Linking rift propagation barriers to excess magmatism at volcanic rifted margins

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ITEM DR1. DISTRIBUTION OF SEAWARD-DIPPING REFLECTORS

Offshore volcanics, known as seaward dipping reflectors (SDRs; e.g. Hinz 1981; Mutter et al., 1982; Planke and Eldholm, 1994) are offset or dissected along transfer zones (Figure DR1). In the South Atlantic an intrinsic relationship between the volumes of SDRs and the presence of segment boundaries has been proposed (Franke et al., 2007). It might be speculated that variations in the volumes of the extrusive magmas are the result of varying accommodation space within the rift, potentially resulting from the presence of segment boundaries but also that rift propagation barriers by themselves cause variations in the volumes of melts. Segmentation has also been described for the Uruguayan margin by Soto et al. (2011).
Figure DR1: Simplified plate reconstruction of the South Atlantic margins to Magnetic Anomaly M4: The term margin segmentation as used in this research is derived from detailed mapping of South American (e.g. Franke et al., 2007) and South African (e.g. Koopmann et al., 2014) margin structures. Most importantly, the onset of and offsets within offshore volcanic units, reflected in seismic data as seaward dipping reflectors (SDRs) allow the definition of segment boundaries or transfer zones along the margin. Further, magnetic anomalies merge with the mapped area of SDRs occurrence, underlining the idea of a segmented continental break-up from south to north.
ITEM DR2. NUMERICAL MODEL DESCRIPTION

DR2.1 Basic setup
The rift evolution has been modelled using the thermo-mechanical forward modelling software SLIM3D (Semi-Lagrangian Implicit Model for 3 Dimensions) described in Popov and Sobolev (2008). This finite-element code has been successfully applied in a number of geodynamic modelling studies at convergent (Quinteros and Sobolev, 2012; Quinteros et al., 2010), divergent (Brune et al., 2014, 2012) and transform (Popov et al., 2012) plate boundaries.

SLIM3D solves the coupled conservation equations of momentum

\[- \frac{\partial p}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} + \rho g_i = 0\]

energy

\[\rho C_p \frac{DT}{Dt} = \frac{\partial}{\partial x_i} \left( \frac{\partial T}{\partial x_i} + \tau_{ij} \dot{\varepsilon}_{ij} + \rho A \right)\]

and mass

\[\frac{1}{K} \frac{Dp}{Dt} + \alpha_T \frac{DT}{Dt} + \frac{\partial v_i}{\partial x_i} = 0\]

with coordinates \(x_i\), time \(t\), material time derivative \(D/Dt\), velocities \(v_i\), temperature \(T\), pressure \(p\), stress deviator \(\tau_{ij}\), strain rate deviator \(\dot{\varepsilon}_{ij}\), density \(\rho\), gravity vector \(g_i\), heat capacity \(C_p\), heat conductivity \(\lambda\), thermal expansivity \(\alpha_T\), radioactive heat production \(A\), and bulk modulus \(K\). The Einstein summation rule applies for repeated indices.

Lithospheric deformation is implemented in terms of an elasto-visco-plastic rheology that is incorporated by decomposing the deviatoric strain rate into elastic, viscous, and plastic components. This method self-consistently reproduces diverse lithospheric-scale deformation processes like faulting, flexure and lower crustal flow.

The Mohr-Coulomb model is used for implementation of plastic failure:

\[F = \frac{1}{2} (\sigma_{\text{max}} - \sigma_{\text{min}}) + \frac{1}{2} (\sigma_{\text{max}} + \sigma_{\text{min}}) \sin \varphi - c \cos \varphi = 0\]

with the yield surface \(F\), maximum and minimum principal stresses \(\sigma_{\text{max}}\) and \(\sigma_{\text{min}}\), effective friction angle \(\varphi\), and cohesion \(c\).

Viscous flow occurs via two creep mechanisms: diffusion and dislocation creep:

\[\dot{\varepsilon}_{\text{Diff}} = B_{\text{Diff}} \tau_{\text{II}} \exp \left(\frac{E_{\text{Diff}} + p V_{\text{Diff}}}{RT}\right)\]

\[\dot{\varepsilon}_{\text{Disloc}} = B_{\text{Disloc}} (\tau_{\text{II}})^n \exp \left(\frac{E_{\text{Disloc}} + p V_{\text{Disloc}}}{RT}\right)\]

where \(\dot{\varepsilon}\) denotes the effective strain rate, \(\tau_{\text{II}}\) the second invariant of deviatoric stress, \(n\) the stress exponent, \(B\) the pre-exponential factor, \(E\) the activation energy, \(V\) the activation volume and \(R\)
the gas constant. The model involves three weakening mechanisms: shear heating, dislocation creep-related strain rate softening, and friction softening. All model parameters are listed in Supplementary Table S1.

Table DR1: Thermo-mechanical parameters. *During frictional strain softening, the friction coefficient $\mu$ reduces linearly from 0.5 to 0.05 (corresponding to effective friction angles $\phi$ 27° and 3°, respectively) for brittle strain between 0 and 1. For strains larger than 1, it remains constant at 0.05.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Upper Crust</th>
<th>Lower Crust</th>
<th>Strong Mantle</th>
<th>Weak Mantle</th>
</tr>
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<tbody>
<tr>
<td>Density, $\rho$ (kg m$^{-3}$)</td>
<td>2700</td>
<td>2850</td>
<td>3280</td>
<td>3300</td>
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<td>Thermal expansivity, $\alpha$ (10$^{-5}$ K$^{-1}$)</td>
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<td>2.7</td>
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<td>Bulk modulus, $K$ (GPa)</td>
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<td>122</td>
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<tr>
<td>Shear modulus, $G$ (GPa)</td>
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<td>40</td>
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<td>74</td>
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<tr>
<td>Heat capacity, $C_T$ (J kg$^{-1}$ K$^{-1}$)</td>
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<td>1200</td>
<td>1200</td>
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<tr>
<td>Heat conductivity, $\lambda$ (W K$^{-1}$ m$^{-1}$)</td>
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<td>2.5</td>
<td>3.3</td>
<td>3.3</td>
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<tr>
<td>Radiogenic heat production, $A$ ($\mu$W m$^{-3}$)</td>
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<td>0.0</td>
<td>0.0</td>
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<tr>
<td>Initial friction coefficient *, $\mu$ (-)</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>Initial friction angle *, $\phi$ (°)</td>
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<td>27</td>
<td>27</td>
<td>27</td>
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<tr>
<td>Cohesion, $c$ (MPa)</td>
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<td>5.0</td>
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<td>Pre-exponential constant for diffusion creep, $\log(B_{\text{Diff}})$ (Pa$^{-1}$ s$^{-1}$)</td>
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<td>-</td>
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<td>-8.65</td>
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<tr>
<td>Activation energy for diffusion creep, $E_{\text{Diff}}$ (kJ / mol)</td>
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<td>-</td>
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<td>335</td>
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<td>Activation volume for diffusion creep, $V_{\text{Diff}}$ (cm$^3$ / mol)</td>
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<td>-</td>
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<td>4</td>
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<tr>
<td>Pre-exponential constant for dislocation creep, $\log(B_{\text{Disloc}})$ (Pa$^{-n}$s$^{-1}$)</td>
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<td>-21.05</td>
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<td>0</td>
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<td>10</td>
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</table>

The model consists of a 1500 km long domain that is 250 km wide and 150 km deep. With a resolution of 5 km in x- and z-direction and 10 km in y-direction, it involves 225,000 elements and 238,731 nodes. It accounts for four distinct petrological layers: The upper and lower crustal layers are each 20 km thick and deform in accordance with a wet quartzite flow law for dislocation creep (Gleason and Tullis, 1995), and a mafic granulite flow law for dislocation creep (Wilks and Carter, 1990), respectively. We use dry olivine rheology (Hirth and Kohlstedt, 2003) to model deformation of the strong, depleted, subcontinental mantle extending to 90 km depth, while we use a wet olivine (i.e., 500 ppm H/Si) flow law (Hirth and Kohlstedt, 2003) for the weak, asthenospheric mantle covering the lowermost 60 km of the model domain. Throughout the mantle, both diffusion and dislocation creep are incorporated.
DR2.2 Boundary Conditions
We impose sequential extension of three rift segments by applying a velocity boundary condition at the model sides facing in the x-direction. Note that these boundaries do not move, but that material flows out of the model domain with a prescribed velocity. This velocity is 10 mm/yr at each model side resulting in 20 mm/yr full extension rate. Rift segments are activated sequentially. At model start, only the first rift segment experiences extension while the other two segments are kept at constantly 0 mm/yr boundary velocity. In the next phase, Segment 1 and 2 extend at the same rate while Segment 3 is still immobile. Finally, all segments experience extension. Boundary conditions in terms of stress are as follows: Zero shear stress (free slip) at the x-facing model sides, continuous stress condition (Brune et al. 2012) at the y-facing sides.

Lateral outflow of material is compensated by dynamic inflow through the bottom boundary. This is accomplished by means of the isostasy-based Winkler foundation, where in- and outflow of material is accounted for during re-meshing (Popov & Sobolev 2008). The top boundary is governed by a free surface allowing self-consistent evolution of topography that is tracked by the finite element mesh.

DR2.3 Thermal model setup
Solving the energy equation, the model’s temperature field results from the material-specific heat conductivity, radiogenic heat production, dissipation of mechanical energy and the following boundary conditions: The surface temperature is held constant at 0°C while the bottom temperature is set to 1350°C and the lateral boundaries are thermally isolated.

As an initial condition, a small thermal heterogeneity is introduced in the model centre in order to avoid rift localisation at the model boundaries. This heterogeneity is generated by elevating the 1350°C isotherm in form of a triangular shape with 20 km height and 20 km width.

DR2.4 Melt generation
We estimate melt generation in a post-processing step by implementing a batch melting model for peridotite (Katz et al., 2003). The solidus temperature for a given pressure amounts to $T_{\text{sol}}=-5.1p^2+132.9p+1085.7$ while the liquidus temperature $T_{\text{liq}}=-3.2p^2+80.0p+1475$, where temperatures are measured in °C and pressure $p$ in GPa. Following the computation of $T_{\text{sol}}$ and $T_{\text{liq}}$ at each element, we quantify the melt fraction of peridotite $X$ at the temperature $T$ as $X = \frac{(T-T_{\text{sol}})}{(T_{\text{liq}}-T_{\text{sol}})}^{3/2}$.

Moreover, we account for endothermic consumption of lattice energy by computing the associated temperature change $\Delta T_{\text{lattice}} = X \Delta S / C_p (T+273)$, with the entropy change during melting (Katz et al., 2003) $\Delta S =300$ J kg$^{-1}$ K$^{-1}$ and the heat capacity $C_p=1200$ J kg$^{-1}$ K$^{-1}$. We evaluate the melt fraction that is self-consistent with lattice energy consumption by iterative computation of $X$ and $T=T_0 - \Delta T_{\text{lattice}}$, where $T_0$ is the initial temperature that does not account for energy consumption during melting.

Vertical integration over the melt-generating area allows us to approximate how much magma is produced by decompression melting. In order to compare our results to SDR distribution, we restrict our analysis to pre-break-up times. We assume vertical and total melt extraction. Since we are only interested in pre-break-up melts, the melt production after crustal separation (i.e. arrival of the wet olivine phase at the surface) is blanked in our Figures.
ITEM DR3. ALTERNATIVE SETUPS AND MODEL ROBUSTNESS

Despite recent advances, current 3D models often require a relatively coarse resolution and a limited computational domain. In order to evaluate the robustness of our results with respect to model resolution and model size, a series of alternative models have been computed that are shown in Supplementary Figure DR2:

- M1 (black) is identical to the model of the main manuscript with a rift delay time of 5 My, except that the model comprises only 2 segments of each 400 km length.
- M2 (red) is identical to M1 but involves two times higher vertical resolution (2.5 km)
- M3 (orange) is identical to M1 but involves two times higher along-strike resolution (5 km)
- M4 (green) is identical to M1 but its model domain extends to 300 km depth.
- M5 (blue) is identical to M1 but its model domain extends to 410 km depth, covering the entire upper mantle above the transition zone. In order to run this model within a reasonable time, we used somewhat coarser elements (6×6×12 km instead of 5×5×10 km).

The overall distribution of magmatic material with respect to the segment boundary is robustly reproduced by all alternative models. Finer resolution even increases the peak in excess melt if compared to the reference model, indicating that the models shown in the main paper provide a lower limit for the excess melt generation due to rift segmentation.

![Final distribution of magmatic material](image)

**Figure DR2:** Results of alternative models. Model extent and resolution have a quantitative influence on the modelled distribution of magmatic material (compare Fig. 4), but they do not affect the existence and the location of the melt generation peak. In fact, the peak becomes more prominent than in the reference model described in the main text for 3 out of 4 alternative models.
ITEM DR4. ANIMATIONS

To improve visualization, we included 5 animations of the model (Animations A1-A5):

**Animation DR1:** Animated model run from 1 My to 25 My, with a rift delay of 5 My. Included are material phases, strain rate and 1300°C isotherm plots.

**Animation DR2:** Animated model run from 1 My to 25 My, with no rift delay. Included are temperature distribution, along-strike velocity, melt thickness production and final distribution plots.

**Animation DR3:** Animated model run from 1 My to 25 My, with a rift delay of 2.5 My. Included are temperature distribution, along-strike velocity, melt thickness production and final distribution plots.

**Animation DR4:** Animated model run from 1 My to 25 My, with a rift delay of 5 My. Included are temperature distribution, along-strike velocity, melt thickness production and final distribution plots.

**Animation DR5:** Animated model run from 1 My to 25 My, with a rift delay of 7.5 My. Included are temperature distribution, along-strike velocity, melt thickness production and final distribution plots.

References


