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Dynamics of slab rollback and induced back-arc basin formation

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ABSTRACT

We performed a numerical study to understand the dynamical mechanisms of back-arc basin formation induced by slab rollback. To this end, we used two-dimensional numerical models of an integrated plate–mantle convection system with 410- and 660-km phase transitions. Retrograde slab migration occurs when the slab stagnates in the transition zone or when the deep section of the slab is vertical. In both cases, slab rollback occurs because the deep slab section obstructs the descending motion of the shallow slab section with an inclination. Buoyancy of the 660-km phase boundary acts as the obstructing force in the case of stagnant slab formation, and an anchoring force against the horizontal motion works similarly in the case of vertical slabs. To balance the horizontal component of the obstructing force, a suction force at the plate boundary pulls the overriding plate toward the ocean. Back-arc spreading is produced by means of slab rollback when the overriding plate with a weak area is fixed to the model boundary. The back-arc deformation becomes compressional when the overriding plate is freely movable despite trench retreat, because the wedge mantle flow viscously drags the overriding plate toward the trench. This implies that forces tending to actuate the overriding plate away from the trench are necessary to generate back-arc extension even when trench retreat is generated by slab rollback.

1. Introduction

The formation of back-arc basins is one of the distinctive characteristics of subduction zones. Dynamical mechanisms of back-arc basin formation are classified into the following three types (Heuret and Lallemand, 2005): (1) overriding plate motion away from the trench with the slab anchored to the deep mantle (Uyeda and Kanamori, 1979; Scholz and Campos, 1995), (2) slab rollback (migration in the direction opposite to the plate motion) driven by mantle flow pressure or asthenosphere injection (Tatsumi et al., 1990; Ricard et al. 1991; Doglioni, 1993; Flower et al., 2001), and (3) rollback of the gravitationally unstable slab (Elsasser, 1971; Molnar and Atwater, 1978; Garfunkel et al., 1986).

Analyses of the motions of the overriding plate and trench (Chase, 1978; Jarrard, 1986; Heuret and Lallemand, 2005) have shown that back-arc deformations are strongly influenced by the overriding plate motion. This is interpreted to mean that the overriding plate motion that is distant from the trench drives the back-arc extension. Active back-arc basins, however, often accompany a fast retreating slab (Heuret and Lallemand, 2005; Sdrolias and Müller, 2006). This suggests that slab rollback is an important driving mechanism of back-arc basin formation. Dependence of the slab rollback speed on the age of the subducting lithosphere, which is expected when backward motion is induced by negative buoyancy of the slab, is not observed (Heuret and Lallemand, 2005; Lallemand et al., 2005). Therefore, the pressure against the slab surface of the arc side caused by the mantle flow is often the preferred explanation for slab rollback (Heuret and Lallemand, 2005; Sdrolias and Müller, 2006). This model is also used to explain the concentration of back-arc basins in the western Pacific margins (Ricard et al., 1991; Doglioni, 1993; Flower et al., 2001).

Seismic tomography images (van der Hilst et al., 1991; Fukao et al., 1992, 2001, 2009) show that stagnant slabs distributed beneath the subduction zones often accompany back-arc basins (van der Hilst and Seno, 1993). Christensen (1996) and others (Čížková et al., 2002; Torii and Yoshioka, 2007; Tagawa et al., 2007a; Billen, 2008) showed the importance of trench retreat to stagnant slab generation. Nakakuki et al. (2010) pointed out that trench retreat reduces down-dip stress at the tip of the slab because the distance between the trench and the slab tip is extended, and thus promotes formation of the stagnant slab. Therefore, stagnant slab formation induced by slab rollback reduces the dip angle of the slab. In contrast, Lallemand et al. (2005) pointed out that the deep part of slabs with back-arc extension has a steeper dip angle on average than that with back-arc compression.

Backward slab motion has been realized in a self-consistent manner in several studies with two-dimensional (2-D) numerical

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models (Gurnis and Hager, 1988; Zhong and Gurnis, 1995; Enns et al., 2005; Capitanio et al., 2007; Tagawa et al., 2007b; Faccenna et al., 2009; Nakakuki et al., 2010), three-dimensional (3-D) numerical models (Stegman et al., 2006) and analog models (Kincaid and Olson, 1987; Schellart, 2005; Faccenna et al., 2007). Numerical models for back-arc basin formation driven by slab buoyancy have been developed in the past decade. For example, Hall et al. (2003) and Gerya et al. (2008) modeled the initiating subduction zone with an old lithosphere thrusting beneath a young oceanic lithosphere. The effects of the overriding plate on the back-arc deformation have been studied in a case with fixed velocity (Arcay et al., 2008) and with free mobility (Capitanio et al., 2010a). Clark et al. (2008) and Capitanio et al. (2010b) have reproduced the extension of a thin back-arc lithosphere by using 3-D and 2-D models with a viscosity jump between the upper mantle and lower mantle. Gerya and his collaborators (Gorczyk et al., 2007; Nikolaeva et al., 2008; Gerya and Meilick, 2011; Vogt et al., 2012) have focused on revealing the effects of magmatic products and hydration, including the fine lithological structure of the subduction zone at the ocean–ocean or ocean–continent boundary, on the formation of the back-arc spreading.

In this study, we aim to understand the dynamics of slab rollback and the accompanying back-arc basin formation under the condition where the subducting lithosphere interacts with 410- and 660-km phase transitions. We focus on deformation and force balance of the subducted and overriding lithospheres in the subduction zone at an ocean–continent boundary. The evolution of the subduction zone structure is numerically simulated from subduction initiation to slab penetration into the lower mantle. For this purpose, we construct 2-D models of an integrated plate–mantle convection system with freely movable lithospheres and a plate boundary driven self-consistently by an internal density anomaly (Nakakuki et al., 2010). Lithosphere-scale (a few to 100 km) deformation is calculated in the numerical mantle convection system on the whole-mantle scale (a few thousand to 10,000 km).

We first investigate the process of stagnant slab generation due to the interaction with the mantle transition zone as a mechanism of slab rollback (van der Hilst and Seno, 1993; Zhong and Gurnis, 1995). In this model, the strength of the arc lithosphere is treated as a varying parameter to examine the magnitude of the driving force. We then consider the dynamics of steeply inclined slabs to identify the slab rollback mechanism that is consistent with the relationship between the dip angle and the retreating motion of the subducted slab (Lallemand et al., 2005), and examine the effects of stiff slab lengths. Finally, we consider the effects of overriding plate mobility (Capitanio et al., 2010a; Nakakuki et al., 2010) on back-arc deformation to understand why back-arc extension is often associated with overriding plate motion distant from the trench (Chase, 1978; Jarrard, 1986; Heuret and Lallemand, 2005). We focus on the interaction of the subducted slab with a freely movable overriding plate via the wedge mantle flow.

2. Model settings and basic equations

2.1. Model configuration

The models are constructed based on the model in Nakakuki et al. (2010). The major difference from the previous model is the introduction of a fixed overriding plate to most cases except for one. In our model, the plate-like motion of the surface layer is generated without an imposed plate velocity. Several assumptions are made to produce the plate-like motion of the surface, as mentioned in Tagawa et al. (2007b). Details of the model settings are described in the Supplementary materials.

A schematic view of the model configuration is shown in Fig. 1. The fixed parameters are described in Table 1. We use an extended Boussinesq fluid model (Christensen and Yuen, 1985) in a 2-D rectangular box. Free-slip conditions are assigned to all boundaries so that the internal buoyancy forces spontaneously drive the motion of the lithospheres and the underlying mantle. Constant temperature is applied to the top boundary, and thermal insulation is assumed on the bottom and side boundaries. We consider back-arc deformation in the subduction zone with an overriding continental lithosphere about 100 km thick in the initial condition. The continental lithosphere includes an intrinsically buoyant layer that simulates a continental crust 34 km thick (Segment 5). To produce deformation in the overriding plate, reduction of the lithospheric strength and wedge mantle viscosity is introduced into the arc area (Segment 4). The descriptions of the variable model settings are found in Table 2. In cases with a fixed overriding plate (Cases 1–7), the low-viscosity zone is not introduced. Only Case 8 includes a "low-viscosity area," which provides freedom of the overriding plate motion in the top-right corner of the model (Segment 6). The viscosity of this zone is fixed at 10^{22} Pa s, which is close to the effective viscosity of the orogenic zone (Gordon, 2000).

The transition zone is modeled with two phase transitions of olivine series minerals at 410- and 660-km depths. The Clapeyron slope for the 410-km phase transition is set to be $+3$ or $+2$ MPa K^{-1}, and that for the 660-km phase transition is -3 or -2 MPa K^{-1} (Table 2). These values are consistent with those inferred from the depression of the 660-km discontinuity obtained by seismological

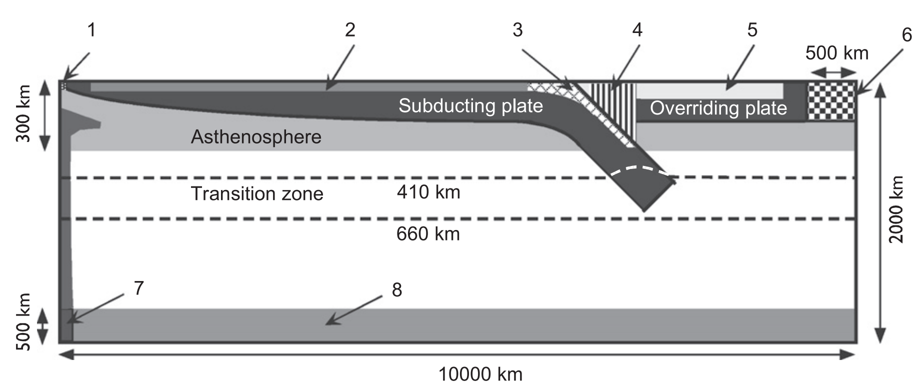

Fig. 1. Schematic illustration of model configuration. The dark segments in the uppermost area show the subducting and overriding plates. The light gray area shows the asthenosphere wherein the viscosity pressure dependence is larger than that in the layer below a depth of 300 km. The dashed lines show the phase transitions at depths of 410 and 660 km. The numbered areas represent the following. 1: Weak zone due to melting, 2: oceanic crust layer with history-dependent yielding, 3: fractured segment of 2, 4: weak arc area with the low yield strength and viscosity, 5: continental crust with intrinsic buoyancy, 6: low-viscosity area to allow free overriding plate motion (only for Case 8), 7: fixed-high-temperature area to generate upwelling plume, and 8: fixed-temperature layer.