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Control on off-rift magmatism: A case study of the Baikal Rift Zone

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A B S T R A C T

The Baikal Rift Zone (BRZ) is an active intracontinental rift zone with almost no rift-related magmatism along the rift axis. Most magmatism is observed outside the rift center, and e.g. the major Vitim volcanic field is displaced more than 200 km from the rift center. The reasons for this regional distribution of the magmatism are enigmatic, in particular regarding the off-rift magmatism. We present results of numerical modeling of rift structures similar to BRZ that develop in the transition between craton and orogenic belt. Geophysical evidence suggests that pre-existing weak zones control the location of the BRZ. The models therefore include a pre-set fault (weak zone) within the transition zone between the craton and the orogenic belt as an initial condition in the model. If the pre-existing fault is close to the craton, partial melting in the mantle, due to sub-crustal extension, is offset from the surface rift graben by over 200 km. A horizontal shear zone in the lower crust transfers the extension from the shallow crust to the lithospheric mantle far from the rift graben. A pre-set fault close to the orogenic belt zone is required to explain magmatism close to the rift center in the southernmost BRZ. We conclude that the location of the pre-existing fault zone relative to a change in lithosphere thickness may shift the magmatism away from the surface expression of the rift graben and may explain the location of the off-rift magmatism, e.g. the basaltic Vitim plateau. The predicted geometry of the sedimentary graben along the rift strike is consistent with seismic observations in the northern and central BRZ.

1. Introduction

The Baikal Rift Zone (BRZ) is located between the southeastern edge of the Siberian craton and the Sayan–Baikal fold belt (Fig. 1). Rifting at the BRZ began in the Oligocene (30–35 Ma) as 'slow rifting' which lasted ca. 30 Myr (until the Early Pliocene). The second stage of rifting (Late Pliocene to Quaternary) is characterized by accelerating basin subsidence and rift shoulder uplift and is considered as the 'fast rifting' stage (Logatchev and Zorin, 1987).

In general, massive magmatism due to decompression partial melting is expected in rift zones (Reid and Jackson, 1981). Extensional rift centers usually develop in the most extensively stretched part of the lithosphere and therefore magma eruptions are expected in the rift center, but the BRZ is an exception (Kiselev et al., 1987; Maccaferri et al., 2014). Three main volcanic fields have developed at the BRZ since the Miocene: (1) in and around the Tunka basin and the Khamar-Daban range; (2) in the Vitim plateau to the east of BRZ; (3) in the Udokan area in the northeast (Fig. 1). No magma eruption events have ever been recorded in the rift axis, except for small eruptions in the Tunka depression at the SW end of the BRZ (Kiselev et al., 1987; Logatchev and Zorin, 1987). The volcanic fields of the Khamar-Daban, the Udokan and the Vitim plateau are all off-axis where the latter is offset from the axial rift by more than 200 km (Fig. 1).

Previous studies link the off-rift location of the volcanic fields to (1) lithosphere-scale low-angle detachment faults (Bosworth, 1987), (2) flexure-induced extension at the footwall of rift master faults (Ellis and King, 1991), or (3) gravitational unloading in the crust (Maccaferri et al., 2014). However, these models are not satisfactory for the BRZ because (1) no deeply-penetrating faults down to the lithosphere–asthenosphere boundary have been imaged by geophysical surveys (Nielsen and Thybo, 2009; Thybo and Nielsen, 2009); (2) the off-rift volcanoes appear to occur in the hanging wall rather than the footwall (Fig. 1); (3) crustal unloading itself cannot explain the volcano distribution variation along the rift strike (Fig. 1). There is indication that thick continental lithosphere (>100 km) might not allow magmatism in the rift (Bialas et al., 2010; Buck, 2006). However, even if magmatism does not

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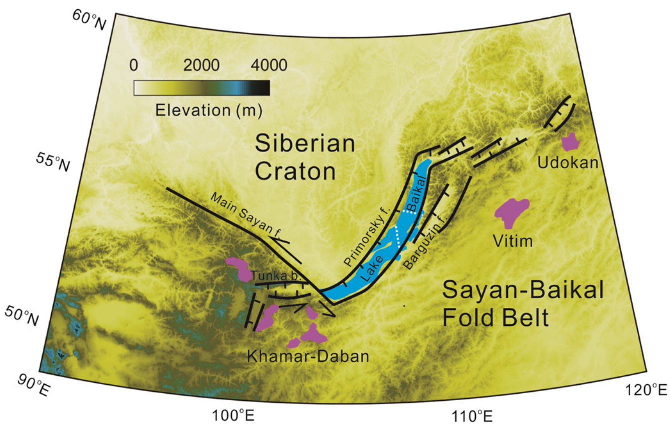

Fig. 1. Map of topography and tectonic setting of the Baikal Rift Zone (BRZ). Faults are named by 'f'. Udokan, Vitim, and Khamar-Daban are three main volcanic fields (red patches, after Kiselev et al., 1987). Black lines roughly mark major tectonic faults (after Sherman, 1978). The white dashed lines across the North and South Baikal lake are two seismic reflection profiles shown in Fig. 5. No magmatism has been recorded along the rift axis since the beginning of stretching at 35–30 Ma except in the Tunka depression. The Vitim volcanic field is located ca. 200 km from the rift axis. (For interpretation of the references to color in this figure, the reader is referred to the web version of this article.)

assist faulting in a rift, synchronous magmatism should not automatically occur far away from the rift center, especially in the presence of a weak zone.

It is commonly recognized that the deformation of the BRZ is guided by inherited weak zones around the suture between craton and orogenic belt (Corti et al., 2011; Petit et al., 1998; Sherman, 1978). The major faults include the Main Sayan fault (>1000 km long) along the southern edge of the Siberian craton, the Primorsky fault bounding the Lake Baikal to the west as a master fault, and the 200 km long Barguzin major fault east of the North Baikal basin (Fig. 1) (Sherman, 1978). Since their initial formation in the early Paleozoic, the faults have been repeatedly reactivated (Konnikov et al., 1993). The role of a pre-existing fault in forming a narrow and deep basin, as is observed in the BRZ, has been tested by analog models (Corti et al., 2011).

In this work, we use numerical geodynamic modeling to investigate the effect of the position of a pre-existing fault on the distribution of volcano eruption centers. Melting processes in the mantle are simulated assuming a batch melting model (Katz et al., 2003), and the modeled melt fractions are compared to local petrological records. In addition, we validate the results of the dynamic processes in the rift zone by comparing predicted rift graben geometry with seismic reflection profiles across the sedimentary basins of the BRZ.

2. Method

2.1. Governing equations

A new 2D finite difference code with a marker-and-cell technique is applied here to simulate the thermomechanical deformation of the crust and upper mantle. We first solve the equations for conservation of mass (1) and momentum (2):

$$\nabla \cdot u = 0 \tag{1}$$

$$\nabla \cdot \sigma' - \nabla P + \rho g = 0 \tag{2}$$

where u is velocity, σ' is deviatoric stress, P is dynamic pressure, ρ is density, and g is gravitational acceleration ($g_x = 0$ and $g_y = 9.81\ \mathrm{m\,s^{-2}}$ for 2D model). For the equation of conservation of mass (1), we assume an incompressibility condition so that changes in density (e.g., due to pressure, temperature, and phase changes) are negligible. To simulate changes in temperature caused by heat transfer, we solve the heat conservation equation

$$\rho c_p \left(\frac{\partial T}{\partial t} + u \cdot \nabla T \right) - \nabla \cdot k \nabla T - H = 0 \tag{3}$$

where c_p is heat capacity, k is thermal conductivity and H is heat production due to radioactive, shear, adiabatic and latent heat sources. The governing conservation equations, of mass, momentum, and energy are solved in the Eulerian frame and material physical property is transported by Lagrangian markers which move in the velocity field interpolated from a fixed Eulerian grid (Gerya, 2009).

2.2. Rheology

The model simulates a visco-elasto-plastic rheology. The elastic material follows Hooke's law for a 2D continuum. For viscous flow, we include the power law dislocation creep

$$\dot{\varepsilon} = A (\sigma')^n \exp\left(-\frac{E + VP}{RT} \right) \tag{4}$$

where $\dot{\varepsilon}$ is strain rate, A is a material constant, σ' is differential stress (the difference between maximal and minimal), n is stress exponent ($n = 1$ for diffusion creep and $n > 1$ for dislocation creep), E is activation energy, V is activation volume, R is the gas constant, and T is temperature. We can reformulate equation (4) to obtain the effective viscosity

$$\eta = \frac{\sigma_{\mathrm{II}}}{2\dot{\varepsilon}_{\mathrm{II}}} \tag{5}$$

where σ_{II} is the second invariant of the deviatoric stress, and $\dot{\varepsilon}_{\mathrm{II}}$ is the second deviatoric strain rate invariant of viscous part of deformation. The effective viscosity is cut at 10^{25} Pa s and 10^{18} Pa s. The Navier–Coulomb yield criterion is used for frictional-plastic deformation when stress reaches a specific limit that marks the transition from viscous to plastic failure (Byerlee, 1978)

$$\sigma_{\mathrm{yield}} = P \sin\phi + C \tag{6}$$

where σ_{yield} is the maximum second deviatoric stress invariant, P is pressure, $\sin\phi$ is friction coefficient and C is cohesion. We follow numerical implementation of eqs. (1)–(6) as described by Gerya (2009).

2.3. Free surface

A free surface condition at the boundary between rock surface and air may have significant effect on lithospheric and mantle dynamics (Kaus and Becker, 2008). It is implemented by introducing a low-density (1000 $\mathrm{kg\,m^{-3}}$), low viscosity (10^{18} Pa s) near-surface layer. Subsequently, the interface between the markers defining the crust and the air behaves similar to a free surface. In order to obtain meaningful results, the following criterion should be satisfied for the top weak layer (Crameri et al., 2012):

$$\frac{\eta_{st}/\eta_{ch}}{(h_{st}/L)^3} \ll 1 \tag{7}$$

where η_{st} and h_{st} are viscosity and thickness of the weak layer, and η_{ch} and L are the characteristic mantle viscosity (which controls topography relaxation) and the length scale of the model, respectively.