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Effects of axially variable diking rates on faulting at slow spreading mid-ocean ridges

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ABSTRACT

Magma supply for dike injection can be highly variable within a segment of a slow-spreading mid-ocean ridge but the tectonic impact of this variability is not fully understood. Here, we use three-dimensional numerical models to quantify the effects of variable diking rates on the faulting mode at a 20 km-long slow spreading ridge segment. In addition to end-member faulting modes in which long-lived detachment faults or short-lived normal faults form along the whole segment, we newly identify a transitional mode in which a detachment and a short-lived normal fault form simultaneously but in respective domains separated by a transfer fault. Different faulting modes can be better correlated with the average dike intrusion rate, rather than the highest or lowest rate along the segment. Along-axis stress coupling tends to homogenize fault offset along the segment, inhibiting the domination of a particular faulting mode associated with an extreme local diking rate. This homogenizing effect explains why detachment faults can sometimes form even in the regions previously considered as unfavorable. Our results further suggest that a long (>15 km) and continuous detachment, partially overlain by younger faults, can create an oceanic core complex when faults weaken fast and diking rate is low. When faults weaken slow and diking rate is moderate, however, faulting occurs in the transitional mode, producing a detachment over only a part of the segment length.

1. Introduction

Modes of faulting at mid-ocean ridges (MORs) are sensitive to the fraction of plate separation that is accommodated by dike intrusion, which is denoted as M (Buck et al., 2005; Tucholke et al., 2008). Numerical models adopting the M-factor parameterization have been successful in explaining major bathymetric contrasts between fast- and slow-spreading ridges (Buck et al., 2005), and variations in faulting and axial morphology in slow to intermediate spreading ridges (Behn and Ito, 2008; Ito and Behn, 2008). M is also useful for quantifying the favorable conditions for the formation of oceanic core complexes (OCCs), where lower crustal or mantle rocks with a corrugated surface are exhumed by a long-lived (>1 Myr) detachment (Cann et al., 1997; Tucholke et al., 1998). Previous modeling studies suggest that OCCs form when M is between 0.3 to 0.5 (Tucholke et al., 2008). When M > 0.5, frequent diking pushes faults away from the ridge axis and, new axial valley faults replace them. This process occurs alternatingly on both sides of the ridge (Buck et al., 2005; Shaw and Lin, 1993), forming symmetric abyssal hills across the ridge axis.

However, recent observations on slow-spreading ridges pose challenges to these models. M values estimated for OCC-forming segments at 13°18′N and 13°48′30″N Mid-Atlantic Ridge (MAR) (MacLeod et al., 2009) can be locally as high as 0.7 (Fig. 1) while as low as 0 to 0.3 at the Atlantis Massif (Grimes et al., 2008) and the Atlantis Bank (Baines et al., 2008). Previous two-dimensional (2D) models cannot fully explain how OCCs can form at location of such high or low M values. In addition, how OCC-forming detachments transition to the abyssal hills-forming normal faults is not clear (MacLeod et al., 2011; Reston and Ranero, 2011). A long-known example is the dome-to-abyssal hills transition found in the Atlantis Massif (Blackman et al., 2008) and a new one is the series of OCCs separated by symmetric abyssal hills between 13°10′N and 13°55′N MAR (MacLeod et al., 2009) (Fig. 1). Clearly, 2D models implicitly assuming that M is uniform along a ridge segment cannot explain these observations.

Addressing these questions requires considering spatially variable M. Especially at slow spreading ridges like the Mid-Atlantic

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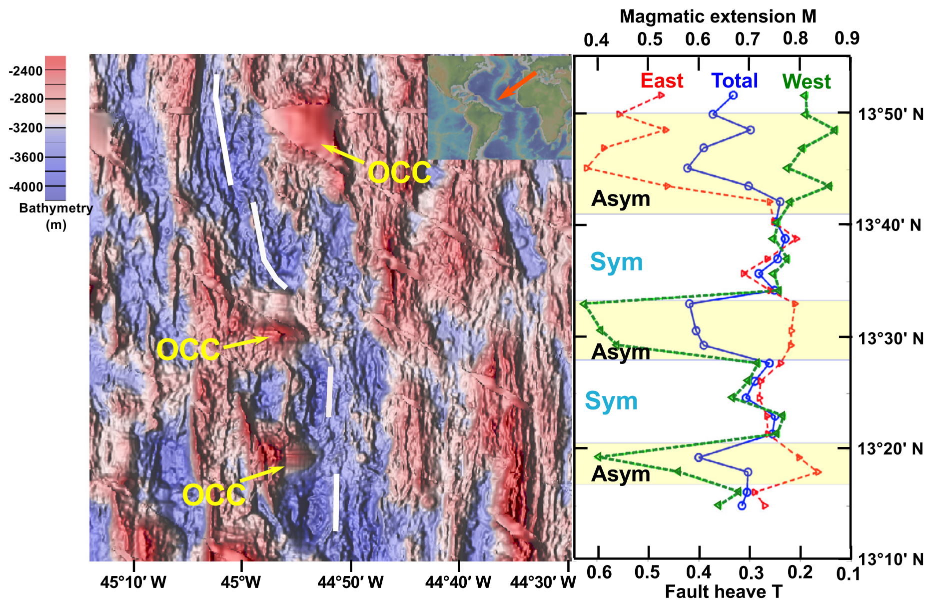

Fig. 1. Bathymetry and the along-ridge variations in M of the Mid-Atlantic Ridge from 13°10′N to 13°55′N. (left) Bathymetry (Ryan et al., 2009) and the inferred axis of the neovolcanic zone (MacLeod et al., 2009) (white lines) with oceanic core complexes (OCCs) annotated. (right) M variations along the ridge adapted from MacLeod et al. (2009). Red and green lines show M variations on the east and the west side of the ridge axis. The mean of the two is the total M (blue line). Fault heave $T = 1 - M$. The yellow shaded areas annotated as "Asym" represent the ridge-parallel extent of asymmetrically spreading OCCs, where the average M is 0.63. The areas annotated as "Sym" show symmetrically spreading abyssal hills and the average M is 0.73.

Ridge (MAR), magma supply is usually higher at the center of the ridge segment and decreases towards the tip of the segment (Carbotte et al., 2015; Lin et al., 1990; Tolstoy et al., 1993). Discrete foci of magmatic accretion along a ridge axis can have wavelengths of 10 to 20 km (Lin et al., 1990). In this study, we investigate how the along-axis variation in magma supply governs the transitions in faulting mode by prescribing variations in M to fully three-dimensional (3D) numerical models.

2. Numerical methods

We use the open source geodynamic modeling code, SNAC (StGermaiN Analysis of Continua) (Choi et al., 2008). SNAC is an explicit Lagrangian finite element code that solves the force and energy balance equations. For each time step, strain and strain rates are updated based on the initial or previous velocity fields under constraints from boundary conditions. A constitutive model returns updated stresses corresponding to these deformation measures. Internal forces are then calculated from the updated stresses and are divided by an artificial mass, which helps damp elastic oscillations, and the resulting accelerations are integrated over time to yield velocities and displacements. A 3D domain is discretized into hexahedral elements, each of which is in turn divided into two sets of tetrahedra. This symmetric discretization prevents faulting from favoring a specific "grain" of a mesh.

Rheology for the oceanic lithosphere is assumed to be elasto-visco-plastic. We use the combination of linear isotropic elasticity, power-law viscosity of dry diabase (Buck et al., 2005; Kirby, 1983) and the Mohr–Coulomb plastic model. We choose rheological structure and parameters that are similar to those used in previous studies (Tucholke et al., 2008) but make further simplifying assumptions such as a steady-state temperature distribution and the use of a single material type for mantle. Weakening of a fault occurs as plasticity goes through strain softening, which is realized by cohesion reduction with an increasing amount of the second invariant of deviatoric plastic strain. We assume this relationship to be linear for simplicity. The initial and final values of cohesion are 44 and 4 MPa. A critical fault offset (Lavier et al., 2000) (ΔX_c), at which cohesion is reduced to a minimum, is used to define the rate of strain weakening. The fault weakening rates based on ΔX_c of 300 m and 1000 m, respectively, are termed "fast" and "slow" in this study. Following previous studies (Buck et al., 2005; Poliakov and Buck, 1998), we also implement a healing process that counteracts the fault weakening process, restoring cohesion in non-deforming regions. In this study, we define the change of plastic strain by healing per time step as $1.0/(1.0 + \Delta t/\tau_h)$, where Δt is the time step and τ_h is the healing time scale, 10^{12} s or ~32 kyrs (Poliakov and Buck, 1998). The complete list of model parameters is given in Table 1.

We follow previous studies to represent the fraction of diking-accommodated plate separation with the 'M' factor (Buck et al., 2005; Poliakov and Buck, 1998). Diking is assumed to occur in the middle of the domain where the lithosphere is the thinnest and dike elements are set to have zero plastic strain. We also calculate pseudo-2D models in order to benchmark against previous studies and to compare with the 3D models. The pseudo-2D model has only one element in the ridge-axis direction to save computational cost and has constant M. All the other settings are the same with 3D models. We verified that pseudo-2D models created with SNAC reproduce the results from selected previous studies (Buck et al., 2005; Tucholke et al., 2008).

The 3D models have a common geometry of 60 × 20 × 20 km (Fig. 2A). The domain is decomposed into 1-km hexahedral elements. Assuming a steady-state temperature field, we reset the temperature field at every time step to a linear profile from 0 °C at a depth of 0 km to 240 °C at 6 km, reflecting enhanced cooling due to hydrothermal circulation, and below 6 km, to the instantaneous cooling of a semi-infinite half-space with plates moving at a constant half-spreading rate of 2.5 cm/yr (Turcotte and Schubert, 2002). Temperature is fixed at 0 °C at the top surface and 1300 °C at the bottom surface. The half spreading rate of 2.5 cm/yr is applied on the model sides that are parallel to the ridge while free-slip boundary conditions are applied to two sides that are perpendicular to the ridge. Hydrostatic seawater pressure is applied on the top surface. Seawater density is assumed to be 1040 kg/m^3. The heights of the water columns are locally determined as $4000 - h(x, z)$ m, where $h(x, z)$ is the topography at horizontal coordinates, x and z. The bottom boundary is free of shear stresses

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