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Effects of temporal plume–slab interaction on the partial melting of the subducted oceanic crust

Chungwan Lim, Changyeol Lee *

Faculty of Earth and Environmental Sciences, Chonnam National University, Gwangju 500-757, Republic of Korea

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ABSTRACT

The plume–slab interaction facilitates the partial melting of the subducted oceanic crust in subduction zones and explains the pulse-like adakite eruptions in the Abukuma region, northeastern Japan from approximately 16 to 14 Ma, as well as the southwest-to-northeast migration of the adakite eruptions from southeastern China to southwestern Japan during the Cretaceous. Despite the potential implication of the plume–slab interaction to arc magmatism, the effects of the plume parameters on the partial melting of the subducted oceanic crust have not been quantitatively evaluated. Thus, we ran two-dimensional kinematic–dynamic numerical subduction experiments to evaluate the effect of the duration and thickness of the injected plume blob into the mantle wedge on the slab surface temperature. Along with the consideration of the injection of the plume blob, we consider diverse convergence rates and ages of the subducting slab with slab dips of 30° and 45°. Our model calculations show that the plume–slab interaction is a promising process that increases the slab surface temperature up to around 60 °C, resulting in the partial melting of the subducted oceanic crust even though the plate age is 100 My (million years). Increases in the duration and thickness of the plume blob positively increase the peak temperatures of the slab surface. Increases in the convergence rate and age of the subducting slab almost systematically increase and decrease the slab surface temperature, respectively, which indicates that the net temperature contribution of the plume–slab interaction is similar regardless of the subduction parameters. We found that the contribution of time-evolving subduction parameters on the plume–slab interaction can be approximated using the subduction parameters of approximately 5 My before the time of the partial melting of the subducted oceanic crust. Our model calculations facilitate the first-order analysis of the plume–slab interaction in the subduction zones of which subduction parameters are not consistent with the observed adakite.

1. Introduction

For decades, arc volcanism has been extensively studied in diverse fields including geochemistry, geophysics, and petrology. Among the numerous studies related to arc volcanism, partial melting of the subducted oceanic crust has been focused upon because of its implications for Archean continental growth (Kelemen et al., 2003b) and the special arc volcanic rocks such as adakites (e.g., Defant and Drummond, 1990; Kelemen et al., 2003b; Yogodzinski and Kelemen, 1998). Although the causes for the partial melting of the oceanic crust have been debated, ridge subduction of the very young oceanic plate (<25 My) has been accepted as a major cause (e.g., Defant and Drummond, 1990; Maruyama et al., 1997; Peacock, 1996).

Introduction of a numerical model allows quantitative evaluations of the thermal structure of the subduction zone, essential for evaluating the thermal environments for the partial melting of the subducted oceanic crust. Early numerical model studies using constant mantle viscosity have shown that the partial melting of the oceanic crust can occur when the age of the incoming oceanic plate is younger than 25 My and/or very large frictional heating along the slab–wedge interface (Peacock, 1996). Sophisticated numerical model experiments including realistic mantle rheology (diffusion and/or dislocation creep) and/or high-mesh resolutions have shown that the partial melting of the subducted oceanic crust can occur in incoming plates of moderate age (~50 My) such as the Aleutians (Kelemen et al., 2003a; Lee and King, 2010). Along with advances in numerical modeling, laboratory experiments on the partial melting of the subducted oceanic crust (sample: the synthetic basalt as a form of powder or glass) conducted using multi-anvil apparatuses have shown that the

* Corresponding author.
E-mail address: changyeol.lee@gmail.com (C. Lee).

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solidus of the subducted oceanic crust ranges from approximately 730 to 780 °C, much lower than that of the anhydrous (dry) oceanic crust (Kessel et al., 2005; Schmidt and Poli, 1998). This indicates that the partial melting of the subducted oceanic crust, which is responsible for adakites, can be evaluated through interdisciplinary studies including geochemical, geophysical and geological studies.

However, the numerical model experiments using the realistic mantle rheology and wet solidus of the subducted oceanic crust have not explained some observed adakites in the subduction zones such as Mariana, Tonga and Java-Sunda (Kim and Lee, 2014). This implies that other possible mechanisms exist, which result in the partial melting of the subducted oceanic crust in the subduction zones. Among the other mechanisms resulting in the partial melting of the subducted oceanic crust, the plume–slab interaction is suggested as a possible candidate. For example, the Mid-Miocene Abukuma adakite had erupted for ~2 My from approximately 16 to 14 Ma: the localized pulse-like eruption of the adakite (diameter of the erupted region: ~30 km) in the Abukuma region, northeastern Japan (Yamamoto and Hoang, 2009). Lee and Lim (2014) suggested that the pulse-like eruption of the Abukuma adakite was a consequence of the pulse-like plume–slab interaction through the injected plume blob from the subslab hot mantle. Another example of a potential plume–slab interaction might be the southwest-to-northeast migration of the peak timing of the Cretaceous adakite from southeastern China to northeastern Japan. The migration of the adakite eruptions has been considered to be related to the same directional migration of the ridge subduction (Kinoshita, 1995; Maruyama et al., 1997), but recent plate reconstruction model does not show the ridge subduction (Gurnis et al., 2012; Sdrolias and Müller, 2006), posing a problem for explaining the southwest-to-northeast migration of the adakite eruptions. A recent study indicates that the apparent migration of the peak timing of the adakite magmatism can be explained by the plume–slab interaction (Kim et al., 2015; Lee and Ryu, 2016), thus this interaction is a potential consideration to explain the 'tricky' partial melting of the subducted oceanic crust in the subduction zones. Although it is not designed to explain the partial melting of the subducted oceanic crust, the bimodal eruptions of the rhyolite and basalts in the Yellowstone region, western USA are interpreted as a consequence of a plume–slab interaction (Kincaid et al., 2013).

Although the plume–slab interaction potentially implies the partial melting of the subducted oceanic crust, the effects of plume parameters, such as duration and thickness, on the partial melting of the subducted oceanic crust, have not been quantitatively studied. In this study, we quantitatively evaluated the effects of plume parameters (duration and thickness of the plume blob injected into the mantle wedge) on the partial melting of the subducted oceanic crust. With the varying plume parameters, we considered diverse convergence rate and age of the incoming plate from 5 to 20 cm/y and 40 to 100 My, respectively, using slab dips of 30° and 45°. Through a series of numerical model experiments, we developed a guideline for geologists who want to evaluate whether a plume–slab interaction exists in the subduction zones.

2. Numerical models

Previous studies indicate that the viscous dissipation occurred in the subducting slab considerably affects the thermal structure of the subducting slab (e.g., Lee and King, 2009; So and Yuen, 2015). However, because the purpose of this study is to evaluate the sole effect of the plume parameters on the slab surface temperatures, essential for the evaluation of the partial melting of the subducted oceanic crust, we employed the incompressible Boussinesq approximation which neglects the viscous dissipation. Two-dimensional numerical kinematic–dynamic subduction models are similar to those used by our previous studies (Kim and Lee, 2014; Lee and King, 2009; Lee and Lim, 2014). Because the mantle behavior in the mantle wedge is dominated by the corner flow created by the viscous coupling between the subducting slab and overlying mantle wedge, the buoyancy of the mantle is neglected (e.g., Kneller and van Keken, 2008; Lee and King, 2010; Wada and Wang, 2009). Thus, the governing equations used in our model calculations are described as follows:

$$0 = \nabla \cdot \vec{v} \quad \text{continuity equation,} \tag{1}$$

$$0 = -\nabla P' + \nabla \cdot \bar{\tau} \quad \text{momentum equation,} \tag{2}$$

$$\rho_c C_p \frac{DT}{Dt} = \nabla \cdot (k \nabla T) + \rho_c H \quad \text{energy equation,} \tag{3}$$

where $\vec{v}$ is the velocity, P' is the dynamic pressure, $\bar{\tau}$ is the deviatoric stress tensor, ρ_c is the density, C_p is the heat capacity at constant pressure, t is the time, T is the temperature, k is the thermal conductivity, and H is the rate of radiogenic heat production. All of the material parameters used are kept constant and described in Table 1.

For the model calculations, we used slab dips of 30° and 45° for the kinematically subducted slab by imposing the constant convergent rate on the slab domain (Fig. 1). The overlying plate was assumed to be a 35-km-thick fixed layer, and a no-slip boundary condition was implemented on the bottom. A stress-free boundary condition was implemented along the boundaries enclosing the back-arc mantle and it allowed dynamic mantle inflow/outflow for the mantle wedge. To consider the buoyant serpentinites, the mantle flow at the corner of the mantle wedge is isolated, which was considered in our previous studies (Kim and Lee, 2014; Lee and Lim, 2014). Radiogenic heat productions of 3.80×10^{-11} W/kg for the upper crust and 7.38×10^{-12} W/kg elsewhere were used (Turcotte and Schubert, 2002). The potential temperature of the mantle is set to 1350 °C and a linear mantle adiabat of 0.35 °C/km is added to this potential temperature.

For the rheology of the mantle wedge, we introduced the composite viscosity of diffusion and dislocation creep using the dry olivine rheology constrained by Karato and Wu (1993). The mathematical form used in our model calculations are described as follows:

$$\eta_{comp} = \left(\frac{1}{\eta_{diff}} + \frac{1}{\eta_{disloc}} \right)^{-1} \quad \text{composite viscosity,} \tag{4}$$

Table 1
Model and rheological parameters.

Density (kg/m^3)	3300
Surface temperature (K)	273
Temperature contrast (K)	1420
Heat capacity (J/kg K)	1200
Thermal conductivity (W/m K)	3.96
E_{diff} (J/mol)	300×10^3
E_{disloc} (J/mol)	540×10^3
V_{diff} (m^3/mol)	6.0×10^{-6}
V_{disloc} (m^3/mol)	1.5×10^{-5}
A_{diff} (m$^{2.5}$/Pa s)	6.1×10^{-19}
A_{disloc} (s$^{1.5}$/Pa$^{3.5}$)	2.4×10^{-16}
d_g (m)	1.0×10^{-3}
m (·)	2.5
n (·)	3.5
R (J/mol)	8.314

E: activation energy, V: activation volume, A: prefactor, d_g: grain size, m: grain size exponent, n: stress exponent, and R: gas constant.