The dependence of planetary tectonics on mantle thermal state: Applications to early Earth evolution

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Abstract

For plate tectonics to operate on a planet, mantle convective forces must be capable of forming weak, localized shear zones in the lithosphere that act as plate boundaries. Otherwise, a planet's mantle will convect in a stagnant lid regime, where subduction and plate motions are absent. Thus, when and how plate tectonics initiated on Earth is intrinsically tied to the ability of mantle convection to form plate boundaries; however, the physics behind this process are still uncertain. Most mantle convection models have employed a simple pseudoplastic model of the lithosphere, where the lithosphere "fails" and develops a mobile lid when stresses in the lithosphere reach the prescribed yield stress. With pseudoplasticity high mantle temperatures and high rates of internal heating, conditions relevant for the early Earth, impede plate boundary formation by decreasing lithospheric stresses, and hence favor a stagnant lid for the early Earth. However, when a model for shear zone formation based on grain size reduction is used, early Earth thermal conditions do not favor a stagnant lid. While lithosphere stress drops with increasing mantle temperature or heat production rate, the deformational work, which drives grain size reduction, increases. Thus the ability of convection to form weak plate boundaries is not impeded by early Earth thermal conditions. However, mantle thermal state does change the style of subduction and lithosphere mobility; high mantle temperatures lead to a more sluggish, drip-like style of subduction. This "sluggish lid" convection may be able to explain many of the key observations of early Earth crust formation processes preserved in the geologic record. Moreover, this work highlights the importance of understanding the microphysics of plate boundary formation for assessing early Earth tectonics, as different plate boundary formation mechanisms are influenced by mantle thermal state in fundamentally different ways.

1 Introduction

Plate tectonics has fundamentally shaped the tectonic, thermal, and chemical evolution of the Earth, leaving our planet in sharp contrast to its neighbors in the solar system. However, despite the importance of plate tectonics for Earth's history, when and how plate tectonics started on Earth is poorly constrained [e.g. *Condie and Kröner*, 2008; *van Hunen and Moyen*, 2012; *Cawood et al.*, 2013; *Korenaga*, 2013]. Few rocks from the Archean survive today, and those that exist are difficult to interpret. For even earlier times, greater than 4 billion years ago, zircon grains are all that is available to constrain the dynamics of the very early Earth [*Valley et al.*, 2002; *Harrison et al.*, 2005; *Kemp et al.*, 2010]. Moreover, exactly what evidence would constitute clear proof for plate tectonics is still debated, as rocks formed by plate tectonic processes on the modern Earth can potentially be formed by non-plate-tectonic processes on the early Earth [*Bédard*, 2006; *Johnson et al.*, 2014; *Sizova et al.*, 2015], among other ambiguities. As a result, estimates for when plate tectonics began based on the geologic record span from > 4.0 Ga [*Harrison et al.*, 2005; *Hopkins et al.*, 2008] to < 1.0 Ga [*Stern*, 2005].

Determining how and when plate tectonics began on Earth is not only an important question for understanding the history of our own planet, but also for the more general question of what factors allow for the development of plate tectonics on any rocky planet. Earth is the only rocky planet or moon known to operate in a plate-tectonic regime; the other rocky bodies in the solar system are thought to operate in some form of stagnant lid regime [e.g. *Breuer and Moore*, 2007]. Interestingly, the icy satellite Europa potentially exhibits subduction, and possibly even full fledged plate tectonics, in its outer icy shell [*Kattenhorn and Prockter*, 2014]. However, more work is needed to determine whether potential subduction zones are spatially limited, akin to Venus as discussed below, or more widespread. In the stagnant lid regime, the surface exists as a single plate with mantle convection taking place beneath this plate, and there is no global network of subduction zones and mid-ocean ridges, or large-scale differential movement between different regions of the lithosphere [e.g. *Ogawa et al.*, 1991; *Solomatov*, 1995]. Stagnant lid planets or satellites can still display volcanism and active tectonics, as seen on Mars [e.g. *Neukum et al.*, 2004; *Robbins et al.*, 2011], Mercury [e.g. *Byrne et al.*, 2014], or Io [e.g. *O'Reilly and Davies*, 1981; *Moore*, 2003]. In fact, Venus even shows evidence for limited subduction in some regions, possibly due to burial of the lithosphere by volcanism, and subsequent "peeling" away and sinking of this buried lithosphere [e.g. *Sandwell and Schubert*, 1992; *Davaille et al.*, 2017], a process called "plume-induced" subduction [*Gerya et al.*, 2015]. However, all of these planets and satellites clearly lack the global system of mobile plates and plate boundaries that characterizes plate tectonics on Earth.

Plate tectonics is also thought to play an important role in Earth's ability to maintain a surface environment suitable for life [e.g. *Kasting and Catling*, 2003; *Foley and Driscoll*, 2016]. Plate tectonics helps power the geodynamo through efficient cooling of the mantle and core [*Nimmo*, 2002; *Driscoll and Bercovici*, 2014], and the resulting magnetic field helps to shield the planet from harmful radiation [e.g. *Grießmeier et al.*, 2005]. Moreover, plate tectonics facilitates carbon cycling between the surface and interior, which helps to maintain temperate surface temperatures on Earth thanks to a feedback between climate, silicate weathering, orogeny, and volcanism [*Walker et al.*, 1981; *Tajika and Matsui*, 1992; *Franck et al.*, 1999; *Sleep and Zahnle*, 2001; *Berner*, 2004; *Abbot et al.*, 2012; *Foley*, 2015]. Stagnant lid planets may also be capable of establishing a carbon cycle that regulates climate [e.g. *Lenardic et al.*, 2016; *Tosi et al.*, 2017; *Noack et al.*, 2017; *Foley and Smye*, 2018], though current estimates still find that plate tectonics is the more favorable tectonic state for long-term climate stability [*Foley and Smye*, 2018]. With the ongoing boom in exoplanet discoveries now revealing that rocky planets are common in the galaxy [*Batalha*, 2014], understanding the dynamical causes of plate tectonics has become an increasingly important goal in astrobiology, in addition to earth science [e.g. *Valencia et al.*, 2007; *O'Neill and Lenardic*, 2007; *Kite et al.*, 2009; *Valencia and O'Connell*, 2009; *Korenaga*, 2010b; *Stamenkovic et al.*, 2011; *van Heck and Tackley*, 2011; *Foley et al.*, 2012; *Lenardic and Crowley*, 2012; *Stein et al.*, 2013].

While continued work in unraveling the Earth's early geologic record is essential for answering the question of when plate tectonics began, theoretical studies on early Earth mantle dynamics provide a key additional constraint, by demonstrating what styles of mantle convection are geodynamically plausible. Theoretical studies can also help in interpreting geologic observations by, for example, testing the feasibility of different processes for generating Archean crust. Moreover, theoretical models can also be readily extended to Venus, Mars, or Mercury, providing additional tests on theories behind the origin of plate tectonics on Earth. That is, any theory must be able to explain the presence of plate tectonics on Earth and its absence on the other rocky planets in our solar system. In this paper I focus on the rheological mechanisms necessary for generating plate like mantle convection, and the constraints we can place on early Earth (specifically the Hadean and Archean) tectonics from our understanding of these mechanisms. In particular, I highlight that attempting to model the microphysics of plate boundary formation, in this case by modeling grain size evolution, leads to new predictions about early Earth tectonics in comparison to previous geodynamic models that treat the lithosphere as a plastic material. The differences in the model results stem from fundamental differences in the physics underlying the two plate generation mechanisms. As a result, in order to truly understand how and when plate tectonics began on Earth, and what factors are necessary for plate tectonics to begin on any rocky planet, we must understand the microphysics of shear localization and plate boundary formation.

1.1 Generating plate tectonics from mantle convection

The geodynamics behind the stagnant lid regime seen on planets such as Mars and Mercury has been well studied, and can be simply understood as a result of the strong temperature dependence of mantle rheology. Temperature dependent viscosity creates a strong, rigid lithosphere and thus confines convection to the warmer mantle interior where viscosity is lower [e.g. *Ogawa et al.*, 1991; *Solomatov*, 1995]. Plate tectonics then requires that some additional mechanism, that allows mantle convective forces to form narrow weak zones within the high viscosity lithosphere, such that plates can slide past each other and subduct beneath one another, be present. A strongly non-linear rheology, where regions of high stress or deformation become weaker than areas of low stress or deformation, has proven successful in forming plate boundaries and generating a "mobile lid" form of convection [e.g. *Weinstein and Olson*, 1992; *Moresi and Solomatov*, 1998; *Tackley*, 2000]. In this study I define the mobile lid regime as one where subduction of surface material into the mantle occurs and drives lithospheric mobility. Mobility of the lithosphere may be slow compared to typical fluid velocities in the mantle interior (a "sluggish lid" regime), exhibit varying degrees of episodicity, or fail to produce truly rigid plates. But as long as subduction of surface material that drives surface motion takes place, convection is in the mobile lid regime. In fact, forming genuinely plate-like behavior is still a challenge today, as models typically fail to form strike-slip faults, truly rigid plates, or single-sided subduction [*Bercovici et al.*, 2015], though some of these features can be created in select cases [e.g. *Gerya*, 2010; *Crameri et al.*, 2012]. As a result, I will focus on the "mobile lid" regime (rather than the "plate-tectonic" regime) in contrast to the stagnant lid regime in this paper.

Although non-linear rheologies can generate mobile lid convection, the mechanism (or mechanisms) behind such rheological behavior are still not well understood [e.g. *Bercovici et al.*, 2015]. One commonly used mechanism is the pseudoplastic yield stress rheology [e.g. *Fowler*, 1993; *Moresi and Solomatov*, 1998; *Tackley*, 2000; *Stein et al.*, 2004; *van Heck and Tackley*, 2008; *Lowman*, 2011]. Here the lithosphere is assumed to have a finite strength, the yield stress, and fails when this stress is reached; failure of the lithosphere results in the formation of weak shear zones and mobility of the lithosphere. With the pseudoplastic rheology, the stress state in the lithosphere is critical for whether convection sits in a mobile lid or stagnant lid regime, as stresses in the lithosphere must reach the yield stress for weak plate boundaries to form.

While the pseudoplastic rheology is capable of producing plate-like mantle convection, yield stress values far below that which are inferred from laboratory experiments [e.g. *Mei et al.*, 2010] or observations of lithospheric flexure [e.g *Zhong and Watts*, 2013], are typically needed to produce mobile lid convection [e.g. *Moresi and Solomatov*, 1998; *Tackley*, 2000; *van Heck and Tackley*, 2008; *Solomatov*, 2004; *Korenaga*, 2010a]. It is thus unclear whether pseudoplasticity is capable of explaining the presence of plate tectonics on Earth; that is, additional weakening mechanisms may be needed. One possible explanation is that hydration of old lithosphere could weaken it, and make the effective strength low enough for yielding at subduction zones [*Korenaga*, 2007]. It has also been proposed that channelized flow in the asthenosphere enhances stresses, and may allow lithospheric yielding even without any weakening via hydration [*Höink et al.*, 2012]. However, more work is needed to confirm whether these hypotheses can fully reconcile the pseudoplastic yielding mechanism for generating plate tectonics with experimental and observational constraints of Earth's lithospheric strength.

Nevertheless the pseudoplastic rheology has been used in previous studies on early Earth tectonics. A key feature of the Hadean and Archean Earth is that the thermal state of the mantle was likely significantly different than at the present. Rates of radiogenic heat production were higher [e.g. *Turcotte and Schubert*, 2002; *Korenaga*, 2006] and the mantle interior was likely hotter as well. Mantle potential temperatures could have been as high as ~ 2000 K just after planetary accretion ended, and the putative magma ocean solidified [e.g. *Abe*, 1997; *Solomatov*, 2000], though there is little physical evidence of the mantle thermal state at this early stage of Earth's history. In the Archean, petrological estimates indicate the mantle was $\sim 100 - 200$ K hotter than the present day potential temperature of $\approx 1350°$ C, and has gradually cooled since the Archean [e.g. *Herzberg et al.*, 2010; *Keller and Schoene*, 2018]. However, high temperatures, similar to those inferred for the Archean, may still prevail in some parts of the lower mantle, as evidenced by rocks, thought to be sourced from a deep seated mantle plume, recording temperatures as high as $\sim 1700^\circ$ C [*Trela et al.*, 2017].

These generally warmer thermal conditions expected in the Hadean and Archean mantle can change the lithospheric stress state, and thus influence the early Earth's propensity for plate tectonics, at least as it would be generated with pseudoplastic yielding. In particular, many studies have found that higher mantle temperatures decrease the magnitude of stresses in the lithosphere, thus requiring lower yield stress values for mobile lid convection; that is, stagnant lid convection becomes more likely with increasing mantle temperature [e.g. *Stein et al.*, 2004; *O'Neill et al.*, 2007]. Moreover, the relative amounts of internal versus basal heating plays an important role as well, as basal heating has been found to feature higher lithospheric stresses than internally heated convection [e.g. *O'Neill et al.*, 2016; *Weller and Lenardic*, 2016; *Korenaga*, 2017].

However, different mechanisms for generating plate tectonics from mantle convection have been proposed, and such mechanisms may respond differently to changes in mantle thermal state. In particular, I focus here on grain size reduction as it is commonly seen in exhumed lithospheric shear zones [*White et al.*, 1980; *Drury et al.*, 1991; *Jin et al.*, 1998; *Warren and Hirth*, 2006; *Skemer et al.*, 2010], and viscosity decreases with decreasing grain size when deformation is accommodated via diffusion creep or grain boundary sliding [e.g. *Hirth and Kohlstedt*, 2003]. Whether grain size reduction is a viable plate generation mechanism has been questioned, however, because grain size reduction by dynamic recrystallization takes place in the dislocation creep regime, while grain size sensitive flow only occurs in the diffusion creep or grain boundary sliding regimes [e.g. *Etheridge and Wilkie*, 1979; *De Bresser et al.*, 1998;

Karato and Wu, 1993]. If grain size reduction cannot continue once deformation has entered a grain size sensitive flow regime, only modest weakening can occur [e.g. *De Bresser et al.*, 2001; *Montési*, 2013]. However, grain size reduction can continue within the diffusion creep regime when a secondary phase (e.g. pyroxene) is dispersed throughout the primary phase (e.g. olivine), because of the combined effects of damage to the interface between phases and Zener pinning (where the secondary phase blocks grain growth of the primary phase). Warping of the interface between phases (or interface damage) itself drives grain size reduction, and such warping can occur via any solid state creep mechanism [*Bercovici and Ricard*, 2012]. Recent experimental studies confirm that the presence of secondary phases significantly enhances grain size reduction, especially as phases begin to mix at high strain [*Bercovici and Skemer*, 2017; *Cross and Skemer*, 2017; *Tasaka et al.*, 2017a, b].

Grain size reduction has been shown to be an effective mechanism for generating mobile lid convection on Earth, and explaining its absence on the other terrestrial planets of our solar system [*Landuyt and Bercovici*, 2009; *Foley et al.*, 2012]. However it should be noted that plate tectonics on Earth can only be explained with grain size reduction for certain ranges of the key model parameters (specifically the damage partitioning fraction and grain growth activation energy, defined below in §2.1), and that these parameters are not well constrained experimentally. As a result, further work constraining these parameters, as well as the theoretical formulation for grain size evolution used in this and similar studies, is needed to test the viability of grain size reduction as a mechanism for explaining the operation of plate tectonics on Earth.

In this paper I show that a grain size reduction mechanism for generating plate boundaries responds to changes in mantle temperature in a fundamentally different way than pseudoplastic yielding. Specifically, new mantle convection calculations are presented that demonstrate that deformational work, which drives grain size reduction in the lithosphere, increases with increasing internal heating rate, and consequently increasing mantle temperature. Thus the situation is opposite that of the pseudoplastic rheology, where higher internal heating rates or mantle temperatures lead to lower stresses, thereby favoring a stagnant lid. As a result, a transition to the stagnant lid regime at high mantle temperatures or high internal heating rates is not seen when grain size reduction is considered. However, subduction and lithospheric mobility does become more sluggish, and subduction becomes more episodic and drip-like. These changes in subduction style have potentially important implications for interpreting early Earth geological and geochemical observations, as discussed in §4.2. The paper is structured as follows: the numerical model setup and governing equations used in this study are described in §2; numerical results and scaling analyses are presented in §3; the implications of the results are described in §4; and the main conclusions and future directions needed to making progress on answering the question of when plate tectonics started are summarized in §5 & §6, respectively.

2 Background Theory and Model Setup

2.1 Grain damage theory

I use a formulation for grain size evolution referred to as grain damage; it is based on energetics, as reducing grain size in a volume of rock is equivalent to increasing the surface energy in that volume [e.g. *Bercovici et al.*, 2001a, b; *Bercovici and Karato*, 2003; *Austin and Evans*, 2007; *Landuyt et al.*, 2008; *Bercovici and Ricard*, 2012, 2013, 2014; *Foley and Bercovici*, 2014; *Foley et al.*, 2014]. The grain damage formulation used in this study is a simplified version of the theory for polymineerallic rocks from *Bercovici and Ricard* [2012], which provides equations for the evolution of grain size in the different mineral phases and for the curvature of the interface between phases. *Bercovici and Ricard* [2012] shows that during deformation of a polymineerallic aggregate, a "pinned state" develops, where secondary phases dispersed through the primary phase "pin" grain boundaries, thereby slowing grain growth. Furthermore, in the pinned state the curvature of the interface between phases, or the interface roughness, controls the average grain size of the mineral phases; i.e. grain size becomes proportional to interface roughness in the pinned state. Assuming the pinned state prevails throughout the mantle thus simplifies the model, as the average grain size of a rock volume can now be tracked by one equation, rather than solving for both mineral phases and the interface roughness separately. That is, one can use the equation governing the interface roughness (equation 4d of *Bercovici and Ricard* [2012]) to solve for the grain size directly. I therefore assume the pinned state prevails throughout the mantle in this study.

An additional simplifying assumption I employ is that diffusion creep is the dominant deformation mechanism throughout the entire mantle, and dislocation creep, or any other deformation mechanism, can be neglected. Neglecting dislocation creep can potentially impact the results, as in reality the rheology will be controlled by whichever mechanism, diffusion creep or dislocation creep, can accommodate deformation the easiest [e.g. *Rozel et al.*, 2011]. In particular, the typical grain size in the mantle interior could grow large enough at high mantle temperatures for dislocation creep to become the preferred deformation mechanism in this region. However, the results are unlikely to be significantly impacted by neglecting dislocation creep, as lid mobility is controlled by the effective rheology of shear zones in the lithosphere, where grains are small and diffusion creep dominates [*Foley and Bercovici*, 2014]. Moreover, in the numerical models grain sizes in the mantle interior remain relatively small, typically lower than a few millimeters, even with high internal heating rates and thus high interior mantle temperatures (see §3.3). Dislocation creep would also cause viscous weakening in high stress regions, such as around downgoing slabs or beneath surface plates [e.g. *Jadamec and Billen*, 2010]. However, damage also leads to weakening in such locations (e.g. see §3.1), so this effect of neglecting dislocation creep is also unlikely to significantly influence the results. I also note that in the shallow lithosphere, additional creep mechanisms, such as low temperature plasticity, can become important [*Karato*, 2008; *Kohlstedt and Mackwell*, 2010; *Mei et al.*, 2010; *Thielmann*, 2017]. However, as explained above, diffusion creep is the dominant deformation mechanism in shear zones that have undergone significant grain size reduction. It is thus unclear if including low temperature plasticity would significantly affect plate generation with grain damage.

Assuming the pinned state prevails also means that mineral phases are assumed to always be well mixed, when in reality mixing occurs only after extensive deformation [e.g. *Bercovici and Skemer*, 2017; *Cross and Skemer*, 2017; *Tasaka et al.*, 2017a, b]. Phase mixing is more efficient at small grain sizes, and thus damage also becomes more effective at small grain sizes due to pinning and interface damage. As a result, grain size itself influences the rate of grain size reduction, an effect that is not captured in the models presented here. Including the effect of phase mixing on grain size evolution is an important goal for future work, as it may improve the plate-like nature of mobile lid convection produced via grain damage, by enhancing shear localization [*Bercovici and Ricard*, 2016].

The above assumptions result in the following theoretical formulation of grain damage used in this study. The mantle viscosity, μ, is a function of grain size and temperature as expected for diffusion creep [e.g. *Hirth and Kohlstedt*, 2003]:

$$\mu = \mu_n \exp\left(\frac{E_v}{RT}\right)\left(\frac{A}{A_0}\right)^{-m}, \tag{1}$$

where μ_n is a constant, E_v is the activation energy for diffusion creep ($E_v = 300$ kJ mol^{-1} [*Karato and Wu*, 1993]), R is the universal gas constant, T is temperature, A is fineness or inverse grain size, A_0 is a reference fineness, and m is a constant. A grain size sensitivity exponent of $m = 2$ is used, the expected value for Nabarro-Herring creep, or diffusion through the grain [e.g. *Hirth and Kohlstedt*, 2003].

The evolution of fineness follows

$$\frac{\mathrm{D}A}{\mathrm{D}t} = \frac{f}{\gamma}\Psi - hA^p, \tag{2}$$

where t is time, f is the damage partitioning fraction, which can vary from zero to one, γ is the surface free energy, Ψ is the rate of deformational work, h is the healing rate, and p is a constant (I use $p = 4$, as estimated for polymineralic rocks by *Bercovici and Ricard* [2012]). Deformational work rate is defined as $\Psi = \nabla \underline{v} : \underline{\underline{\tau}}$, where $\underline{v}$ is velocity and $\underline{\underline{\tau}}$ is the stress tensor. The first term on the right-hand side of (2) states that some fraction, f, of the deformational work partitions into surface energy, thereby reducing the grain size (or increasing fineness). The second term on the right-hand side of (2) represents normal grain growth, which acts to increase grain size (or reduce fineness). The healing rate constant, h, is a function of temperature with an Arrhenius form:

$$h = h_n \exp\left(-\frac{E_h}{RT}\right) \tag{3}$$

where h_n is a constant and E_h is the activation energy for grain growth. The value of E_h for polymineralic rocks is not well known, because most grain growth experiments have been conducted on monomineralic samples. However, *Mulyukova and Bercovici* [2017] found that dynamic recrystallization experiments and observations from natural shear zones, where smaller grain sizes are found at lower temperatures, can be well fit with a grain growth activation energy of $E_h \approx 400$ kJ/mol. Moreover, the damage partitioning fraction, f, could also be temperature-dependent, with larger values of f found at lower temperatures [*Rozel et al.*, 2011; *Mulyukova and Bercovici*, 2017]. However, with $E_h \approx 400$ kJ/mol, f is found to be nearly independent of temperature when grain damage theory is used to fit laboratory deformation experiments [*Mulyukova and Bercovici*, 2017; *Foley*, 2018]; as a result a constant f is used in this study.

The Arrhenius temperature-dependent laws for viscosity and healing, (1) & (3), are simplified in this study by using a Frank-Kamenetskii approximation, which gives the following modified equations for μ and h:

$$\mu = \mu_n \exp\left(\gamma_v T\right) \left(\frac{A}{A_0}\right)^{-m} \tag{4}$$

$$h = h_n \exp\left(-\gamma_h T\right). \tag{5}$$

(4) & (5) are then used, after non-dimensionalization (see §2.2), in the numerical models shown in this study. Here γ_v and γ_h describe the temperature-dependence of viscosity and healing, respectively. When non-dimensionalized these become the Frank-Kamenetskii parameters for viscosity and healing, and can be related to E_v and E_h as described in §2.2. The Frank-Kamenetskii approximation reduces the number of free parameters involved in the viscosity law after non-dimensionalization [see *Korenaga*, 2009], thus simplifying the modeling and analysis. The Frank-Kamenetskii approximation also results in lower viscosities at cold temperatures, i.e. in the lithosphere, than would be reached using the Arrhenius relation given by (1). However, the very large viscosity contrasts produced by the Arrhenius viscosity law with laboratory constrained activation energies are numerically challenging, and therefore

lower activation energies are typically used in modeling studies [e.g. *Tackley*, 2000; *Foley et al.*, 2012; *O'Neill et al.*, 2016]. The value of the Frank-Kamenetskii parameter chosen in this study results in viscosity variation that is consistent with other plate generation studies (see §3.1). A larger viscosity variation across the lithosphere would shift the boundary between stagnant lid and mobile lid convection, shown in Figure 1 below, to higher Frank-Kamenteskii parameters for healing, but is otherwise not expected to change the overall conclusions of this study.

2.2 Governing equations

The grain damage formulation is incorporated into a model of infinite Prandtl number, Boussinesq thermal convection. The damage formulation, (2), is non-dimensionalized using the following scales, where primes denote non-dimensional variables: $\underline{x} = \underline{x}'d$, where d is the depth of the mantle; $t = t'd^2/\kappa$, where κ is the thermal diffusivity; $\underline{v} = \underline{v}'\kappa/d$; $T = T'\Delta T + T_s$, where ΔT is the temperature difference across the mantle and T_s is the surface temperature; $A = A'A_0$; $\underline{\underline{\tau}} = \underline{\underline{\tau}}'\mu_m\kappa/d^2$, where μ_m is the reference viscosity defined at $T_m = \Delta T + T_s$ in the absence of damage (i.e. at $A = A_0$); and $\gamma_v = \theta_v/\Delta T$ and $\gamma_h = \theta_h/\Delta T$. θ_v and θ_h are the Frank-Kamenetskii parameters for viscosity and healing, respectively. The Frank-Kamenteskii parameter is related to the activation energy from the Arrhenius law as

$$\theta = \frac{E\Delta T}{R(T_s + \Delta T)^2}, \tag{6}$$

which gives θ_v for $E = E_v$ and θ_h for $E = E_h$.

Dropping the primes and continuing with non-dimensional variables, the resulting non-dimensional equation governing fineness evolution is:

$$\frac{\mathrm{D}A}{\mathrm{D}t} = D\psi \exp\left(\theta_v(1-T)\right)A^{-m} - H\exp\left(-\theta_h(1-T)\right)A^{p} \tag{7}$$

where $\psi = \nabla \underline{v} : (\nabla \underline{v} + (\nabla \underline{v})^T)$, D is the non-dimensional damage number, and H is the non-dimensional healing number. These quantities are defined as $D = f\mu_m\kappa/(\gamma A_0 d^2)$ and $H = h_m A_0^{(p-1)} d^2/\kappa$ where $h_m = h(T_m)$.

The equations for conservation of mass, momentum, and energy, expressed in terms of non-dimensional variables using the same scales as above, are:

$$\nabla \cdot \underline{v} = 0 \tag{8}$$

$$0 = -\nabla P + \nabla \cdot (2\mu\underline{\underline{\dot{\varepsilon}}}) + Ra_0 T\hat{\underline{z}} \tag{9}$$

$$\frac{\partial T}{\partial t} + \underline{v} \cdot \nabla T = \nabla^2 T + Q \tag{10}$$

where P is the non-hydrostatic pressure, $\dot{\varepsilon_{ij}} = (\partial v_i/\partial x_j + \partial v_j/\partial x_i)/2$ is the strain rate, $\hat{\underline{z}}$ is the unit vector in the vertical direction, and Ra_0 is the reference Rayleigh number defined at the reference interior mantle viscosity and temperature; $Ra_0 = (\rho g\alpha\Delta T d^3)/(\kappa\mu_m)$ where ρ is density, α is thermal expansivity, and g is acceleration due to gravity. The non-dimensional internal heating rate, Q, is defined as $Q = Hd^2/(\Delta T\kappa)$, where $H = H_v/(\rho C_p)$, H_v is the volumetric heating rate (with units of

Wm^{-3}), and C_p is the specific heat. The non-dimensional viscosity, μ, is defined as

$$\mu = \exp\left(\theta_v(1-T)\right)A^{-m}. \tag{11}$$

It is also helpful to define parameters to describe the variation of viscosity and healing across the mantle due to temperature dependence; μ_l/μ_m, the viscosity ratio in the absence of damage, and h_l/h_m, the healing ratio, where the subscript l denotes the value in the lithosphere (i.e. at $T = 0$ for all numerical models). The viscosity and healing ratios are therefore equivalent to the non-dimensional viscosity and healing, respectively, at the non-dimensional surface temperature of $T = 0$.

2.3 Numerical Model Setup

The models use free-slip top and bottom boundary conditions and periodic side boundaries. With periodic side boundaries, a uniform horizontal velocity can be added to the velocity solution while still satisfying the governing equations; in the numerical model this uniform horizontal velocity is set to be zero at each timestep. The temperature boundary conditions are $T = 0$ imposed at the surface and an insulating boundary condition at the base of the mantle; the models are therefore purely internally heated. Models have an aspect ratio of 4×1 and a resolution of 512×128. A finite-volume code, which was explained in *Foley and Bercovici* [2014], is used to solve equations (7)-(11). As the models are purely internally heated, the interior temperature that results when the system reaches steady-state can vary from the assumed value of T_m used in the non-dimensionalization scheme. Thus, T

in the mantle interior (T_i) can vary from 1. As the goal of this work is constraining how mantle temperature affects plate boundary formation via grain damage, this variation in T_i is the desired result of the modeling setup.

All numerical models are started with a mobile lid initial condition. In pseudoplastic models, whether the initial condition is mobile or stagnant lid has been shown to influence the steady-state tectonic regime a model reaches [e.g. *Crowley and O'Connell*, 2012; *Weller and Lenardic*, 2012; *Weller et al.*, 2015]. Specifically, it is harder to initiate plate tectonics from a stagnant lid state than it is to maintain mobile lid convection once it has commenced. Thus, the choice of initial condition for this study favors mobile lid convection as the final state reached in the models. However, *Weller et al.* [2015] shows that increasing internal heating rate expands the stagnant lid regime, and shrinks the mobile lid regime, for both mobile lid and stagnant lid initial conditions, except at very low heat production rates where the opposite trend is seen. Thus the trend of how internal heating rate affects the mode of surface tectonics appears to hold independent of chosen initial condition. Moreover, *Foley et al.* [2014] found that mobile lid convection can readily develop from an initially stagnant lid when grain damage is used, though a thorough study on the effect of initial conditions on the final regime models with grain damage reach is still needed.

A number of metrics and other quantities are calculated from the numerical models as a post-processing step. The base of the lithosphere is determined from the horizontally averaged temperature profile in the mantle. As pure internal heating results in a temperature profile that increases with height from the base of the mantle to the bottom of the lithosphere [e.g. *Moore*, 2008], I define the base of the lithosphere using the maximum of the horizontally averaged temperature profile, $\bar{T}_{\rm max}$; the base of the lithosphere is defined as $T_{\rm bl} = 0.9\bar{T}_{\rm max}$. The total deformational work within the layer, Φ, is calculated by integrating $\tau_{ij}\dot{\varepsilon}_{ij}$ over the whole model domain, and deformational work within the lithosphere ($\Phi_{\rm lith}$) and sub-lithospheric mantle ($\Phi_{\rm man}$) are integrals over just the regions lying above and below the $T_{\rm bl}$ isotherm, respectively. I also define the interior mantle temperature, T_i, as the average temperature within the domain between the height where $\bar{T}_{\rm max}$ occurs and 0.1 units below this height. The average mantle velocity, $v_{\rm man}$, is defined as the whole mantle RMS velocity, and the surface (or plate) velocity, $v_{\rm surf}$, is defined as half the difference between the maximum and minimum horizontal surface velocities.

3 Results

3.1 Regime diagram and convective planform

A large suite of models varying internal heating rate, Q, and the Frank-Kamenetskii parameter for healing, θ_h, are run to map out the boundary between the mobile lid and stagnant lid regimes. The regime boundary is mapped out in terms of the variable θ_h for the following reasons. The damage to healing ratio in the lithosphere, $Dh_m/(Hh_l)$, exerts a major control on the boundary between the stagnant and mobile lid regimes [*Landuyt et al.*, 2008; *Foley et al.*, 2012], and changing θ_h is a way to vary the damage to healing ratio. A lower θ_h results in more rapid grain growth in the lithosphere, i.e. increases h_l, as healing rate decreases less with decreasing temperature than it would with a larger θ_h. Furthermore, a lower θ_h is a rough approximation of the effect of higher surface temperatures, which act to increase the temperature in the mid-lithosphere and thus increase grain growth rates in this region as well [*Landuyt and Bercovici*, 2009; *Foley et al.*, 2012; *Bercovici and Ricard*, 2014]. All models use $Ra_0 = 10^7$, $D = 10^{-2}$, $H = 1.5 \times 10^5$, and $\theta_v = 13.82$ (see Tables 1 & 2 for a compilation of all model parameters and results). The values of Ra_0, D, and H are reasonable estimates for the modern day Earth [*Foley and Rizo*, 2017], which these reference parameters are meant to represent, though it should be noted that uncertainty in these parameters, in particular for f, h_n, and γ_h, is high [see, e.g. *Mulyukova and Bercovici*, 2017]. The chosen θ_v results in 6 orders of magnitude viscosity variation across the lithosphere due to temperature dependence when $T_i = 1$, which is well within the stagnant lid regime in the absence of damage. As discussed in §2.1, θ_h is likely larger than θ_v for peridotite; however, here lower values of θ_h are used to find the boundary between mobile lid and stagnant lid regimes.

The regime diagram (Figure 1) shows two clear trends: at low Q, increasing Q requires larger θ_h for mobile lid convection; at higher Q (greater than $Q \approx 15$), the boundary between regimes is found to be largely insensitive to Q, and actually shows a gradual decrease in the θ_h value required for mobile lid convection as Q increases. The shape of the regime boundary at low Q implies that increasing internal heating rate impedes mobile lid convection, as a larger damage to healing ratio is required in order to produce lid mobility as Q increases. However, this effect is a result of low rates of internal heating leading to low viscosity ratios across the mantle; at low Q,

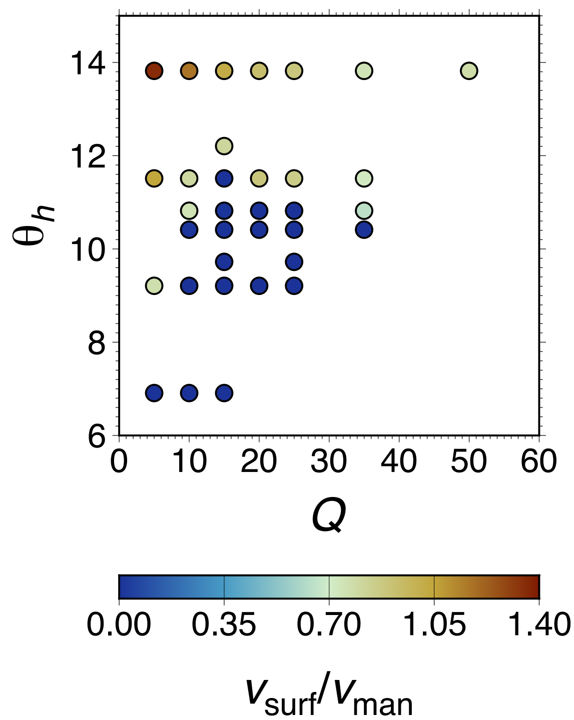

Figure 1: Regime diagram in θ_h-Q parameter space. Model results are colored by the ratio of surface velocity to whole mantle RMS velocity. Thus those colored dark blue, indicating surface velocities are negligibly small, are in the stagnant lid regime.

the viscosity variation between lithosphere and mantle interior is so low that only very modest levels of damage are needed to produce lithosphere mobility. Such a situation is not applicable to the early Earth, where mantle temperatures were high. Moreover, θ_v for mantle rock is higher than used in the convection models, as the very large viscosity variations that result from more realistic values of θ_v result in numerical convergence problems. The more relevant trend is that at higher Q, where the boundary between mobile and stagnant lid is approximately insensitive to Q. Thus, in contrast to the results from models with pseudoplastic yielding, increasing the internal heating rate or the interior mantle temperature (which increases with increasing Q in the models) does not lead to a transition to stagnant lid convection.

However, there are clear changes in the style of lithosphere mobility and subduction that accompany higher rates of internal heating. Examining the planform of convection for mobile lid results with increasing Q (Figure 2) shows a trend towards increasing wavelength (e.g. plates are longer) and a more drip-like style of subduction (example stagnant lid results are also shown in Figure 3). At high Q, the slab undergoes frequent necking, such that subduction takes place via a series of drips rather than with a continuous downgoing slab. An increased frequency of slab necking events at higher mantle temperatures was also seen in regional scale subduction models [*van Hunen and van den Berg*, 2008; *Sizova et al.*, 2010], and occurs for the same physical reasons as in the models shown here. At higher mantle temperatures the viscous resistance to slab sinking in the mantle interior is lower, so slabs sink faster. At the same time, the strength of the convergent boundary in the lithosphere is resisting slab sinking, and the combination of rapid sinking in the mantle interior

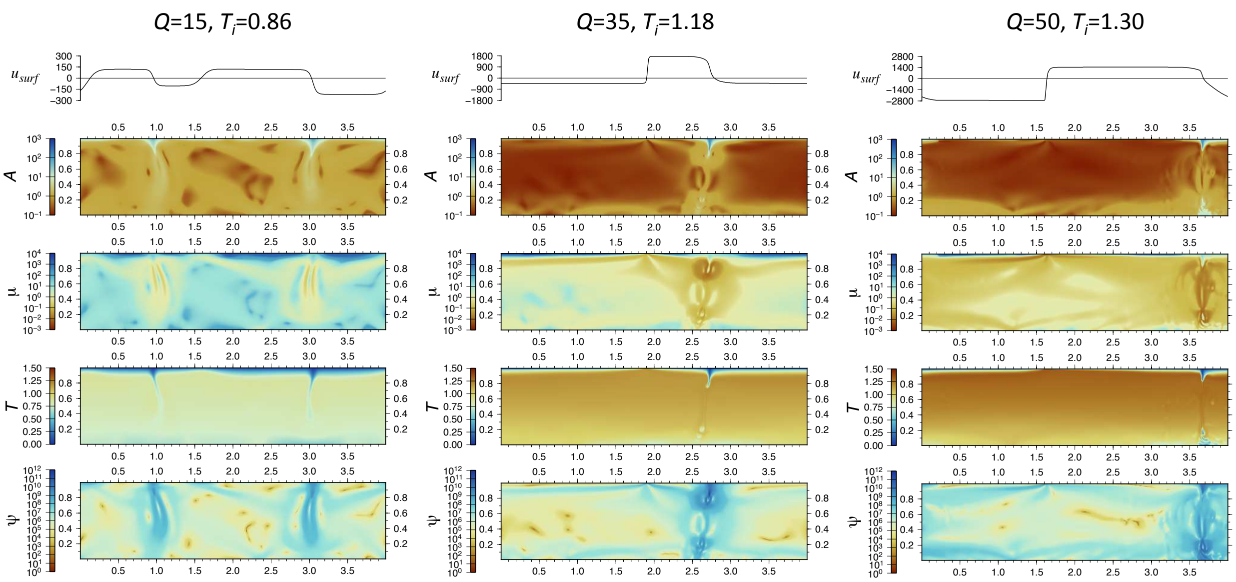

Figure 2: Planform of convection showing the horizontal surface velocity (u_{surf}), fineness field (A), viscosity field (μ), temperature field (T), and the deformational work rate (Ψ). Models with $\theta_h = 13.82$ and $Q = 15$, $Q = 35$, and $Q = 50$ are shown, which result in interior mantle temperatures of $T_i = 0.86$, $T_i = 1.18$, and $T_i = 1.30$, respectively. The perceptually-uniform color map "roma" is used in this study to prevent visual distortion of the data [*Crameri*, 2018a, b].

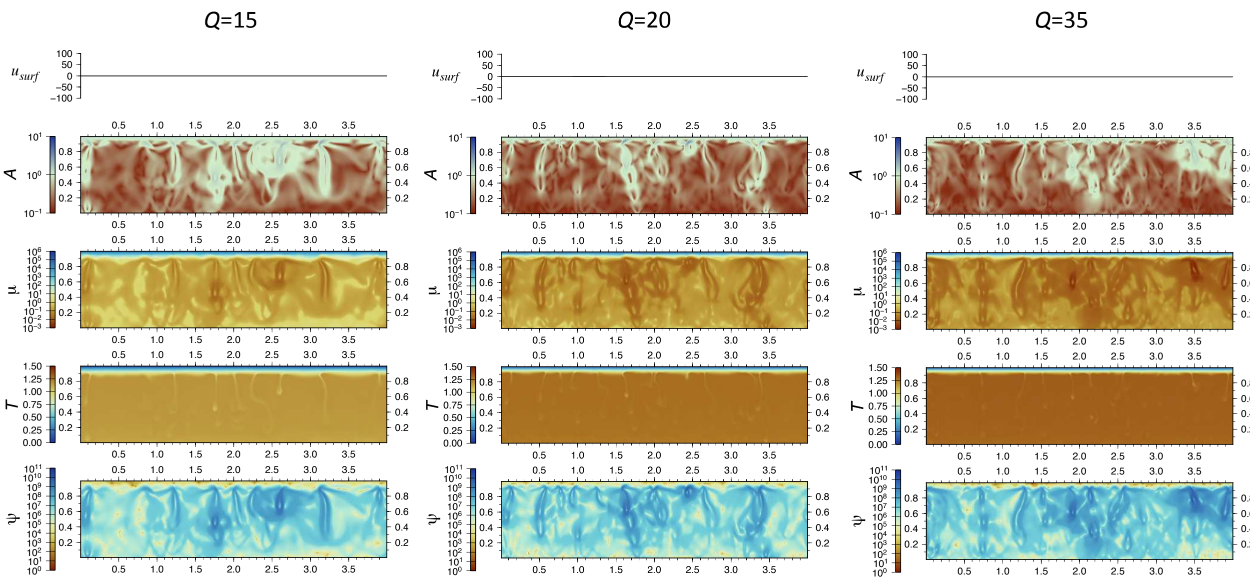

Figure 3: Planform of convection showing the horizontal surface velocity (u_{surf}), fineness field (A), viscosity field (μ), temperature field (T), and the deformational work rate (Ψ) for stagnant lid results. Models with $\theta_h = 10.414$ and $Q = 15$, $Q = 20$, and $Q = 35$ are shown.

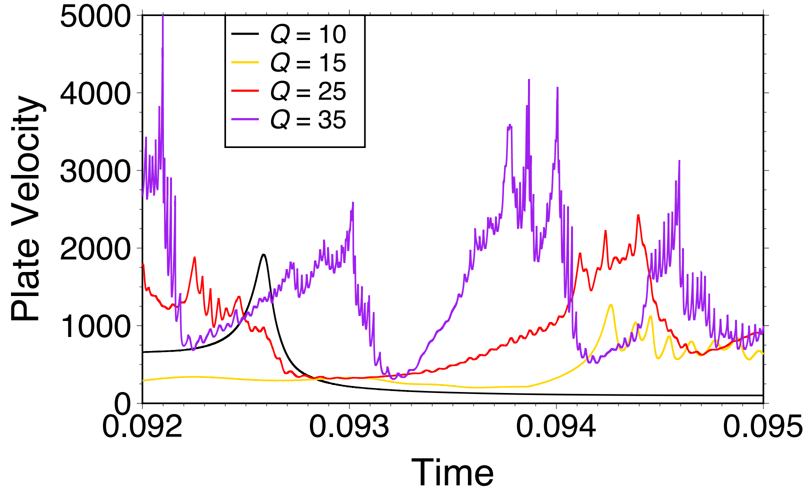

Figure 4: Surface (or plate) velocity as a function of time for models with $\theta_h = 13.82$ and $Q = 10$ (black line), $Q = 15$ (yellow line), $Q = 25$ (red line), and $Q = 35$ (purple line).

and resistance to sinking from the lithosphere leads to slab stretching and eventual tear-off. However, unlike what was commonly seen in subduction models, slab tear-off does not shutdown lithosphere mobility. Convergence in the lithosphere continues until a new slab forms, the new slab then tears off, and the cycle repeats.

The slab necking events lead to a distinct episodicity in surface velocity, which peaks as the slab forms and then decays when the slab breaks away, only to climb again as convergence continues and a new slab forms (Figure 4). Specifically, slab necking leads to a high frequency mode of surface velocity episodicity, which becomes prominent with increasing internal heating rate (e.g. it appears at $Q >\approx 25$ in the models shown here). As a result, the average time spacing between peaks in the surface velocity time series plot decreases from $\approx 3.3 \times 10^{-4}$ at $Q = 15$, to $\approx 6.1 \times 10^{-5}$ at $Q = 25$ and $\approx 2.2 \times 10^{-5}$ at $Q = 35$; in terms of dimensional values these correspond to ≈ 89, ≈ 16, and ≈ 6 Myrs, respectively. I also note that there is clearly a range of frequencies of episodicity, in addition to the high frequency slab necking oscillations. Specifically, there is also a longer wavelength episodicity in surface velocity, likely reflecting reorganizations of the larger scale convection pattern in the mantle. Longer wavelength episodicity is present in the models at all values of Q, but the wavelength decreases with increasing Q as a result of a hotter mantle and more vigorous convection.

3.2 Scaling of deformational work rate during mantle convection

The numerical model results show that the boundary between mobile lid and stagnant lid convection has a different dependency on internal heating rate when grain damage is used than when pseudoplasticity is used. The reason for this difference stems from fundamental physical differences in the plate generation mechanisms. With pseudoplasticity, the stress in the lithosphere is the only factor that determines whether mobile lid convection can be produced, as stresses must be high enough to cause yielding deep into the lithosphere for lid mobility [e.g *Wong and Solomatov*, 2015]. However, grain size reduction is driven by the deformational work done by mantle convection in the lithosphere, so it is the rate of deformational work, $\tau_{ij}\dot{\varepsilon}_{ij}$, that is responsible for plate boundary formation with grain damage, rather than the stress state.

In steady-state (or statistical steady-state for time-dependent convection), the total deformational work in the convecting fluid must balance with surface heat loss

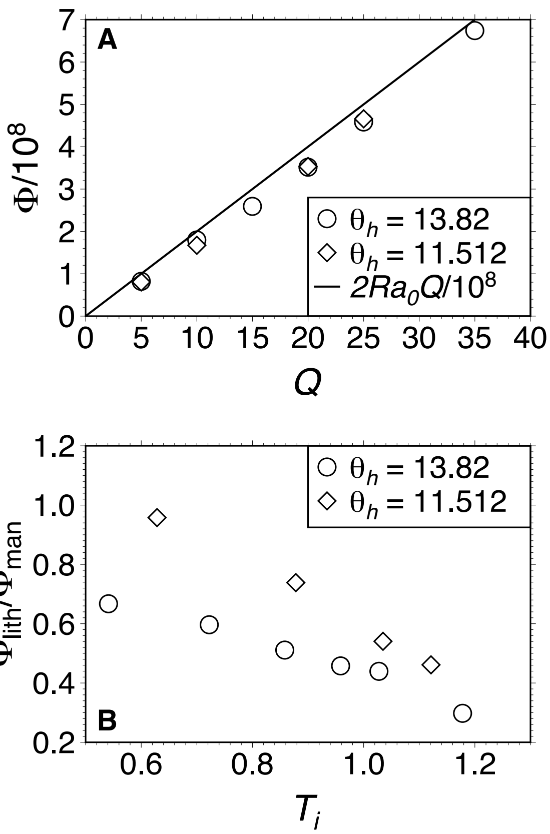

Figure 5: (A) Total deformational work within the convecting layer, Φ, as a function of internal heating rate, Q, and (B) ratio of total deformational work in the lithosphere (Φ_{lith}) to total deformational work in the sub-lithospheric mantle (Φ_{man}). Mobile lid models with $\theta_h = 13.82$ and $\theta_h = 11.512$ are shown. In Figure 5A the scaling law for total deformational work (13) is also plotted, with $L = 4$.

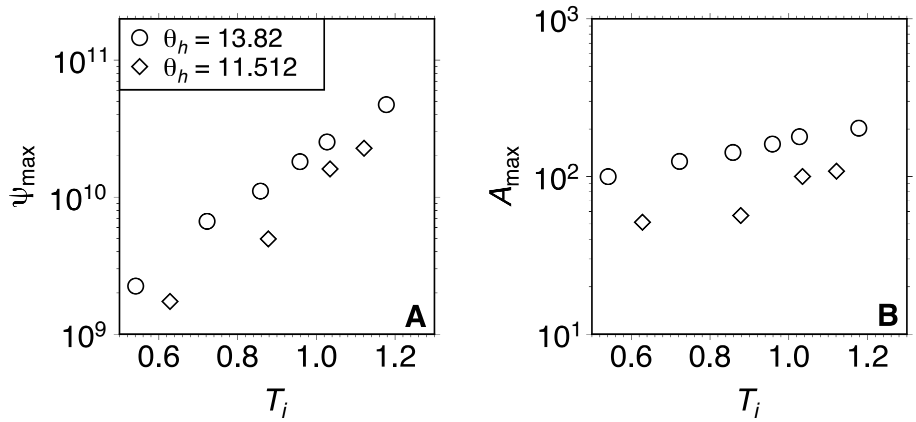

Figure 6: Maximum rate of deformational work in the lithosphere, $\Psi_{\max}$, as a function of interior mantle temperature, T_i for mobile lid models with $\theta_h = 13.82$ and $\theta_h = 11.512$ (A), and maximum lithospheric fineness, $A_{\max}$ as a function of T_i for the same models (B).

[e.g. *Hewitt et al.*, 1975; *Solomatov*, 1995; *Crowley and O'Connell*, 2012], giving

$$\Phi = \int_V \tau_{ij}\dot{\varepsilon}_{ij}dV = Ra_0\left(Nu - \frac{Q}{2}\right)L \tag{12}$$

in the limit of large Nusselt number, Nu; L in (12) is the aspect ratio of the convecting domain, so $L = 4$ for the models presented in this study. For two-dimensional Cartesian convection in thermal steady-state, $Nu \approx Q$, so

$$\Phi \approx Ra_0 QL/2. \tag{13}$$

The numerical models follow (13) (see Figure 5A), confirming that with increasing Q, the total deformational work increases. Thus the energy source for forming plate boundaries with grain damage actually increases with larger internal heating rates, in contrast to pseudoplastic yielding.

The rate of deformational work, Ψ, varies significantly in space throughout the convecting layer, but a substantial fraction occurs in the high viscosity lithosphere, concentrated in shear zones where the lithosphere converges and downwells (Figure 2). The percentage of total deformational work in the lithosphere generally decreases with increasing mantle temperature, ranging from up to ≈ 50 % at low T_i to ≈ 15 % at high T_i (Figure 5B). The total lithospheric deformational work is dominated by work done in shear zones, such that the maximum value of Ψ in the lithosphere is always within convergent shear zones when convection is in a mobile lid regime. $\Psi_{\max}$, the maximum deformational work rate in the lithosphere, thus directly measures the energy source available for grain size reduction in lithospheric shear zones. The numerical models show that $\Psi_{\max}$ increases with increasing mantle interior temperature, further demonstrating that the energy source for plate boundary formation with grain damage is enhanced at the mantle thermal conditions expected for the early Earth (Figure 6A). As a result, the maximum fineness in lithospheric shear zones, $A_{\max}$, also increases with increasing T_i (Figure 6B). Thus, when grain damage is used, high rates of internal heating do not cause a transition to stagnant lid convection, because the lithospheric deformational work rate increases with Q, and thus lithospheric shear zone fineness also increases.

Why stress and deformational work rate can scale in opposite ways with increasing mantle temperature or internal heating rate is a result of tradeoffs between how strain rate and viscosity change. Increasing mantle temperature generally decreases mantle viscosity, and in turn increases the typical mantle flow speeds, and thus the characteristic strain rate of mantle flow, as a result of higher thermal buoyancy forces and a lower viscous resistance to flow. As the characteristic stress, $\tau = 2\mu\dot{\varepsilon}$, where $\dot{\varepsilon}$ is the characteristic strain rate, increasing mantle temperature has competing effects on how stresses in the mantle and lithosphere will scale. Previous studies have found that the decrease in viscosity generally wins out over the increase in strain rate, causing stress to drop as mantle temperature climbs [e.g. *O'Neill et al.*, 2007]. However, as shown here, the energetics of convection require that the total deformational work increase with Nusselt number and internal heating rate. As a result, the rate of deformational work in lithospheric shear zones is positively correlated with mantle temperature (Figure 6A). As with the characteristic stress, the way the characteristic deformational work rate, $\Psi_c = 2\mu\dot{\varepsilon}^2$, scales with mantle temperature is determined by the competing effects of mantle temperature on viscosity and strain rate. However, in this case the strain rate squared term must win out over the decrease in viscosity to produce the observed trends in deformational work rate.

3.3 Changes in subduction and lithosphere mobility style

Although a transition to stagnant lid convection at high internal heating rates or high mantle temperatures is not seen, a hotter mantle does significantly change the style of subduction and lithosphere mobility. Lithospheric shear zones are weaker due to smaller grain sizes resulting from more effective damage, but the mantle interior also gets weaker with increasing temperature, and at a faster rate than lithospheric shear zones (Figure 7). As a result, the ratio of lithospheric shear zone viscosity to sub-lithospheric mantle viscosity increases with T_i (Figure 8A). Higher mantle temperatures also cause higher grain growth rates, and thus larger grain sizes in the mantle interior. However, the effect of grain size in the ambient mantle interior, away from downwellings where damage causes localized zones of grain size reduction that significantly influence the rheology around downwellings, is small compared to viscosity temperature-dependence; average fineness in the mantle interior decreases by only a factor of ≈ 6 when going from $T_i \approx 0.5$ to $T_i \approx 1.2$ with $\theta_h = 13.82$, resulting in a factor of 36 increase in viscosity. Meanwhile temperature-dependence acts to decrease viscosity by over four orders of magnitude over the same temperature interval. The way mantle and lithosphere viscosities change with increasing mantle temperature also causes the ratio of surface velocity to average mantle velocity to decrease (Figure 8B). Convection therefore enters a "sluggish lid" regime at higher mantle temperatures, as defined by *Crowley and O'Connell* [2012], where the lithosphere remains mobile and subduction active, but lithosphere velocities are much slower than interior mantle velocities. Lithosphere velocities can still be as large as modern Earth plate speeds, or larger, but as long as interior mantle velocities exceed surface plate velocities, then convection is considered to be in a sluggish lid regime.

The ratio of shear zone viscosity to mantle interior viscosity measures the strength of plate boundaries relative to the mantle, and the model results indicate this relative strength increases with T_i. In addition to causing the ratio of surface to mantle velocity to decrease with T_i, the increasing relative strength of plate boundaries also causes the change in subduction style from continuous, long-lived subduction, to drip-like subduction with frequent slab necking, as discussed in §3.1, and likely plays a role in the generally longer wavelength convection patterns seen at high heat production rates as well. As the relative strength of plate boundaries increases, thicker plates

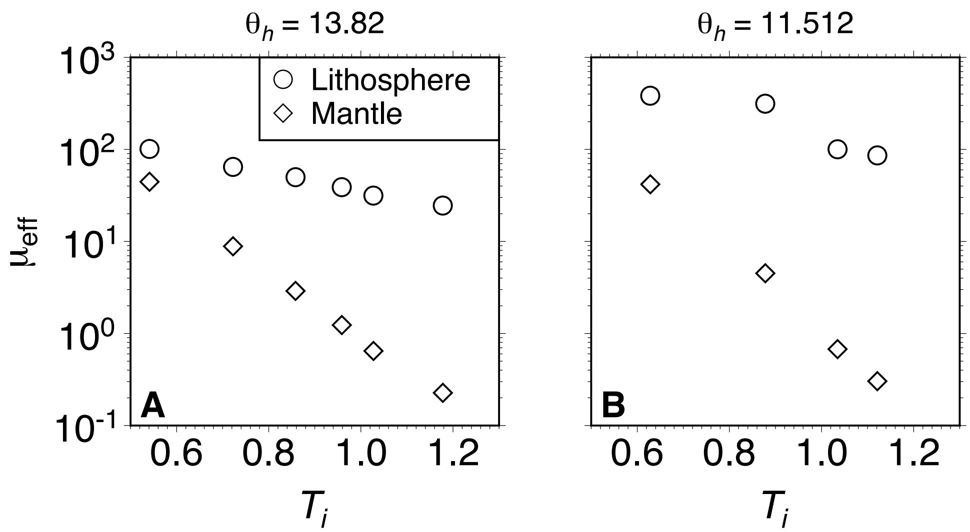

Figure 7: Effective viscosity of lithospheric shear zones (circles) and the mantle interior (diamonds) as a function of T_i, for mobile lid models with $\theta_h = 13.82$ (A) and $\theta_h = 11.512$ (B).

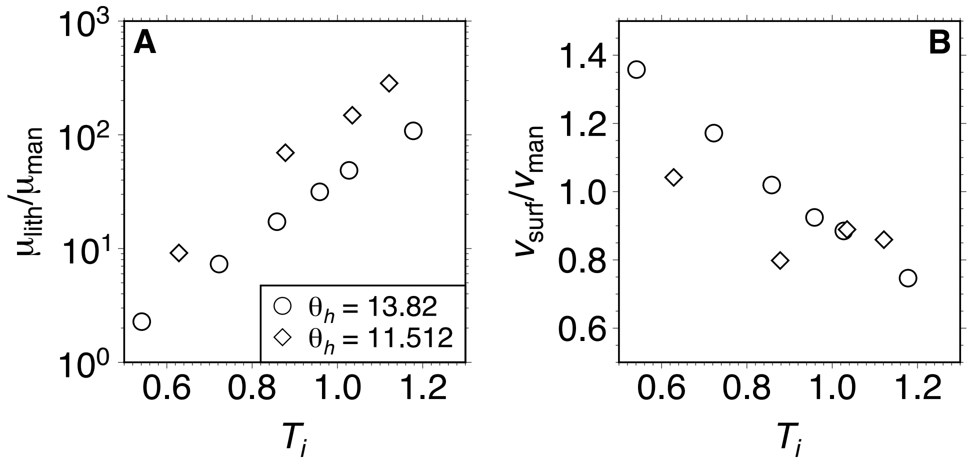

Figure 8: Ratio of the effective viscosity of lithospheric shear zones to the interior mantle viscosity as a function of T_i (A), and ratio of surface velocity to whole mantle RMS velocity (B), for mobile lid models with $\theta_h = 13.82$ and $\theta_h = 11.512$.

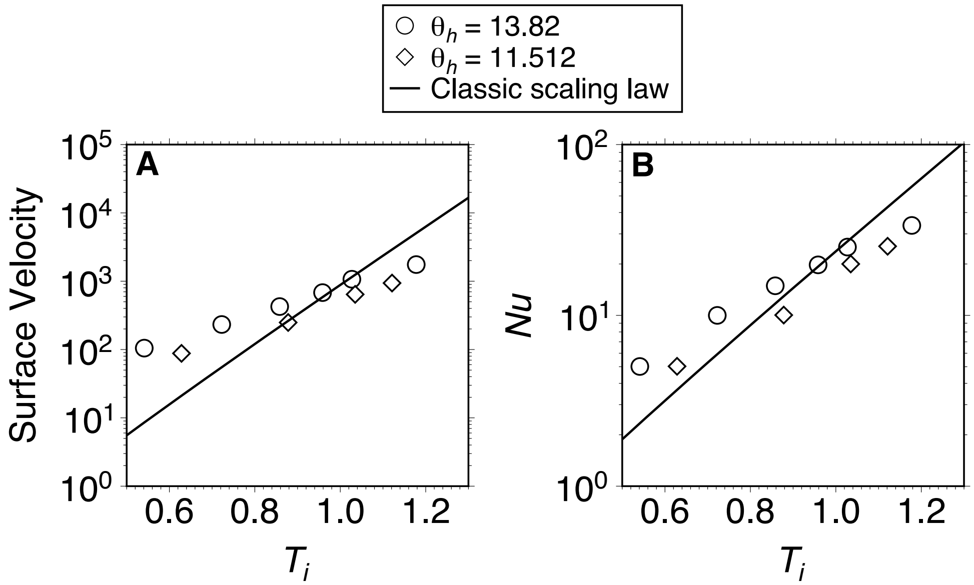

Figure 9: Surface velocity (A) and Nusselt number (B) as a function of T_i for mobile lid models with $\theta_h = 13.82$ and $\theta_h = 11.512$. Also plotted are the constant viscosity scaling laws for v_{surf} (16) and Nu (15). Scaling law constants b and a, from (16) and (15), repsectively, are picked to fit the models with $\theta_h = 13.82$ at $T_i = 1$.

are required before subduction can occur, therefore resulting in longer wavelength convection [e.g. *McNamara and Zhong*, 2005]. Increasing internal heating rate also increases the sub-adiabatic temperature gradient in the mantle, where temperatures beneath the lithosphere are warmer than one would expect along an adiabat, and the deeper mantle is cooler; this temperature profile also favors long-wavelength convection patterns as a result of the depth-dependent viscosity profile that ensues [e.g. *Bunge et al.*, 1996; *Tackley*, 1996; *Lenardic et al.*, 2006].

Another important effect of how increasing mantle temperature influences mobile lid convection produced via grain damage, is that surface velocity and heat flux scale differently than classic scaling laws for mantle convection predict. These classic scaling laws are often used in thermal evolution models to estimate how plate tectonics would change at different mantle thermal conditions [e.g. *Davies*, 2007]. The typical scaling law for surface heat flux from constant viscosity convection, or convection that can be approximated with a constant viscosity in the mantle interior, is $Nu = aRa_i^{1/3}$ [e.g. *Turcotte and Schubert*, 2002; *Solomatov*, 1995], where Ra_i is the internal Rayleigh number, defined as

$$Ra_i = \frac{\rho g \alpha (T_i - T_s) d^3}{\kappa \mu(T_i)}. \tag{14}$$

Thus, Nusselt number scales as

$$Nu = a \left(\frac{Ra_0 T_i}{\exp\left(\theta_v (1 - T_i)\right)} \right)^{1/3}, \tag{15}$$

when the Frank-Kamenetskii viscosity law is used. Likewise, the scaling law for surface velocity for constant viscosity convection is

$$v_{\text{surf}} = b \left(\frac{Ra_0 T_i}{\exp\left(\theta_v (1 - T_i)\right)} \right)^{2/3}. \tag{16}$$

While the above scaling laws for Nu and v_{surf} strictly only apply to high Rayleigh number, constant viscosity convection, mobile lid convection with pseudoplasticity has been found to follow $Nu \sim Ra_i^{1/3}$ and $v_{\text{surf}} \sim Ra_i^{2/3}$ scaling relationships in previous studies [*Moresi and Solomatov*, 1998; *Korenaga*, 2010a], at least when additional complications, such as viscosity layering, are not present [*Crowley and O'Connell*, 2012]. Thus, these classic mantle convection scaling laws have been used to estimate that the early Earth would experience rapid plate speeds and have a high convective heat flux, if plate tectonics were operating at this time [e.g. *Davies*, 2007].

However, mobile lid convection produced with grain damage does not follow these classic scaling laws, because: 1) the relative strength of lithospheric shear zones compared to the underlying mantle increases with interior mantle temperature (Figure 8A); and, 2) the wavelength of convection shows clear changes with increasing temperature, and hence Ra_i, an effect which is known to cause deviations from the classic convection scaling laws [*Zhong*, 2005]. As a result, surface velocities at high temperatures (e.g. above $T_i = 1$) are slower than would be expected if convection followed the classic convection scaling law. Thus mobile lid convection, if it were active on the early Earth, need not result in rapid plate speeds and high surface heat fluxes, and instead may lie a sluggish lid regime as in the convection models shown in this paper. The style of lithospheric mobility is important for interpreting a range of geological and geochemical observations of early Earth evolution, as discussed in more detail in §4.2. In fact, early Earth plate velocities could even be lower than present day plate speeds, while still in a mobile lid regime. Higher mantle temperatures also enhance the effective grain growth rate in the lithosphere, an effect not captured in the convection models presented here, which would cause plate speeds to actually decrease with increasing mantle temperature as a result of strengthening lithospheric shear zones [*Foley et al.*, 2014]. A similar result has also been proposed on the basis of dehydration stiffening of the mantle lithosphere during ridge melting [*Korenaga*, 2006].

4 Discussion

A major implication of the results of this study is that mantle thermal state is overall less important for whether a planet shows mobile lid or stagnant lid tectonics than many previous studies have indicated. Instead, surface temperature and planet size may have a much stronger control over a planet's style of tectonics. Specifically, low surface temperatures and larger planet sizes have been found to favor a mobile lid for convection with grain damage, a result that is also consistent with the planets in our solar system [*Foley et al.*, 2012]. Low surface temperatures suppress healing in the lithosphere, thereby enhancing damage and making mobile lid convection more likely [*Landuyt and Bercovici*, 2009; *Foley et al.*, 2012; *Bercovici and Ricard*, 2014], while deformational work rate increases with increasing planet size [*Foley et al.*, 2012]. That different mechanisms for generating plate boundaries produce different results for early Earth tectonics is not surprising, but does highlight the importance of the microphysics behind shear localization, as discussed more in the next section.

Furthermore, as early Earth thermal conditions do not impede the formation of weak plate boundaries with grain damage, no significant external factors or effects, beyond normal mantle convective forces, are needed to initiate and sustain a mobile lid. *Foley et al.* [2014] found that mobile lid convection can rapidly commence from an initial stagnant lid, solely as a result of Hadean mantle convection. Many authors have argued for additional mechanisms, beyond the forces from simple thermal convection, as the key for initiating plate tectonics; proposals include subduction induced by plumes [*Gerya et al.*, 2015], formation of early continents [*Rey et al.*, 2014], or meteorite impacts [*O'Neill et al.*, 2017]. Such extra mechanisms would be necessary if, as results from using a pseudoplastic rheology, high temperatures or high heat production rates impede shear zone formation. However, the results presented here and in *Foley et al.* [2014] indicate that such external subduction initiation mechanisms are not needed to start mobile lid tectonics on Earth, as weak plate boundaries can readily form from mantle convection at early Earth conditions.

4.1 Rheology and geodynamic predictions of planetary surface tectonics

As illustrated in §3.2, the primary reason why models with pseudoplasticity and those with grain damage produce different predictions for early Earth tectonics stems from fundamental physical differences in how these mechanisms are formulated. This discrepancy highlights the importance of understanding the microphysics of shear localization and plate boundary formation for studying early Earth evolution; ultimately a thorough understanding of lithosphere rheology, and incorporation of this understanding into geodynamic models, is needed to determine how and why plate tectonics developed on Earth. For example, it has been argued that the pseudoplastic rheology can be thought of as an approximation of the more complicated microscale processes leading to lithospheric failure, such as grain size evolution and fabric development. In the case of grain size reduction, it is true that grain damage produces an effectively non-Newtonian stress-strain rate relationship, similar to pseudoplasticity, when grain size is given by (2) at steady-state [e.g *Foley et al.*, 2012; *Foley and Driscoll*, 2016]. However, this effective constitutive relationship is itself a function of mantle thermal state, among other factors, and these dependences are not captured by a simple pseudoplastic formulation. Thus, pseudoplasticity is not a good approximation of processes such as grain size reduction, when trying to constrain the physical factors that allow mobile lid convection to develop on a planet.

4.2 Sluggish lid convection and Archean tectonics

Sluggish lid convection in the Hadean and Archean can potentially explain geological and geochemical observations of early Earth crust formation and tectonics. A number of lines of evidence indicate that the style of tectonics was different from modern day plate tectonics, or at least different from what simple models of plate tectonics on a hotter Earth would predict. Many craton cores from the Eoarchean and Archean are thought to have formed in ocean plateau settings rather than subduction zone settings. Detailed studies of crustal geochemistry of the ≈ 4.0 Ga Acasta Gneiss complex in Canada indicate this early section of felsic crust formed in a setting similar to modern day Iceland [*Reimink et al.*, 2014; *Reimink et al.*, 2016]. The "dome and keel" structures of Archean felsic rocks in the Pilbara terrane in Australia, as well as the geochemistry of this crust, also point towards formation in an ocean plateau-like setting [e.g. *Pease et al.*, 2008; *Smithies et al.*, 2009; *Van Kranendonk*, 2010; *Van Kranendonk et al.*, 2015; *Johnson et al.*, 2017]; in particular it's not clear how subduction could form the large spatial scale felsic "domes" seen in the Pilbara [e.g. *Collins et al.*, 1998; *Thébaud and Rey*, 2013; *François et al.*, 2014]. There is also geochemical evidence for episodes of subduction "failure" (i.e. slab breakoff), rather than sustained subduction, seen in Archean greenstone belts [*Smithies et al.*, 2018].

On a global scale, some studies argue for transitions in crust formation processes in the Archean, around $\sim$ 3.0 Ga, possibly indicating a change in Earth's tectonic mode. *Dhuime et al.* [2012] argues that the rate of continental crust growth changed at $\approx$ 3.0 Ga, possibly as a result of a changing style of mantle dynamics. However, the history of continental crust growth is debated, and alternative scenarios with no change in crust growth rates in the Archean are also possible [*Korenaga*, 2018; *Rosas and Korenaga*, 2018]. Studies from global geochemistry datasets indicate that the composition of the continental crust appears to have changed, from a more mafic composition prior to $\sim$ 3.0 Ga to a more felsic composition after this time [e.g. *Keller and Schoene*, 2012; *Dhuime et al.*, 2015; *Tang et al.*, 2016]. This trend in crust composition has often been interpreted as representing the onset of plate tectonics, though not all studies agree; *Keller and Schoene* [2018] argue that the evolution of continental basaltic geochemistry over time actually represents plate tectonic processes throughout Earth history. Moreover, there are significant uncertainties in reconstructing the history of continental crust composition, and felsic early continents, potentially formed via plate tectonics, are still a viable possibility [e.g. *Greber et al.*, 2017].

Paired metamorphic belts (assemblages of low temperature, high pressure metamorphic rocks and low pressure, high temperature rocks) are thought to be indicative of subduction, as they are formed in arc and back arc settings; these are notably absent before $\sim$ 3.0 Ga [*Brown*, 2014]. Similarly *Shirey and Richardson* [2011] argue that the appearance of eclogite inclusions in diamonds at $\sim$ 3.0 Ga records the start of not only plate tectonics, but the full Wilson cycle. Finally, mantle chemical heterogeneity formed during the first few hundred million years of Earth history persists in the mantle for $\sim 1-2$ Gyrs [e.g. *Rizo et al.*, 2013; *Debaille et al.*, 2013]. Such long mixing times are difficult to explain with simple models of early Earth plate tectonics, where a hot mantle is expected to result in vigorous convection and rapid mantle mixing [*Debaille et al.*, 2013; *O'Neill and Debaille*, 2014]. In fact early formed heterogeneity is even seen in modern rocks thought to be sourced from the large low shear velocity provinces (LLSVPs) in the lower mantle [*Rizo et al.*, 2016; *Mundl et al.*, 2017; *Horan et al.*, 2018; *Peters et al.*, 2018]. As the LLSVPs are typically hypothesized to be chemically dense, and thus resistant to mixing with the rest of the mantle [e.g. *Garnero and McNamara*, 2008], heterogeneity sourced from these regions probably does not constrain the efficiency of mantle mixing. However, the Archean rocks showing resolvable signatures of early formed heterogeneity do not show evidence of being sourced from LLSVPs, so they likely do constrain the efficiency of ancient mantle mixing.

On the other hand, there is also evidence for subduction in the Archean and Eoarchean, and possibly even the Hadean. Hadean zircons from the Jack Hills region of Australia have been used to argue for subduction on the Hadean Earth, based on the chemistry of their inclusions and pressure-temperature conditions where they formed [*Harrison et al.*, 2005; *Hopkins et al.*, 2008], though non-subduction formation scenarios for these zircons are also possible [*Kemp et al.*, 2010; *Nebel et al.*, 2014]. Furthermore, evidence for subduction is seen in the 3.8 Ga Isua greenstone belt in Greenland [*Jenner et al.*, 2009; *Polat et al.*, 2011], the variably dated $4.4-3.8$ Ga Nuvvuagittuq supracrustal belt in Canada [e.g. *Turner et al.*, 2014], the ≈ 3.6 Ga Acasta Gneiss Complex in Canada [*Koshida et al.*, 2016], and in Australia at ≈ 3.2

Ga [e.g. *Van Kranendonk*, 2010] and South Africa at $\approx$ 3.5 Ga [e.g. *Moyen et al.*, 2007]; see also *Moyen and Martin* [2012]. Thus early Earth tectonics likely featured at least spatially limited, transient episodes of subduction.

A stagnant lid with short-lived episodic bursts of subduction has often been invoked to explain the above observations from the Hadean and Archean geologic record [e.g. *Moore and Webb*, 2013; *Griffin et al.*, 2014; *O'Neill et al.*, 2015; *Bédard*, 2018; *Condie*, 2018]. However, a sluggish lid, where subduction occurs more frequently and on a locally confined scale as compared to the short-lived, global subduction and resurfacing events seen during episodic overturns, may also be able to explain early Earth geology. A sluggish lid, featuring subduction and lithospheric mobility, can explain long-term preservation of early-formed mantle chemical heterogeneity, as a sluggish lid also results in slow mixing [*Foley and Rizo*, 2017]. In fact, even plate-tectonic style convection throughout the Hadean and Archean can allow early formed heterogeneity to survive when variations in lower mantle mineralogy are taken into account, as high viscosity, silica rich regions that are resistant to mixing tend to form in the cores of convection cells [*Ballmer et al.*, 2017]. Paleomagnetic reconstructions of super-continent cycles show a trend of gradually increasing plate speeds over time, rather than the decreasing trend one would expect from models based on the classic mantle convection scaling laws discussed in §3.3 [*Condie et al.*, 2015]. Such a trend is also consistent with predictions from grain damage models that explicitly include the effect of higher mantle temperatures on lithospheric grain growth rates [*Foley et al.*, 2014], an effect which is not included in the convection models shown in this study (see §3.3).

Moreover, a sluggish lid regime would result in less efficient convective heat transport than modern style plate tectonics (see §3.3). Earth's heat budget, where less than 50 % of Earth's present day heat flow is thought to be derived from radiogenic heat production [e.g. *Jaupart et al.*, 2007], is difficult to reconcile with rapid plate speeds and a high heat flux on the early Earth; extrapolating back in time from modern day conditions results in a "thermal catastrophe," where Earth is predicted to be in a molten state only ~ 1 Gyrs ago [*Korenaga*, 2003]. One possible solution to the "thermal catastrophe" problem is if heat flux does not increase strongly with mantle temperature, as this allows the modern Earth's heat loss to be dominated by secular cooling without requiring mantle temperatures to increase sharply when extrapolated back in time [*Korenaga*, 2006]. Models of early Earth sluggish lid convection with grain damage indicate that heat flux is a weak function of mantle temperature, especially if the effect of mantle temperature on lithospheric grain growth rate is included [*Foley et al.*, 2014]. Thus sluggish lid convection may also work to reconcile Earth's thermal history, though more detailed thermal evolution models are needed to demonstrate this robustly.

Inefficient heat transport on the early Earth may even explain the dominant crust formation processes thought to be operating at that time. Based on the estimates in *Foley et al.* [2014], sluggish lid convection at a mantle temperature of 2000-2100 K, representative of the Earth just after magma ocean solidification, would only result in ≈ 60 TW of convective heat flow, while radiogenic heat production would yield $\approx 75 - 100$ TW. Thus significant heat loss via mantle melting and volcanism is required to balance the heat budget, similar to, though less extreme, than the heat-pipe volcanism proposed by *Moore and Webb* [2013] for the Hadean. As a result, there would be extensive volcanism capable of creating thick ocean plateaus that could internally re-melt to form felsic crust, all while subduction is still active. If this extensive volcanism, which would occur largely as a result of passive mantle upwelling, produces significantly more felsic crust than any arc volcanism present at this time, then felsic crust production predominantly forming outside of subduction zones can be explained without needing a stagnant lid. Stagnant lid convection would of course also result in felsic crust formation in non-subduction settings, since subduction is absent in this regime, but a sluggish lid allows for subduction without requiring sudden, global overturn events. In the sluggish lid conceptual model, mantle upwelling volcanism will wane as the mantle cools and volcanic heat loss becomes less significant, leading to subduction zones then becoming the primary setting for continental crust production, an idea also proposed by *Van Kranendonk* [2010]. The above hypothesis is admittedly speculative at the present, and thus requires more detailed models of crust formation in a sluggish lid regime to be fully tested. However, given the potential for reconciling a wide range of geological observations of the early Earth with a sluggish lid, this hypothesis is worth pursuing further.

5 Conclusions

Convection models with grain damage, a theoretical formulation for grain size evolution that allows weak lithospheric shear zones to form via grain size reduction, show that the boundary between stagnant lid and mobile lid convection is only weakly sensitive to mantle interior temperature or heat production rate. The near insensitivity of the mode of surface tectonics to mantle thermal state is in contrast to previous modeling studies performed with a pseudoplastic rheology; these studies found that increasing internal heating rate or mantle temperature decreases stresses in the lithosphere, therefore favoring a stagnant lid regime. The discrepancy in results stems from fundamental physical differences in the two mechanisms for generating weak plate boundaries. With grain damage, the work done by deformation in the lithosphere drives grain size reduction and subsequent plate boundary formation, while with pseudoplasticity the lithospheric stress state determines whether weak shear zones can form. The deformational work increases with increasing internal heating rate or mantle interior temperature, therefore allowing weak plate boundaries to form by grain size reduction at early Earth thermal conditions. Thus the question of early Earth geodynamics is intimately tied to lithosphere rheology. The microphysical processes that lead to shear localization in the lithosphere must be thoroughly understood, in order to constrain the geodynamics of the early Earth.

Although increasing mantle temperature towards early Earth conditions does not have a significant impact on the overall regime of mantle convection (stagnant or mobile lid) when grain damage is used, it does change the style of subduction and lithospheric mobility. Surface, or plate, motions become increasingly sluggish compared to flow velocities in the mantle interior as internal temperature increases. As a result, subduction becomes drip-like, and no longer features a coherent slab extending deep into the mantle. Drip-like subduction is a result of slabs constantly tearing-off as they form and sink into the mantle, which also leads to a distinct, high frequency oscillation in plate speeds at the surface. This style of sluggish lid convection can potentially explain key observations of the early Earth geologic record, such as long-term preservation of early formed mantle heterogeneity, or changes in the rate of continental crust formation and average composition of the continental crust in the Archean. Sluggish lid convection leads to slow mantle mixing and hence long-term preservation of mantle chemical heterogeneity. Significant volcanism outside of normal ridge or arc settings would also be expected for a sluggish lid at early Earth thermal conditions, potentially allowing for crust formation in ocean plateau-like settings to be the dominant crust forming process. However, more detailed models combining geodynamics and petrology are needed to better test the hypothesis of early Earth sluggish lid convection, as outlined in the next section.

6 Future Directions

More work on lithospheric rheology, in particular the microphysics of plate boundary formation, is clearly needed to make significant progress on the question of early Earth evolution. However, another vital area of future work is integrating geodynamics, geochemistry, and petrology to better test models of early Earth tectonics. In particular, geodynamic models of early Earth tectonics need to track the conditions under which melting and crust formation would take place, so they can be compared with the geologic record. Such an approach is only recently being attempted [e.g. *Walzer and Hendel*, 2008; *Gerya*, 2011; *Johnson et al.*, 2014; *Sizova et al.*, 2015; *Rozel et al.*, 2017; *Walzer and Hendel*, 2017], and is a promising direction for progress in the coming years. In addition to testing whether the tectonic modes produced in theoretical models are consistent with geological observations, systematic testing of different hypotheses for early Earth tectonics is also needed. In particular, working through the various key observations of early Earth crust formation to determine which styles of global tectonics are, or are not, consistent with these observations would be a major advance in our understanding of early Earth evolution. Presently it is still not clear which features of the Archean geologic record provide strong constraints on the style of tectonics, versus those that can be created by multiple different mechanisms and thus are compatible with multiple different modes of surface tectonics. With better integration of geodynamics, geochemistry, and petrology, and in particular systematic testing of different geodynamic scenarios against the ancient geologic record, progress on these critical questions can be achieved in the future.

7 Data Availability

The code used to produce the results shown in this paper can be accessed here: https://github.com/bradfordjfoley/foley-convection-code. Tables of the numerical model results, showing relevant parameters and key output are given in Tables 1 & 2.

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Table 1: Table of mobile lid model results

Ra_0	θ_v	θ_h	D	H	Q	Resolution	Φ	v_{surf}	v_{man}	Ψ_{max}	A_{max}
10^7	13.82	13.82	10^{-2}	1.5×10^5	5	512×128	8.21×10^7	142.2	104.7	2.24×10^9	99.67
10^7	13.82	13.82	10^{-2}	1.5×10^5	10	512×128	1.80×10^8	272.4	232.5	6.66×10^9	124.7
10^7	13.82	13.82	10^{-2}	1.5×10^5	15	512×128	2.59×10^8	433.4	425.1	1.11×10^{10}	141.9
10^7	13.82	13.82	10^{-2}	1.5×10^5	20	512×128	3.52×10^8	630.1	681.7	1.81×10^{10}	160.6
10^7	13.82	13.82	10^{-2}	1.5×10^5	25	512×128	4.58×10^8	945.5	1068	2.52×10^{10}	178.8
10^7	13.82	13.82	10^{-2}	1.5×10^5	35	512×128	6.74×10^8	1308	1752	4.73×10^{10}	202.4
10^7	13.82	13.82	10^{-2}	1.5×10^5	50	512×128	6.82×10^8	2214	2849	6.36×10^{10}	212.6
10^7	13.82	12.206	10^{-2}	1.5×10^5	15	512×128	2.61×10^8	391.4	481.8	1.04×10^{10}	93.29
10^7	13.82	11.513	10^{-2}	1.5×10^5	5	512×128	8.07×10^7	91.53	87.85	1.73×10^9	51.26
10^7	13.82	11.513	10^{-2}	1.5×10^5	10	512×128	1.67×10^8	198.6	248.8	4.95×10^9	56.56
10^7	13.82	11.513	10^{-2}	1.5×10^5	20	512×128	3.53×10^8	570.1	641.2	1.60×10^{10}	100.0
10^7	13.82	11.513	10^{-2}	1.5×10^5	25	512×128	4.66×10^8	807.7	939.8	2.27×10^{10}	108.0
10^7	13.82	11.513	10^{-2}	1.5×10^5	35	512×128	5.98×10^8	1292	1797	2.61×10^{10}	109.6
10^7	13.82	10.82	10^{-2}	1.5×10^5	5	512×128	1.64×10^8	195.0	259.8	5.28×10^9	56.57
10^7	13.82	10.82	10^{-2}	1.5×10^5	35	512×128	5.59×10^8	1236	1906	2.36×10^{10}	91.61
10^7	13.82	9.21	10^{-2}	1.5×10^5	5	512×128	7.72×10^7	65.65	85.44	1.21×10^9	27.84

Table 2: Table of stagnant lid model results

Ra_0	θ_v	θ_h	D	H	Q	Resolution	v_{surf}	v_{man}
10^7	13.82	6.908	10^{-2}	1.5×10^5	5	512×128	0.074	272.9
10^7	13.82	6.908	10^{-2}	1.5×10^5	10	512×128	0.027	612.2
10^7	13.82	6.908	10^{-2}	1.5×10^5	15	512×128	0.033	614.4
10^7	13.82	9.21	10^{-2}	1.5×10^5	10	512×128	0.25	774.0
10^7	13.82	9.21	10^{-2}	1.5×10^5	15	512×128	0.20	1035
10^7	13.82	9.21	10^{-2}	1.5×10^5	20	512×128	0.070	1404
10^7	13.82	9.21	10^{-2}	1.5×10^5	25	512×128	0.055	1851
10^7	13.82	9.721	10^{-2}	1.5×10^5	15	512×128	0.67	600.6
10^7	13.82	9.721	10^{-2}	1.5×10^5	25	512×128	0.20	865.4
10^7	13.82	10.414	10^{-2}	1.5×10^5	10	512×128	3.07	416.1
10^7	13.82	10.414	10^{-2}	1.5×10^5	15	512×128	0.79	847.4
10^7	13.82	10.414	10^{-2}	1.5×10^5	20	512×128	0.20	1905
10^7	13.82	10.414	10^{-2}	1.5×10^5	25	512×128	0.27	1742
10^7	13.82	10.414	10^{-2}	1.5×10^5	35	512×128	0.17	2057
10^7	13.82	10.82	10^{-2}	1.5×10^5	15	512×128	1.84	583.0
10^7	13.82	10.82	10^{-2}	1.5×10^5	20	512×128	1.94	695.8
10^7	13.82	10.82	10^{-2}	1.5×10^5	25	512×128	0.36	1550
10^7	13.82	11.513	10^{-2}	1.5×10^5	15	512×128	4.52	1611

Notes: Stagnant lid models are not run all the way to statistical steady-state, and instead were stopped when it was clear that convection had entered a stagnant lid regime and would not return to a mobile lid. Velocities are averages over the model period when stagnant lid behavior began, but do not represent statistical steady-state velocities reached if the models were run longer.

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