Earth and Planetary Science Letters 331–332 (2012) 281–290

Contents lists available at SciVerse ScienceDirect

Earth and Planetary Science Letters

journal homepage: www.elsevier.com/locate/epsl

The conditions for plate tectonics on super-Earths: Inferences from convection models with damage

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ARTICLE INFO

Article history:
Received 4 November 2010
Received in revised form 19 March 2012
Accepted 22 March 2012
Available online 20 April 2012

Editor: T. Spohn

Keywords:
mantle dynamics
exo-planets
plate tectonics
damage theory

ABSTRACT

Numerical simulations of mantle convection with a damage-grainsize feedback are used to develop scaling laws for predicting conditions at which super-Earths would have plate tectonics. In particular, the numerical simulations are used to determine how large a viscosity ratio between pristine lithosphere and mantle (μ_l/μ_m) can be offset by damage to allow mobile (plate-like) convection. Regime diagrams of μ_l/μ_m versus the damage number (D) show that the transition from stagnant lid to mobile convection occurs for higher μ_l/μ_m as D increases; a similar trend occurs for increasing Rayleigh number. We hypothesize a new criterion for the onset of plate tectonics on terrestrial planets: that damage must reduce the viscosity of shear zones in the lithosphere to a critical value equivalent to the underlying mantle viscosity; a scaling law based on this hypothesis reproduces the numerical results. For the Earth, damage is efficient in the lithosphere and provides a viable mechanism for the operation of plate tectonics. We scale our theory to super-Earths and map out the transition between plate-like and stagnant-lid convection with a "planetary plate-tectonic phase" diagram in planet size-surface temperature space. Both size and surface conditions are found to be important, with plate tectonics being favored for larger, cooler planets. This gives a natural explanation for Earth, Venus, and Mars, and implies that plate tectonics on exoplanets should correlate with size, incident solar radiation, and atmospheric composition.

1. Introduction

Within the last decade, many terrestrial extra-solar planets have been discovered. A subset of these planets, approximately 1 to 10 times as massive as Earth, have been termed "super-Earths," because they are presumably analogous to our own planet (e.g. Rivera et al., 2005; Valencia et al., 2007b). One of the most important questions concerning super-Earths is whether they maintain conditions that potentially support life. Plate tectonics is probably necessary for providing the energy source for early chemosynthetic life (Martin et al., 2008) and for buffering atmospheric CO_2 through the negative feedbacks involving orogeny, erosion, weathering, and volcanism (e.g. Walker et al., 1981); thus determining whether super-Earths have plate tectonics is an important aspect of exoplanetary research. This topic is controversial, because recent studies have reached different conclusions about the propensity for super-Earths to have plate tectonics (Korenaga, 2010; O'Neill and Lenardic, 2007; Valencia and O'Connell, 2009; Valencia et al., 2007a). To provide some clarity to this debate, we present a self-consistent model of plate-generation to determine the tectonic regime of a planet.

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doi:10.1016/j.epsl.2012.03.028

1.1. Previous plate generation mechanisms

Plate tectonics on Earth arises from shear localization in, and weakening of, a highly viscous lithosphere that is part of the convecting mantle system (e.g. Bercovici, 2003; Bercovici et al., 2000; Tackley, 2000a). The lithosphere deforms by a variety of mechanisms; frictional sliding occurs near the surface where confining pressure is low, a complex rheology (including combined ductile and semi-brittle deformation) occurs at intermediate depths, and finally viscous flow occurs in the lower lithosphere where temperature is high. The undeformed lithosphere is strong below the weak frictional sliding layer (e.g. Karato, 2008; Kohlstedt et al., 1995); thus the generation of plate tectonics on a planet requires a mechanism capable of weakening the strong mid-lithosphere through convective stresses. However, the mechanism responsible for this weakening and localization is unknown and highly debated (e.g. Bercovici, 2003).

One simple deformation mechanism, used in many convection models, is a viscoplastic yield stress rheology (where the yield stress physically represents the strength of the mid-lithosphere) that reduces the viscosity of an area where the stress exceeds the yield stress (e.g. Moresi and Solomatov, 1998; Tackley, 2000b). This method produces a plate-like style of convection in numerical models (Foley and Becker, 2009; Tackley, 2000b; van Heck and Tackley, 2008). However, there is a large mismatch between experimentally determined lithospheric strength and convective stress, meaning that unrealistically low yield stress values are necessary to produce convection with a mobile plate-like surface (e.g. Moresi and Solomatov, 1998; Tackley, 2000b). In addition, this rheology only produces instantaneous shear localization. A material that has yielded will immediately regain its strength when stresses are no longer at the yield stress. On the Earth, however, memory of deformation in the form of dormant weak zones is a crucial aspect of plate tectonics (Gurnis et al., 2000).

1.2. Damage mechanics

To capture both dynamic localization and weak-zone memory, we use a grainsize feedback we refer to as grain-damage (e.g. Bercovici and Ricard, 2003, 2005; Bercovici et al., 2001; Landuyt et al., 2008). The damage mechanism is motivated by its effectiveness in the mid-lithosphere, its allowance of dormant weak zones, and the pervasiveness of peridotitic mylonites (deformed mantle rocks with significant grainsize reduction) in lithospheric shear zones (White et al., 1980). Damage theory generally refers to weakening of materials due to cracking, defects, and grainsize reduction and our formulation is based on interface thermodynamics (Bercovici and Ricard, 2003, 2005, in press; Bercovici et al., 2001; Landuyt et al., 2008; Ricard and Bercovici, 2009) as cracking is synonymous with storing energy on the newly formed surface (Griffith, 1921). While the formulation employed here focuses on grain reduction and associated weakening in ductile flow, the physics of simultaneous micro-crack formation and weakening in semi-brittle behavior is essentially the same in the damage formalism; thus we use one formulation to capture damage across the whole mid-lithosphere.

Our damage mechanism uses a feedback between deformation induced grainsize reduction and a grainsize dependent viscosity. As deformation reduces grainsize, viscosity decreases, causing more deformation. Numerical convection simulations show that damage theory is effective at making a plate-like style of convection by reducing the viscosity of the lithosphere (Landuyt and Bercovici, 2009; Landuyt et al., 2008). However, most previous plate generation studies using damage assume a low (4–5 orders of magnitude) viscosity ratio between the lithosphere and mantle (e.g. Landuyt and Bercovici, 2009; Landuyt et al., 2008) and are not applicable to exoplanets. Therefore we perform new numerical convection simulations and use the results to derive analytical scaling laws describing a new condition for the onset of plate tectonics on a planet. We use these scaling laws to develop a "planetary plate-tectonic phase diagram" which describes the influence of leading order and astronomically observable properties, like planet size and surface temperature, on the conditions for plate-like mantle convection.

2. Theory

Here we briefly discuss the key assumptions behind the theoretical formulation of our damage model (Bercovici and Ricard, 2003, 2005; Bercovici et al., 2001; Landuyt et al., 2008). The damage formulation is not a microphysical description of the mechanics of grainsize reduction and growth, but is instead a phenomenological law based on thermodynamic constraints and designed to capture the continuum physics of a self-weakening rheology.

We assume that grainsize reduction and grainsize sensitive creep occur simultaneously. In mantle rock, grainsize reduction is thought to occur in a dislocation creep regime, where creep occurs due to propagation of dislocations through the crystal lattice, while grainsize sensitive creep is thought to occur in a diffusional regime, where creep occurs by diffusion of material through the mineral grain or along the grain boundary. These regimes of deformation occur by distinct micro-mechanical mechanisms in separate domains of deformation space (depending on differential stress, temperature and grainsize) and therefore do not necessarily coexist in a single grain (Karato, 2008). However, a control volume of rock will contain a large distribution of grainsizes, such that dislocation creep and diffusion creep can occur simultaneously (e.g. large grains deform by dislocation creep and small grains by diffusion creep) (Bercovici and Karato, 2003; Ricard and Bercovici, 2009; Rozel et al., 2011). In natural two-phase or polymineralic materials like peridotite, deformation and damage to the interface between phases (e.g., olivine and pyroxene) combined with pinning effects allow damage, grain-reduction, and diffusion creep to co-exist (Bercovici and Ricard, in press); this leads to a state of small grain permanent diffusion creep, which is observed in natural peridotitic mylonites (Warren and Hirth, 2006).

We further assume that the average viscosity of the volume is controlled by the grainsize sensitive diffusion creep, though in reality the rheology will be controlled by whichever mechanism allows for the easiest deformation (e.g. Rozel et al., 2011). This assumption is supported by field observations in some shear zones (Jin et al., 1998; Mehl and Hirth, 2008; Warren and Hirth, 2006), and it allows us to focus on the grainsize feedback mechanism alone for this study.

2.1. Damage formulation

The viscosity is sensitive to grainsize and temperature as expected for diffusion creep or grain boundary sliding (Hirth and Kohlstedt, 2003):

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

where μ_n is a constant, E_v the diffusion creep activation energy ($E_v = 300$ kJ/mol (Karato and Wu, 1993)), T the temperature, R the universal gas constant, A the fineness, or inverse grainsize (Bercovici and Ricard, 2005; Landuyt et al., 2008), and A_{ref} the reference fineness. The constant m is usually set to 2 or 3 based on the mechanism of diffusion creep (Hirth and Kohlstedt, 2003; Karato and Wu, 1993). Specifically, diffusion along the grain boundary, or Coble creep, leads to $m = 3$ while diffusion through the grain, Nabarro–Herring creep, results in $m = 2$ (e.g. Evans and Kohlstedt, 1995). This difference is potentially important due to feedbacks in our damage model. We choose $m = 2$ as this produces a modest level of softening more consistent with a composite rheology, that combines dislocation and diffusion creep. We show later that our model with $m = 2$ produces a power law rheology, similar to that produced by a composite rheology with damage for $m = 3$ (Rozel et al., 2011).

Fineness is governed by the following evolution equation:

$$\frac{DA}{Dt} = \frac{f}{\gamma}\Psi - hA^p \tag{2}$$

where t is time, f is the damage partitioning fraction, which can vary from zero to one, γ the surface free energy, Ψ the deformational work, h the healing rate, and p a constant. Deformational work is defined as $\Psi = \nabla \underline{v} : \underline{\underline{\tau}}$, where $\underline{v}$ is the velocity and $\underline{\underline{\tau}}$ is the stress tensor (Bercovici and Ricard, 2005; Landuyt et al., 2008), see also Austin and Evans (2007).

The first term on the right side of Eq. (2) represents partitioning of a fraction (f) of deformational work towards creation of surface free energy by reducing grainsize and increasing fineness. The second term on the right side represents reduction of fineness due to normal grain growth, where, as in Landuyt and Bercovici (2009) and Landuyt et al. (2008), the constant p is set to 3 to reproduce the experimentally inferred dependence of grainsize on the square root