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On the possibility of a folded crustal layer stored in the hydrous mantle transition zone

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ABSTRACT

Previous numerical studies of mantle convection focusing on subduction dynamics have indicated that viscosity (rheological) heterogeneity in the subducting plates and the surrounding mantle may have a primary effect on the behavior of subducting plates. Meanwhile, the existence of a low-viscosity layer (LVL) beneath the mantle transition zone (MTZ) has been suggested by several geodynamic studies. We investigated the effects of a rheologically weaker crustal layer in a hydrous (wet) MTZ with an underlying LVL on the behavior of subducting plates and determined the trace of crustal material using a numerical simulation model of subduction dynamics in three-dimensional regional spherical geometry. The combined effect of a rheologically weaker crustal layer and an LVL beneath the MTZ produces characteristic transient behavior of the crustal layer, whose trace shows a folded, loop-like feature in the wet MTZ along with a folded subducting slab in the LVL. This phenomenon is caused by mechanical interaction of the crustal layer with an upward positive bending force due to the main body of the subducting slab folding in the LVL and an enhanced downward negative buoyancy force due to an increase in density from the garnet to the post-garnet phase of crustal material. The transiently folded crustal layer behind the root of the stagnant slab is also observed. This result is consistent with seismological evidence suggesting a piece of subducted oceanic crust in the uppermost lower mantle beneath the subducting slab in the Mariana trench. Over half of the subducted crust is temporarily stored in the MTZ. However, this stored crustal material begins to penetrate into the lower mantle within 10 Myr, a relatively short timeframe relative to the overturn time of mantle convection. Eventually, most of the crustal material is entrained into the lower mantle, but a non-negligible portion remains stored in the MTZ for >50 Myr after the beginning of the accumulation of subducted crust in the MTZ.

1. Introduction

Seismic tomography models produce varying predictions of the behavior of subducted plates as they sink into the mantle transition zone (MTZ) (e.g., Fukao et al., 1992, 2001; Li et al., 2008; van der Hilst et al., 1991, 1997; Zhao, 2004; Zhao et al., 2013). While tomography techniques are robust in capturing seismic images of velocity anomalies, they have limitations for resolving detailed structures, including traces of the subducted oceanic crustal layer, associated with the mantle fabric distribution at the bottom of the MTZ. There remains substantial variation between the models in shorter wavelength features (Grand, 2002).

Many dynamic or semi-dynamic numerical models of mantle convection have not considered subducted oceanic crustal material (e.g., Čížková et al., 2007, 2002; Christensen, 1996; Han and Gurnis, 1999; Torii and Yoshioka, 2007; Yoshioka and Naganoda, 2010). There has been debate over whether crustal material becomes trapped above the ringwoodite to perovskite + magnesiowüstite phase decomposition boundary at a depth of 660 km (hereafter, the "660-km phase boundary") due to its chemical buoyancy and the density inversion between the surrounding mantle and the oceanic crust at the phase boundary (Ringwood and Irifune, 1988). The crustal material is lighter than the surrounding mantle, which is a higher-pressure phase.

However, a few numerical studies have included the crustal layer. Richards and Davies (1989) and Gaherty and Hager (1994) performed numerical simulations of two-dimensional (2D) mantle convection and suggested that separation of the crustal component (eclogite) from the subducting lithosphere (bulk peridotite) should not occur at the 660-km phase boundary. van Keken et al. (1996) conducted 2D numerical experiments to test the influence of the density and viscosity of the crust while descending into the lower mantle. They concluded that if a thin, weak layer exists between the strong garnetite crust and the cold slab interior, it can effectively decouple the crust from the main body of the slab and the

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lighter crust can rise, leading to a garnet-enriched MTZ. Christensen (1997) conducted a 2D semi-dynamic simulation of plate subduction and concluded that the compositional buoyancy of crustal material has negligible effect in determining the behavior of stagnant slabs associated with old plates (i.e., 100 Myr) in the MTZ.

Folding of subducting slabs in the Mariana trench has been imaged in the MTZ using seismic tomography models (e.g., Widiyantoro et al., 2000). These features were also identified by some numerical models of mantle convection (Christensen, 1997; van Hunen and van den Berg, 2008). van Hunen and van den Berg (2008) showed variations in the tectonic style of subduction with changing mantle temperature using a series of 2D numerical mantle convection models of the crust and a highly viscous harzburgite layer. They determined that under present-day mantle temperatures, crustal separation from the buckled subducting plate in the MTZ is not produced by varying the yield stress, the coupling strength of a fault located between the subducting and overriding plates, the viscosity of the harzburgite layer, etc.

These numerical studies suggest that the chemical buoyancy of crustal material does not appear to control the behavior of subducting plates at the bottom of the MTZ, whereas viscosity (or rheological) heterogeneity in the subducting plates and the surrounding mantle may be the primary influence.

Recent high-pressure experiments in mineral physics suggest deep water transport by subducting plates (e.g., Ohtani et al., 2004) and a rheological contrast between the garnet-rich crustal material and the surrounding mantle (olivine-rich) under hydrous conditions in the MTZ (Katayama and Karato, 2008). Yoshida et al. (2012) carried out a three-dimensional (3D) numerical simulation and suggested that there is a substantial difference in the behavior of a subducting plate and the trace of crustal material in their models under dehydrated ("dry") and hydrous ("wet") MTZ conditions. Under wet conditions, the crustal material, which is weaker (less viscous) than the surrounding peridotite (i.e., subducting plate) in the MTZ, is temporarily stored at the bottom of the MTZ.

There is controversy surrounding viscosity variations in the MTZ and the uppermost lower mantle (Karato, 2008). Postglacial rebound studies have suggested that no significant viscosity decrease occurs at the 410-km seismic discontinuity even under wet MTZ conditions; rather, there may be a viscosity increase with depth, likely due to pressure dependence (Peltier, 1998). Čadek et al. (1997), Kido and Čadek (1997) and Kido and Yuen (2000) suggested a low-viscosity layer (LVL) just beneath the MTZ based on their geoid inversion study. The LVL is considered to be of thermal origin—a thermal barrier at the 660-km boundary arising from the negative Clapeyron slope induces a high-temperature thermal boundary layer beneath the MTZ. Numerical studies of mantle convection have suggested that the secondary upwelling plumes originated from the MTZ are attributed to the LVL (i.e., "second asthenosphere") (Cserepes and Yuen, 1997; Cserepes and Yuen, 2000; Cserepes et al., 2000). They concluded that the influence of the second asthenosphere is important in inducing convection layering at 660 km and the "mid-mantle plumes" that have no root in the deep lower mantle. More recently, Yoshioka and Naganoda (2010) concluded that a subducting slab tends to stagnate if an LVL exists just below the 660-km boundary, because viscous resistance force between the slab and the LVL is reduced there.

Recently, we developed a 3D numerical simulation model using regional spherical geometry and carried out a series of simulations focusing on the behavior of subducted oceanic plates and the role of the crust in the MTZ. We concluded that under wet conditions, the subducting plate tends to stagnate, with a maximum lateral extent of over 1000 km, when the viscosity increase at the 660-km phase boundary is larger than a factor of ~60 and the viscosity and density contrasts between the crustal material and the surrounding peridotite permit "plate segmentation" (Yoshida et al., 2012). However, quantitative analysis to determine how much of the crustal material stagnates in the MTZ and how much penetrates into the lower mantle has not yet been conducted.

In the present study, we examined the effects of a rheologically weaker crustal layer in the wet MTZ and an underlying LVL on subduction dynamics, i.e., the behavior of the subducting plate and determined the trace of crustal material using our recently developed 3D simulation code.

2. Model setup

2.1. Overview of the numerical model

The mantle was numerically modeled as an incompressible Boussinesq fluid with an infinite Prandtl number. A simulation code, ConvRS, was used to simulate mantle convection using particle-in-cell, finite-volume (FV) discretization on a staggered grid (e.g., Yoshida, 2008a,b; Yoshida et al., 2012). The computational domain of mantle convection was confined to 3D regional spherical-shell geometry along spherical polar coordinates (r, θ, φ) with a thickness of 1000 km and a lateral extent of $10^\circ \times 40^\circ$ (i.e., $\sim$1112 km $\times$ $\sim$4448 km at the top surface boundary) in the θ and φ directions (Fig. 1a). The numbers of FVs used were 128 (in r) $\times$128 (in θ) $\times$512 (in φ), resulting in a numerical resolutions of $\sim$8.7 km in the horizontal directions and $\sim$7.8 km in the radial direction.

Following Yoshida et al. (2012), the top surface boundary of the convecting domain was divided into two regions with overriding and subducting (i.e., oceanic) plates. The subducting plate initially had a lateral extent of $7.5^\circ \times 20^\circ$ in the θ and φ directions. The trench was set to $\varphi = 180^\circ$ for the subducting plate. To allow the subducting plate to freely extend latitudinally in the process of subduction, a "quasi-transform fault" was established at the northern boundary of the subducting plate ($\theta = 82.5^\circ$).

To simulate a subducting plate at a trench, we applied our recently developed semi-dynamic model of the subduction zone (Yoshida et al., 2012). The overriding and subducting plates were given the velocities V_o ($\equiv V_{os}\sin\theta$) and V_s ($\equiv V_{ss}\sin\theta$), where V_{os} and V_{ss} are the surface velocities of the overriding and subducting plates at $\theta = 90^\circ$, respectively (see supplemental materials). The velocity of trench retreat is given by $V_{tr}\sin\theta$, where V_{tr} is the surface velocity of trench retreat at $\theta = 90^\circ$ (Fig. 1b). The flow was imposed *a priori* at the top surface boundary and in a small region around the trench ("boundary region," see Fig. 1b) by maintaining mass conservation at each time-step (see supplemental materials for details). The dip angle of the subducting plate was fixed at 45° in the boundary region (Fig. 1b).

To simulate plate segmentation (Tajima et al., 2009; Yoshida et al., 2012), we introduced narrow weak (low-viscosity) fault zones (WFZs) on the top of the subducting plate accounting for the fault displacements associated with large seismic ruptures (Fig. 1a and b). These WFZs presumably correspond to the fault boundaries of large subduction earthquakes. In the present model, three WFZs were initially located along the subduction direction with a width of 0.16° (=$\sim$17 km) and a depth of $\sim$47 km (Fig. 1b).

2.2. Governing equations and parameters

In formulating the basic equations that govern mantle convection, the length L, velocity $\boldsymbol{v}$, stress (or pressure) $\boldsymbol{\sigma}$, time t, and temperature T are non-dimensionalized as follows:

$$L = r_1 L', \quad \boldsymbol{v} = \frac{\kappa_0}{r_1}\boldsymbol{v}', \quad \boldsymbol{\sigma} = \frac{\eta_0 \kappa_0}{r_1^2}\boldsymbol{\sigma}', \quad t = \frac{r_1^2}{\kappa_0}t', \quad T = \Delta T T' \qquad (1)$$

where r_1 denotes the radius of the Earth; κ_0, the reference thermal diffusivity; η_0, the reference viscosity; and ΔT, the temperature