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Influence of geometry and eclogitization on oceanic plateau subduction

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ABSTRACT

Subduction is driven by the negative buoyancy of the slab, which depends on both the temperature and composition of the lithospheric plate. In the case of subduction of an oceanic plateau, larger thickness of the lower-density crust and harzburgite layers can locally decrease the negative buoyancy available to drive subduction. However, the ability of a plateau to have an impact on subduction depends on the total buoyancy of the slab, and therefore also depends on size of the plateau and the basalt to eclogite transition, which substantially increases the density of the crust. Using 2-D numerical finite element models of subduction, we investigate the role of eclogitization and oceanic plateau size (thickness and length) in the process of oceanic plateau subduction. Model results show that eclogitization of the crust substantially increases the chances that a plateau will be subducted because the extra buoyancy of larger plateaus is lost through the phase transition. For kinematically driven models, all plateaus are subducted regardless of the thickness or width. Extrapolation of these results to 3-D geometry shows that sufficient slab buoyancy is available to sustain subduction for narrow plateaus with slab widths 6–7 times the plateau width (depending on plateau thickness and length). In dynamically driven models, subduction of oceanic plateaus of various sizes can substantially slow subduction, and large plateaus (>200 km long and >25 km thick) can lead to break-off of the slab and underplating of the plateau.

1. Introduction

Subduction relies on dense material sinking into the mantle and pulling the surface plate behind it (e.g., Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Funiciello et al., 2003a). Therefore, it is expected that the arrival of a buoyant structure, such as an oceanic plateau, into a subduction zone would resist subduction and may even cause subduction to fail (slab break-off). Normal oceanic crust is about 7 km thick, underlain by an ~ 18 km thick harzburgitic layer formed as the residue of melting at the mid-ocean ridge (Oxburgh and Parmentier, 1977). Both the basaltic crust with a density of 3000 kg/m^3 and the harzburgitic layer with a density of 3330 kg/m^3 are more buoyant than the underlying mantle, which has a density of 3400 kg/m^3 (Fig. 1a). However, regions of thicker oceanic crust, including oceanic plateaus, aseismic ridges and seamounts, formed by mantle plumes (Coffin and Eldhom, 1994) or intra-plate volcanism (Conrad et al., 2011) are not uncommon. As the oceanic crust is produced by the partial melting of the mantle, a thicker oceanic crust has to be produced by the partial melting of a higher volume of mantle and thus leads to a proportionally thicker harzburgite layer (Oxburgh and Parmentier, 1977) (Fig. 1a).

Early consideration of the feasibility of subducting an oceanic plateau focused primarily on the local isostatic balance of buoyancy forces within the trench and concluded that features with buoyant crust thinner than 15–17 km would subduct, while features with crust thicker than 30 km would not (Cloos, 1993). In recent years seismic and drill-core observations of oceanic plateau structure show that while some thin plateaus have escaped subduction (e.g., Caribbean Plateau, ~ 15 km, Kerr et al., 1999; Mann and Taira, 2004), others have not (e.g., Yakutat Plateau, 25 km, Bruns, 1985; Brocher et al., 1991; Sea of Okhost, Mann and Taira, 2004), and even the thickest plateau, Ontong-Java (30–40 km thick, Gladczenko et al., 1997; Miura et al., 2004; Mann and Taira, 2004), is being subducted to depths of at least 200 km (Phinney et al., 2004).

Even in cases where buoyant features are subducted they can continue to modify the subduction process within the shallow asthenosphere. Around 10% of modern subduction zones have a shallow-dipping or flat slab segment (Gutscher et al., 1999) and subduction of thickened oceanic crust is often put forward to explain such anomalous segments. For example, the Pacific plate slab has a dip of 5–10° for more than 500 km from the trench where the Yakutat Plateau is subducting beneath southern Alaska (Ratchkovski and Hansen, 2002; Gudmundsson and Sambridge, 1998; Ferris et al., 2003). However, not all buoyant features are

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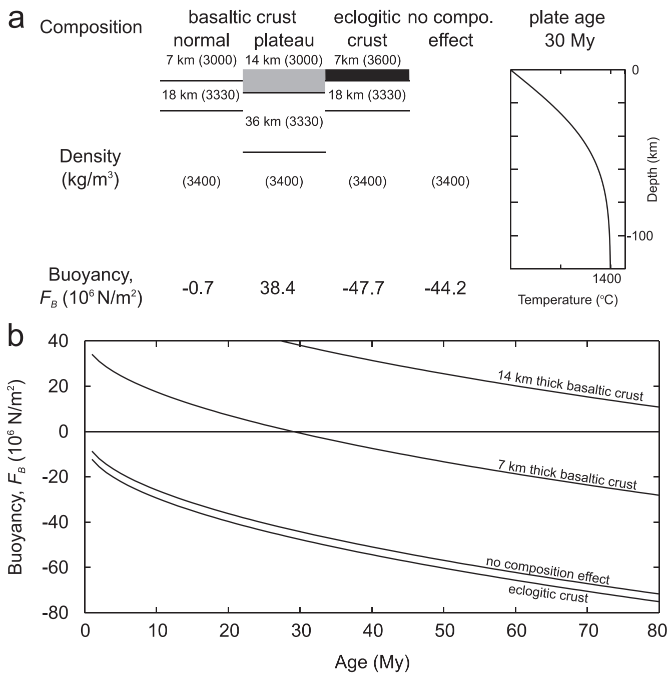

Fig. 1. Isostatic buoyancy of the lithosphere. (a) Isostatic column for normal and overthicken basaltic crust, both with and without eclogitic crust for a plate age of 30 Myr and temperature given by a half-space cooling model. Net buoyancy force per unit area is given at the bottom of profile for a 100 km thick column (where $F_b = g\int_y \rho(1-\alpha T)-\rho_0(1-\alpha T_0)\, dy$ with $\rho_0 = 3400$ kg/m^3 and $T_0 = 1400$ °C). (b) Comparison of buoyancy force for 100 km thick column as a function of plate age for normal oceanic crust before (7 km basaltic layer) and after the eclogite reaction (eclogitic crust), an oceanic plateau (14 km basaltic layer), and neglecting compositional layering (no composition effect).

associated with flat slabs (e.g., Louisville Ridge subducting in the Tonga-Kermadec Trench, Iquique Ridge in the Central Chile trench; Lallemand et al., 2005; Tassara et al., 2006).

In addition to thickness of the crust and residual layers, the net buoyancy of a subducting slab also depends on: (1) plate age as the lithosphere cools and becomes denser, (2) the transition of basalt to eclogite, which starts at a pressure of ∼ 15 kbar and a temperature of ∼400 °C (e.g., Hacker et al., 2003; Bousquet et al., 2005; Yamato et al., 2007a), and (3) the length of slab available to drive subduction. Normal oceanic plate becomes negatively buoyant after ∼30 Myr, while over thickened crust (e.g., 14 km) has a positive buoyancy force (Fig. 1b). However, as the plate subducts the basaltic crust will convert to eclogite (with a density of 3600 kg/m^3) at a pressure of around 15–30 kbar depending on temperature (Hacker, 1996). Once the crust is fully converted to eclogite, the plate has negative buoyancy regardless of plate age (Fig. 1b).

Comparison of the plate buoyancy force following conversion to eclogite (e.g., -47.6×10^6 N/m^2 at 30 Myr) to the buoyancy force due to temperature alone (no compositional effect: e.g., -44.2×10^6 N/m^2 at 30 Myr) shows that the negative buoyancy of the eclogite layer is compensated by the positive buoyancy of the harzburgite layer (Fig. 1b). In other words, after the eclogite reaction takes place there is no net compositional contribution to the slab buoyancy. This is true regardless of crustal thickness because the thickness of the residual harzburgite layer is proportional to the thickness of the basaltic crust (Oxburgh and Parmentier, 1977). One exception to this is if conversion to eclogite is incomplete (partial eclogitization), in which case a plate with thickened crust will remain buoyant and can lead to slab flattening beneath the overriding plate (van Hunen et al., 2002a, 2002b, 2004).

Finally, in order for a region of thickened crust to interfere with the subduction process its positive buoyancy must be greater than the total negative buoyancy of the slab in the upper-most mantle. Therefore, it is not only the thickness of the buoyant feature, but also its length (and width in three dimensions) that will determine if the subduction is slowed or even completely halted by subduction of this feature. In this case, the process of eclogitization may limit the buoyancy effects of long features.

The aim of this paper is to present two-dimensional (2-D) numerical models of oceanic plateau subduction to investigate the influence of plateau geometry (thickness and length) and eclogitization on the behavior of the slab. We present two types of geodynamic simulations: (1) models with an imposed convergence rate to model the case in which adjacent plateau-free regions of the plate are able to maintain subduction as a buoyant feature enters the subduction zone and (2) models with dynamic boundary conditions in which subduction is only driven by negative slab buoyancy. Furthermore, we extrapolate these 2-D model results into 3-D in order to place constraints on the conditions necessary to modify subduction rate and slab behavior, or even cause subduction to fail (slab-break off).

2. Methods

2.1. Equations

To model an oceanic plate with an embedded oceanic plateau subducting beneath another oceanic plate of the same age in two dimensions (Fig. 2), we solved the classic fluid mechanics equations using the Boussinesq approximation for incompressible viscous flow

$$\nabla\cdot\sigma+\mathbf{f}=0 \tag{1}$$

$$\nabla\cdot\mathbf{v}=0 \tag{2}$$

$$\frac{\partial T}{\partial t}+\mathbf{v}\cdot\nabla T=\kappa\nabla^2 T \tag{3}$$

where $\sigma_{ij}=-P\delta_{ij}+2\eta_{ef}\dot{\epsilon}_{ij}$ is the stress tensor defining the constitutive relation with $\dot{\epsilon}_{ij}$ being the strain rate tensor and η_{ef} the effective viscosity. P is defined as the second invariant of the stress tensor, and can be expressed as the sum of the lithostatic pressure and the dynamic pressure, resulting from the viscous flow, $\mathbf{f}=\rho(C)(1-\alpha T)g\boldsymbol{\delta}_{\mathbf{yy}}$ is the force due to density variations related to temperature (T) and composition (C). The other parameters are the velocity vector $\mathbf{v}$, the time t, and the thermal diffusivity κ. Internal heat sources and shear heating are not included.

2.2. Domain size and initial conditions

We present two sets of models: kinematic models with a constant horizontal velocity imposed on the subducting plate, and dynamic models with no imposed subduction velocity (Fig. 2a). For both sets of models a dipping weak crustal layer (described below), with initial dip increasing from 0° to 30° from the surface to 100 km depth, defines the boundary between the subducting and overriding plates at x=500 km (Fig. 2a). The initial shape of the dipping section of the weak crustal layer is in agreement with the average observed dip angle of slabs (32°) for depths of 0–125 km (Lallemand et al., 2005). The overriding plate is fixed.

The initial temperature condition for the subducting and overriding plates is given by a half-space cooling model defined by the age of the plate (with a thermal diffusivity of 0.8×10^{-6} m^2/s; Parsons and Slater, 1977). For the kinematic models, the initial