 1 and 9 is the negative thermal buoyancy of cold, convectively unstable mantle, i.e., slabs. The simulations are also similar in that the maximum stress levels the lithosphere can withstand are limited by the yield stress of any lithospheric section. The key difference is that preexisting, lateral yield stress variations exist within the simulation of Figure 9. In the simulation of Figure 1, oceanic and continental lithosphere experienced the same stress levels as they came into contact with a subducting slab. The relatively low friction

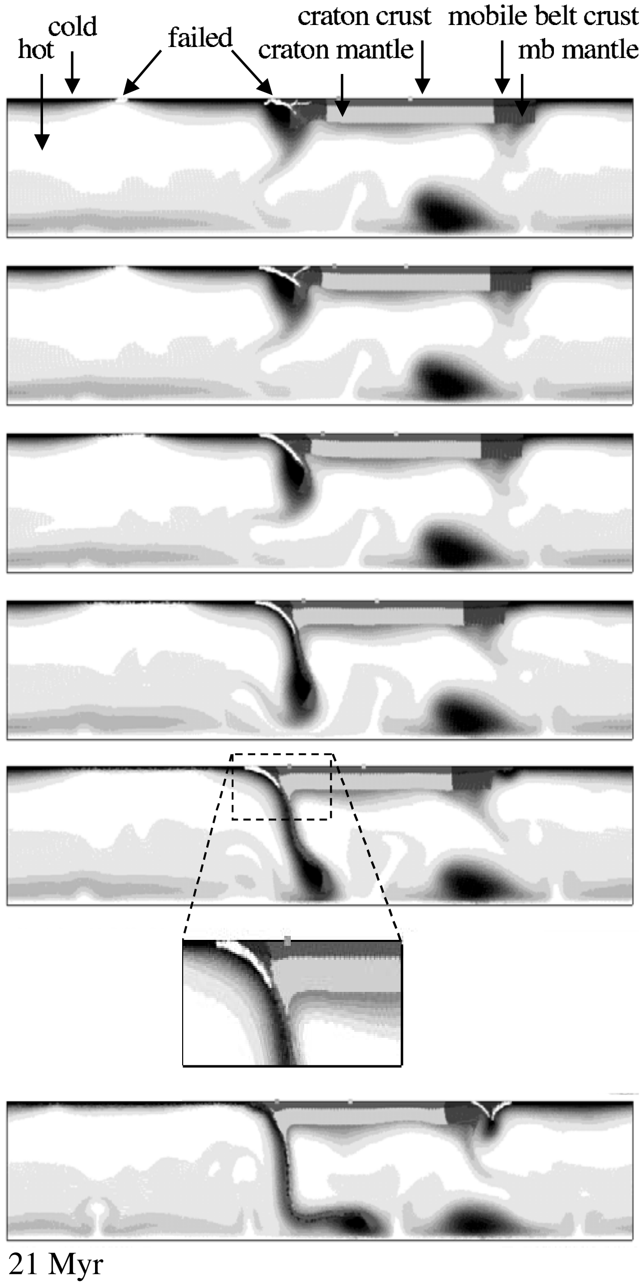

Figure 9. Crumple zone model evolution.

coefficient of mobile belt components in the simulation of Figure 9 means that the maximum stress that the mobile belt lithosphere can maintain is lower than the stress required to cause failure in the cratonic lithosphere. Although cratonic lithosphere does feel the influence of the sinking slab when it is near a subduction zone, the fact that slab induced stresses are transmitted to it through the weak mobile belt means that it does not experience stress levels in excess of its yield stress. Thus the failure surfaces associated with lithospheric recycling do not form within the cratonic lithosphere over the evolution time shown in Figure 9.

[43] The effects of mobile belts are further quantified in Figure 10. Figure 10a shows that root longevity increases dramatically once a critical mobile belt to craton yield ratio is exceeded. Figure 10b shows that increasing the width of mobile belts can also add to root longevity by removing cratons from subduction induced stresses for longer time periods. Over longer timescales than those shown, cratons did rift which caused portions to be exposed to oceanic mantle. Once subduction initiated at these newly formed continental margins cratonic lithosphere was relatively easily recycled as it now had no peripheral crumple zone to buffer it. This suggests that either a mechanism for buffer zone regeneration needs to be included into the simulations if they are to preserve cratonic lithosphere over a timescale of 10^9 years or other craton stabilizing effects must operate mutually with the crumple zone effect. Section 3.6. explores the mutual operation of several stabilization mechanisms acting in tandem.

3.6. Mixed Parameter Effects

[44] A number of simulations have been performed that combine several or all of the parameter effects discussed

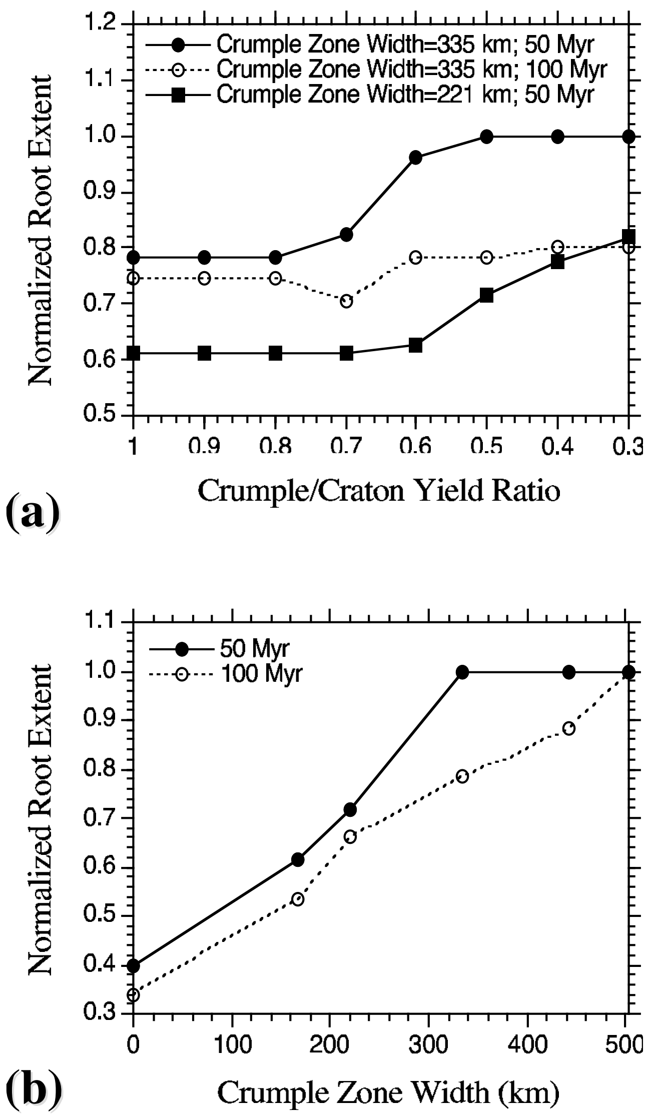

Figure 10. (a) Normalized root extent versus crumple zone to mantle and craton yield stress ratio. (b) Normalized root extent versus crumple zone width.

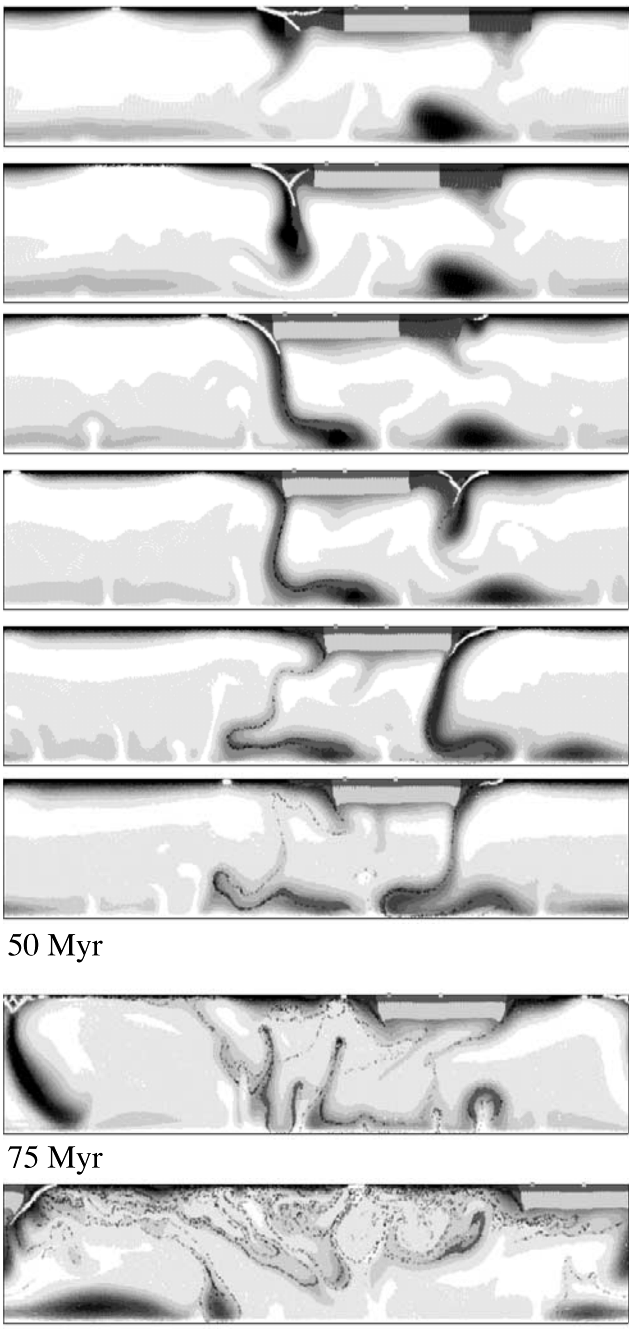

Figure 11. Mixed model evolution.

above. Figure 11 shows the evolution of a mixed effect model that does a particularly good job in preserving deep cratonic lithosphere and maintaining tectonic stability of the crust above. The simulation does not involve extreme values of any one parameter. The reference density of root material is within the range of geochemical estimates [*Poudjom Djomani et al.*, 2001]. The intrinsic viscosity of root material is only a factor of 10 greater than reference mantle which is within the range of rheologic estimates [*Hirth and Kohlstedt*, 1996]. The effective friction coefficients of cratonic and mobile belt components are a factor of 2 greater and a factor of 2 less than, respectively, reference mantle. These variations can be accounted for by pore fluid pressure variations [e.g., *Hubbert and Rubey*, 1959].

[45] The simulation of Figure 11 spans several mantle overturn times. Mobile belt mantle lithosphere is almost completely recycled into the deeper mantle save for small portions that remain attached to the craton peripheries. Sections of the mobile belt crust are also rifted from the craton but significant portions do also remain attached to the craton peripheries as well. Thus, at the final evolution time shown the craton does maintain peripheral weak zones. To explore whether the weak zones, for this mixed effect simulation, did have a nontrivial role we ran a similar simulation that started with a craton exposed to a subduction zone. In this case, the cratonic crust was deformed and portions of the deeper cratonic lithosphere where recycled in a mode similar to that of Figure 1.

[46] We also ran additional variations for the mixed effect case of Figure 11 to further gauge the relative roles of specific parameter effects. The dominant effect was the relatively high effective friction coefficient of cratonic crust and mantle lithosphere. Embedding the craton deeper into a continent could add to cratonic stability and longevity as could providing cratonic roots with a higher viscosity and/or buoyancy but these effects alone or in tandem could not fully provide for stability and longevity within the parameter ranges explored within the previous sections for any effect of its own. Allowing the effective friction coefficient of both cratonic crust and mantle lithosphere to be a factor of 4–5 greater than reference mantle could, on the other hand, provide for craton stability and longevity independent of the other parameter effects. The other parameter effects could lower the relative effective friction coefficient increase required for cratonic components, but as already noted, a higher friction coefficient in the cratonic crust and uppermost mantle was always needed.

4. Conclusions and Discussion

[47] Our first main conclusion, based on our simulations, is that the chemical buoyancy of cratonic roots is likely not the key factor that provides for the stability and longevity of cratonic lithosphere. Even if root material is chemically buoyant relative to reference mantle, this cannot preserve large portions of it from being recycled in subduction zone settings. It cannot prevent a root from being rifted into smaller and smaller pieces nor can it prevent root material from detaching from the crust above. This contrasts with isostatic arguments which have been used to suggest that root buoyancy can provide for stability and longevity of cratonic lithosphere [e.g., *Poudjom Djomani et al.*, 2001; *Lee et al.*, 2001]. Such arguments compare the integrated density of a 1-D column of cratonic lithosphere to that of a reference asthenospheric column and inherently assume that a cratonic column remains coherent, that lateral cratonic extent does not have a large effect, and that the sinking of a cratonic column into the mantle under its own weight is the principal potential means of root recycling. The simulations of this paper show that accounting for the formation of lithospheric detachments, for the effects of lateral root

extent, and for subduction driven recycling significantly decreases the survival potential of cratonic roots relative to what would be inferred from isostatic arguments. The key point is that it is subduction which recycles the root in our simulations. The simulations do not challenge the idea that the lower density of chemically distinct cratonic lithosphere, relative to the bulk mantle, can prevent it from sinking into the mantle under its own weight via a Rayleigh-Taylor type instability [e.g., *Poudjom Djomani et al.*, 2001; *Lee et al.*, 2001].

[48] Our next main conclusion relates to root viscosity. Dry conditions within deep cratonic lithosphere, relative to reference mantle, can increase the viscosity of cratonic roots [*Pollack*, 1986; *Hirth and Kohlstedt*, 1996]. However, this does not guarantee root stability and longevity. Just as a high ductile viscosity for oceanic lithosphere can not prevent it from being recycled if brittle behavior is considered [*Moresi and Solomatov*, 1998], so too the brittle properties of cratonic lithosphere could potentially offset the effects of an intrinsic increase of deep root viscosity. Although previous dynamic modeling studies have concluded that a high root viscosity can provide stability and longevity for cratonic lithosphere [*Shapiro*, 1995; *Doin et al.*, 1997; *Lenardic and Moresi*, 1999], these studies did not consider brittle failure.

[49] Not including brittle failure as a strength-limiting process can lead to a drastic overestimation of lithospheric strength [e.g., *Sawyer*, 1985]. On the basis of previous dynamic models that do not consider brittle failure within the cratonic lithosphere [*Shapiro*, 1995; *Doin et al.*, 1997] and on scaling results that also assume purely viscous behavior [*Lenardic and Moresi*, 1999], the viscosity assigned to deep cratonic lithosphere in the simulation of Figure 4a, for example, should be more than adequate to provide for stability and longevity. The reason this does not prove to be the case within the simulation itself is because brittle failure is allowed for. Our simulations suggest that very viscous root material can be recycled unless the thickness of chemically distinct, high-viscosity cratonic lithosphere exceeds that of the oceanic lithosphere by a factor of 2 or more. Thus root viscosity may be the key to cratonic stability and longevity, but if this is the case, then a critical root thickness must be exceeded. In tandem, these conditions imply that if root viscosity is the key, then the effective mechanical thickness of cratonic lithosphere should approach 100 km or more. This then is our second conclusion.

[50] Our third main conclusion is that the most effective means of providing for cratonic stability and longevity, within our simulations, is if both cratonic crust and mantle lithosphere have a brittle yield stress that is greater than that of the oceanic lithosphere. If the yield properties are parameterized in terms of an effective friction coefficient, then the value for cratonic components needs to be approximately of a factor of 4 greater if this is the sole craton stabilizing effect. The implied effective friction coefficient for cratonic components is in accord with laboratory results for dry rock friction [*Byerlee*, 1968; *Kohlstedt et al.*, 1995]. The value applied within our simulations for oceanic lithosphere is, on the other hand, low relative to dry rock friction. The low value is required to allow for the subduction of oceanic lithosphere in our simulations. This is in accord with previous modeling studies of oceanic lithospheric subduction and the low effective friction coefficient values have been attributed to hydration effects [*Bird*, 1978; *Bercovici*, 1996; *Moresi and Solomatov*, 1998; *Tackley*, 1998]. Thus, if the stability mode above is correct, it suggests that dry conditions within cratons allow them to withstand relatively high stresses before failing in a brittle mode while both oceanic lithosphere and deforming continental lithosphere contain sufficient quantities of water to allow them to fail at stresses lower than would be predicted based on assuming a dry rock friction coefficient.

[51] Our fourth main conclusion is that the ratio of cratonic to oceanic yield stress values required for craton stability and deep lithospheric longevity can be lowered by a factor of 2, relative to that noted above, if cratons are buffered by relatively low yield stress mobile belts. Such a peripheral mobile belt can provide added stability to the cratonic interior by absorbing the stresses associated with an active margin. This is analogous to the crumple zone of an automobile or, to use a thermal analogy, to ablative heat shields which can keep the interior of a spacecraft cool not by offering high thermal resistance but by burning away on reentry and thus carrying heat away from the craft. These analogies make it clear that if mobile belts are to protect cratons through more than a single potential deformation episode, then they must be renewable. At present, the generally accepted view of continental growth involves island arcs colliding with existing continental cores. It is likely that arc lithosphere is hydrated, given it is formed in a subduction setting, and therefore weak relative to the continental core. If this picture of continental growth is correct, then it suggests a means of regenerating weak buffer zones at the periphery of continental interiors. To fully explore the implications this would have for long-term cratonic stability requires that we incorporate chemical differentiation into our numerical simulations, and this is under way.

[52] In summary, our simulations reveal two end-member modes of stability for cratonic crust and for longevity of deeper roots, one in which strength principally resides at depth in a thick, strong continental root, and one in which inherently stronger cratonic upper and lower crust yields at higher stresses than the crust in tectonically mobile regions. These modes have distinguishable dynamic consequences. In regions where stability originates in deep roots, material which makes up the root should not be younger than the stable crust. If the crust is strong, then it might be possible, using xenolith mapping of the deep lithosphere, to find regions where the root has been disrupted by mantle processes without observable surface deformation [e.g., *Jacob et al.*, 2000]. Further, the strength in the uppermost lithosphere can be measured in terms of long-term elastic response as a function of wavelength [e.g., *Zuber et al.*, 1989; *Simons et al.*, 2000] and can also be inferred from the thickness of the seismogenic layer [*Maggi et al.*, 2000]. Such data-based inference of cratonic strength could potentially discriminate between the two stability modes mapped out by our simulations. It should be noted, however, that these two modes need not be mutually exclusive nor must their relative contributions be equal for cratonic lithosphere worldwide. A careful examination of data from specific

cratonic regions, together with comparisons to simulation predictions, will be required to show if one mode dominates or if it is a mixture that contributes to the stability of any given craton.

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