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3-D thermo-mechanical modeling of plume-induced subduction initiation

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ABSTRACT

Here, we study the 3-D subduction initiation process induced by the interaction between a hot thermo-chemical mantle plume and oceanic lithosphere using thermo-mechanical viscoplastic finite difference marker-in-cell models. Our numerical modeling results show that self-sustaining subduction is induced by plume–lithosphere interaction when the plume is sufficiently buoyant, the oceanic lithosphere is sufficiently old and the plate is weak enough to allow the buoyant plume to pass through it. Subduction initiation occurs following penetration of the lithosphere by the hot plume and the downward displacement of broken, nearly circular segments of lithosphere (proto-slabs) as a result of partially molten plume rocks overriding the proto-slabs. Our experiments show four different deformation regimes in response to plume–lithosphere interaction: a) self-sustaining subduction initiation, in which subduction becomes self-sustaining; b) frozen subduction initiation, in which subduction stops at shallow depths; c) slab break-off, in which the subducting circular slab breaks off soon after formation; and d) plume underplating, in which the plume does not pass through the lithosphere and instead spreads beneath it (i.e., failed subduction initiation). These regimes depend on several parameters, such as the size, composition, and temperature of the plume, the brittle/plastic strength and age of the oceanic lithosphere, and the presence/absence of lithospheric heterogeneities. The results show that subduction initiates and becomes self-sustaining when the lithosphere is older than 10 Myr and the non-dimensional ratio of the plume buoyancy force and lithospheric strength above the plume is higher than approximately 2. The outcomes of our numerical experiments are applicable for subduction initiation in the modern and Precambrian Earth and for the origin of plume-related corona structures on Venus.

1. Introduction

The subduction process is widely considered a key factor of plate tectonics. Despite its importance, many aspects of subduction remain poorly understood, and the most controversial questions involve where new subduction zones initiate and how this process proceeds. Some hypotheses have been proposed to answer the first, "where" question and include subduction nucleation at passive continental margins (Dickinson and Seely, 1979; McKenzie, 1977; Karig, 1982; Nikolaeva et al., 2010) and subduction initiation along pre-existing weakness zones containing transform/fracture zones (Uyeda and Ben-Avraham, 1972; Cooper et al., 1976; Hilde et al., 1977; Casey and Dewey, 1984), extinct mid-ocean ridges (Jones, 1971; Casey and Dewey, 1984), STEP (Subduction-Transform Edge Propagator) faults (Baes et al., 2011), and back-arc regions of mature subduction zones through subduction polarity reversal (McKenzie, 1969; Uyeda and Ben-Avraham, 1972; Cooper and Taylor, 1985; Kroenke, 1986). Testing the feasibility of these hypotheses requires answering the second "how" question. Previous studies have proposed several mechanisms for the subduction initiation process, such as lithospheric failure due to aging of the oceanic plate leading to gravitational instability (McKenzie, 1977), lithospheric weakening due to sedimentary load (Cloetingh et al., 1982, 1984, 1989), sedimentary load together with water weakening (Regenauer-Lieb et al., 2001), and external extensional regimes (Kemp and Stevenson, 1996) for the initiation of subduction at passive margins and far-field compression (Toth and Gurnis, 1998; Hall et al., 2003; Gurnis et al., 2004) for the initiation of subduction at pre-existing weakness zones. Based on 2-D numerical modeling, Ueda et al. (2008) proposed a new mechanism for the formation of new subduction zones resulting from hot

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mantle plume–oceanic lithosphere interaction. According to this scenario, under certain circumstances, a hot and positively buoyant plume overcomes the strength of the lithosphere and passes through it. Then, partially molten plume material spreads across the top of broken lithospheric segments (proto-slabs) and pushes them downward into the asthenosphere, eventually initiating a new subduction zone. In this scenario, the subduction initiation does not require any external forces or lithospheric heterogeneities, and the buoyancy of the plume and magmatism-induced plate weakening are the key factors. Following this study, Burov and Cloetingh (2010) showed that the interaction between a plume and the continental lithosphere can lead to the initiation of continental lithosphere subduction. They indicated that the rheological stratification of the visco-elasto-plastic lithosphere and its free surface are essential factors for subducting continental lithosphere to depths of 300–500 km following plume–lithosphere interaction. Gerya et al. (2015), using 3-D numerical models, proposed that plume–lithosphere interactions in the Archean led to the initiation of subduction and the onset of plate tectonics. The feasibility of plume-induced subduction has been recently supported by natural data from the Central American region, where plume-induced subduction initiation along the southern margin of Caribbean and northwestern South America in the Late Cretaceous (ca. 100 Ma) has been proposed (Whattam and Stern, 2014). The closely related newly published articles of Koptev et al. (2015a, 2015b) investigated the plume–lithosphere interaction in an extensional tectonic setting. Using 3-D numerical models, the authors showed that a plume rising beneath a pre-stretched lithosphere resulted in deflection of the plume by the cratonic keel. Consequently, the plume was channeled along one of the craton sides, leading to the development of coeval magma-rich and magma-poor rifts along opposite sides of the craton.

It is important to note that the previous numerical studies of both Ueda et al. (2008) and Burov and Cloetingh (2010) used 2-D thermo-mechanical models, whereas the process of plume–lithosphere interaction is intrinsically three-dimensional (e.g., Burov and Gerya, 2014; Gerya et al., 2015). Gerya et al. (2015) focused on investigating the conditions leading to the onset of plate tectonic in the Archean. Therefore, they set up experiments to simulate the early Earth, for instance by considering a hotter mantle and a thicker crust. In addition, the mantle plumes modeled by Gerya et al. (2015) were assumed to be purely thermal, whereas the influences of thermal-chemical plumes (e.g., Van der Lee et al., 2008; Sobolev et al., 2016) and compositional plume buoyancy (Ueda et al., 2008) on the 3-D plume-induced subduction initiation process have remained unexplored. In this study, we aim to further the work of Ueda et al. (2008) by using 3-D thermo-mechanical models to investigate the conditions leading to oceanic subduction initiation as a result of thermal-chemical mantle plume–lithosphere interaction in the modern Earth. We use numerical models to demonstrate how different physical parameters can result in different responses to thermal-chemical mantle plume–oceanic lithosphere interactions. In addition, we also investigate how extensional or compressional plate tectonic forces and heterogeneities in the lithosphere (e.g., Burov and Gerya, 2014, Koptev et al., 2015a, 2015b), such as faults and plateaus, can affect the subduction initiation process, as these factors are completely unexplored in previous studies.

We first describe the numerical model, which we use to investigate subduction initiation induced by interactions between a hot, thermally and chemically buoyant mantle plume and oceanic lithosphere. Then, we present the numerical results and a detailed analysis of the effects of different model parameters, such as the age of the lithosphere, the internal friction coefficient, and the composition, size, and excess temperature of the plume, on the results. Finally, we end with a discussion of the results and their implications for terrestrial and planetary tectonic processes.

2. Numerical model

2.1. Model geometry, boundary conditions and initial thermal field

We designed thermo-mechanical visco-plastic models of plume–lithosphere interactions using the I3ELVIS code, which is based on finite difference staggered grid and marker-in-cell methods with an efficient OpenMP-parallelized multigrid solver (Gerya and Yuen, 2007; Gerya, 2014). The model dimensions are 732 km $\times$ 200 km $\times$ 732 km, with a uniform rectangular grid of 245 $\times$ 101 $\times$ 245 nodes and 79 million randomly distributed Lagrangian markers. The coordinate system is a right-handed Cartesian system with the positive y-axis in the direction of depth. The model consists of an oceanic lithosphere, asthenosphere and an ellipsoidal or spherical plume at a depth of 125 km, which corresponds to the onset of plume–lithosphere interaction (Fig. 1a). Table 1 lists the material properties that are used in our numerical models. The oceanic crust consists of 5 km of gabbroic rock and 3 km of basalt. The lithospheric mantle and asthenosphere both have a dry olivine rheology (the different colors for the asthenospheric and lithospheric mantle in the figures of this paper are only for better visualization of the slabs). The plume is modeled using an ellipsoidal (or spherical) volume of hot, partially molten mantle rocks and varies in size, composition (its compositional buoyancy is defined relative to the ambient mantle and ranges from -20 kg/m^3 to $+300$ kg/m^3, Table 1 and Table 2 in Supplementary material) and temperature among the different models (Table 2 in Supplementary material), similar to the 2-D models of Ueda et al. (2008). The free surface boundary condition at the top of the crust is modeled by a 20-km-thick sticky air layer with a low density of 1 kg/m^3 and a viscosity of 10^{18} Pa s. All mechanical boundary conditions are free slip, except the lower boundary, which is defined as an open boundary (Ueda et al., 2008). The thickness of the oceanic lithosphere is defined based on the depth of the 1000 °C isotherm, which changes as a function of the cooling age.

The initial temperature profile in the oceanic lithosphere is calculated according to the prescribed lithospheric age (Turcotte and Schubert, 1982) which varies from 5 to 80 Myr (Table 2 in Supplementary material). The temperature in the asthenosphere is computed according to the adiabatic thermal gradient of 0.5 K/km. The constant elevated temperature within the plume varies between 1900 K and 2500 K in different experiments (Table 2 in Supplementary material).

2.2. Rheological model

The deformation mechanism in the models is based on visco-plastic rock rheology. The effective ductile creep viscosity is defined as:

$$\frac{1}{\eta_{effective}} = \frac{1}{\eta_{diffusion}} + \frac{1}{\eta_{dislocation}}$$

in which $\eta_{diffusion}$ and $\eta_{dislocation}$ are the effective viscosities for diffusion and dislocation creep, respectively, and are computed as:

$$\eta_{diffusion} = 1/2\frac{A_D}{\sigma_{cr}^{n-1}}exp\left(\frac{E+PV}{RT}\right)$$

$$\eta_{dislocation} = 1/2A_D^{\frac{1}{n}}exp\left(\frac{E+PV}{RT}\right)\varepsilon_{II}^{\prime\,\frac{(1-n)}{n}}$$

where P is the dynamic pressure, T is the temperature (in K), $\varepsilon'_{II} = \sqrt{\frac{1}{2}\varepsilon'_{ij}\varepsilon'_{ij}}$ is the second invariant of the strain rate tensor, σ_{cr}