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Dike propagation driven by melt accumulation at the lithosphere–asthenosphere boundary

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ABSTRACT

This study examines the combined effects of melt migration by porous flow in the asthenosphere and dike propagation in the lithosphere. Melt collecting at the base of the lithosphere forms a decompacting boundary layer (DBL), in which the overpressure is sufficient to nucleate dikes that propagate buoyantly upward into the lithosphere. The asthenosphere melt flux determines the excess pressure and melt accumulation rate in the DBL, which together with the state of lithospheric stress, control dike growth rate, dike recurrence interval and the height to which dikes propagate. The vertical propagation and subsequent freezing of melt filled dikes heats and thins the lithosphere.

Our model couples fundamental aspects of dike propagation and porous flow that are commonly treated separately. Our model allows us to estimate conditions under which vertically propagating dikes can thin the lithosphere, given a melt flux determined by the rate of melt production in the asthenosphere. The model also provides an estimate of the amount of melt present at the base of the lithosphere. We find that a steady state high-porosity boundary layer at the lithosphere–asthenosphere boundary, with a melt fraction about 2.5–4 times higher than the asthenosphere melt fraction. Diking occurs at melt fractions much less than the disaggregation limit, so dikes are only a few km tall and about 1 cm wide. Though dikes are small, their recurrence on the order of days can lead to lithosphere erosion rates on the order of a few km/Myr with a melt fraction of a few percent at the base of the lithosphere. The steady state boundary layer melt fraction is controlled by differences in lithospheric stress state and or asthenosphere melt flux, indicating that seismic discontinuities associated with melt accumulation at the lithosphere–asthenosphere boundary should vary systematically with variations in asthenosphere melt generation and tectonic setting.

1. Introduction

1.1. Motivation and background

The near global observation of a sharp drop in seismic velocity near the base of the lithosphere (e.g. Gaherty et al., 1996; Tan and Helmberger, 2007; Rychert and Shearer, 2009; Yuan and Romanowicz, 2010; Fischer et al., 2010) has been attributed in some locations to melt accumulating beneath a solidus isotherm (e.g. Kawakatsu et al., 2009; Schmerr, 2012). Accumulation of melt at the lithosphere–asthenosphere boundary (LAB) has important consequences for the transport of melt into the lithosphere by diking, a significant factor in creation of divergent plate boundaries (e.g. Buck, 2004; Bialas et al., 2010). In this study, we focus on vertical dike propagation from a melt-rich boundary layer at the LAB. We synthesize the results of previous studies on melt migration in the asthenosphere and buoyant dike propagation in the lithosphere in a coupled model that captures the important physical characteristics of both porous flow and dike propagation. Our new approach allows us to quantify the melt flux and heat flux into the lithosphere, as well as the stability of melt at the LAB.

The melt flux into the lithosphere via diking controls tectonic behavior at mid-ocean ridges (MORs) and active continental rifts where continental margins are created. Dike propagation accommodates extension (Buck et al., 2005; Tucholke et al., 2008) and provides heat that weakens the lithosphere by increasing the fraction of the lithosphere deforming by ductile creep (Royden et al., 1980; Bialas et al., 2010), resulting in differences between regions of magma rich and magma poor extension. Magma rich margins, for example, display narrow zones of localized extension with little detachment faulting (White et al., 1987; Menzies et al., 2002) while amagmatic margins are characterized by deep extensional basins formed by diffuse normal faulting (Whitmarsh et al., 2001; Cochran, 1983). The supply of melt is tied to production of melt in the asthenosphere, evident in the correlation of rifting location with continental flood basalt eruption and plume head arrival (Morgan, 1971, 1983).

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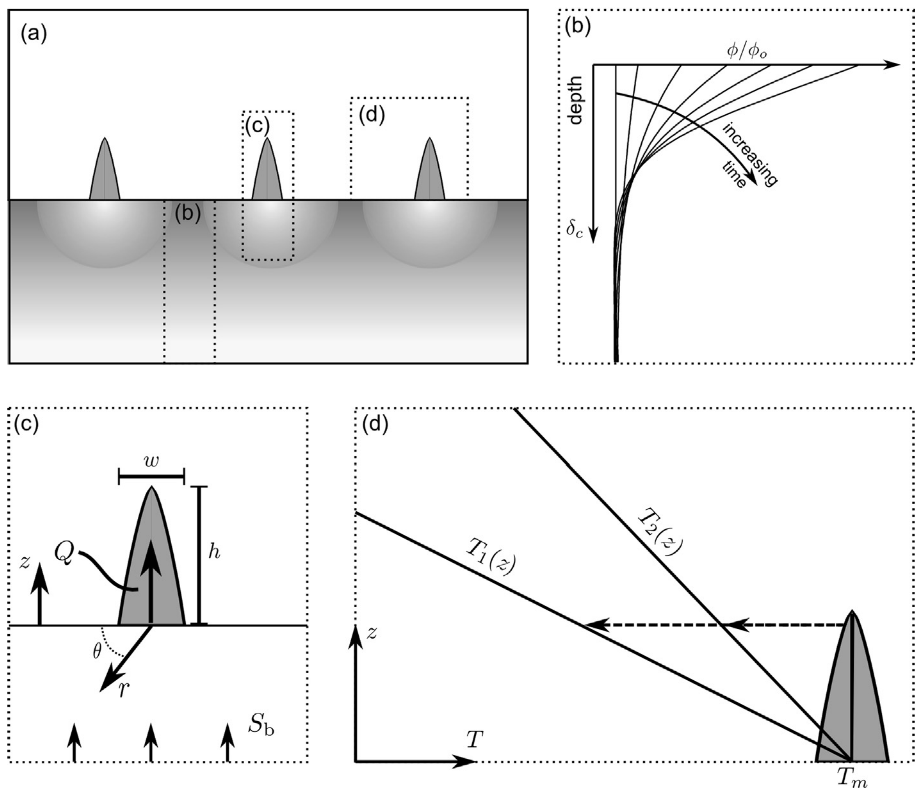

Fig. 1. The accumulation of melt at the lithosphere–asthenosphere boundary and subsequent penetration of melt into the lithosphere (inset a) is controlled by the interaction of the processes illustrated in insets b–d. (b) Melt formed in the asthenosphere rises buoyantly by porous flow and accumulates beneath a freezing boundary in a decompacting boundary layer. ϕ_o is the reference background melt fraction and δ_c is the reference compaction length. For $\phi_o = 0.01$ and typical parameters in Table 1, $\delta_c \approx 20$ km. (c) If the excess pressure is sufficiently high, dikes may propagate into the lithosphere. The growth rate and ultimate volume of a dike is limited by the rate at which fluid can move into the dike as well as the supply rate of new fluid into the decompacting boundary layer, $S_\mathrm{b} = (1-\phi_o)\Delta\rho g k(\phi_o)/\mu$. (d) The extent of dike propagation is limited by the freezing rate, which depends on the horizontal temperature gradient at each height in the dike. A geotherm with a larger gradient, $T_1(z)$, results in a larger freezing rate than a geotherm with a smaller gradient, $T_2(z)$.

The melt distribution below the lithosphere, which determines the source volume for diking, is constrained by seismic studies. Average melt fractions in upwelling mantle beneath a mid-ocean ridge are between 10^{-4} and 0.02 (Forsyth, 1998; Evans et al., 1999), but may increase on length scales shorter than the resolution of seismic tomography and magnetotelluric inversion. The sharp drop in seismic velocity at depths consistent with the base of the oceanic and Phanerozoic continental lithosphere, may in some locations be due to the accumulation of melt at a permeability or freezing boundary (e.g. Rychert et al., 2005; Kawakatsu et al., 2009; Fischer et al., 2010). Beneath Hawaii, where melt is generated by a mantle plume, the similarity in discontinuity depth ($\sim$ 75 km; Rychert and Shearer, 2011; Schmerr, 2012), and the maximum depth of dike-associated seismicity ($\sim$ 60 km; Wright and Klein, 2006), suggests that dikes propagate upwards from these zones of melt accumulation, which may contribute to the local thinning of the lithosphere above the Hawaiian plume as observed by Li et al. (2004).

Despite the relationship between lithosphere dike propagation and asthenosphere porous flow, most studies treat the two regimes independently. Studies on melt migration in the asthenosphere typically assume that melt generated beneath the lithosphere or a ridge axis is extracted from the asthenosphere, forming new crust. Some studies use a partial coupling by assuming a critical melt fraction above which melt is removed (Ghods and Arkani Hamed, 2000), by imposing a narrow window through which melt can escape (Katz, 2008) or by applying a suction due to an existing buoyantly pressurized dike (Katz, 2010). In contrast, fracture mechanics models of dike propagation assume a constant fluid flux (Spence and Turcotte, 1985) or constant fluid excess pressure (Lister and Kerr, 1991) at the base of an upwardly propagating dike. Studies that focus on the effect of melt emplacement on tectonics (Tucholke et al., 2008; Bialas et al., 2010) treat the supply of melt at the base of the lithosphere as an adjustable parameter.

1.2. Conceptual model

The degree to which melt remains stable at the top of the asthenosphere or penetrates the lithosphere, subsequently affecting lithosphere strength and structure, is a function of melt availability, tectonic environment, and the ability for melt to migrate without freezing. To investigate the coupling among these variables, we synthesize results from earlier studies to develop a coupled model of melt migration from source to surface. These processes are illustrated schematically in Fig. 1.

Melt generated by adiabatic decompression in the mantle rises buoyantly. When melt collects at a freezing boundary (Fig. 1b), the solid matrix decompacts to accommodate the additional melt. If the pressure difference between fluid and solid (excess pressure, P) is large enough, dikes will propagate from the decompacting boundary layer (DBL) into the lithosphere. Dike growth (Fig. 1c) is limited by the rate at which fluid flows into the dike from the source region as well as the flux of melt into the DBL from below.

The porosity, ϕ, of a DBL when dike propagation begins controls the kinetics of dike growth. If dike propagation begins after ϕ increases to the rheologically critical melt fraction ($\phi_{RCMF} \approx$ 0.25; Scott and Kohlstedt, 2006), where solid grains lose contact, dike growth is controlled by the combination of magma buoyancy, source excess pressure and viscous pressure gradients due to flow (e.g. Lister, 1990; Roper and Lister, 2005). If dikes