

Linking rift propagation barriers to excess magmatism at volcanic rifted margins

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ABSTRACT

Break-up-related extrusive magmatism, imaged in reflection seismic data as seaward-dipping reflectors (SDRs), extends symmetrically along the volcanic margins of the Atlantic Ocean. Recent research found distinct along-margin variations in the distribution of SDRs, and abundance of volcanic material was found to be spatially linked to transfer fault systems. These segmented the propagating rift that later developed into the ocean, and are interpreted as rift propagation barriers. Based on these observations, we develop a numerical model, which shows that rift-parallel mantle flow and locally enhanced rates of volcanism are the result of delays in rift propagation and segmented opening. Our model suggests that segmentation is one of the major factors in the distribution and localization of rift-related extrusive magmatism. We conclude that in addition to mantle temperature and inherited crustal structures (e.g., weaknesses from previous rift episodes), rift propagation delay plays an important role in the distribution of extrusive volcanism at volcanic passive margins by controlling the mantle flow beneath the rift axis.

INTRODUCTION

Extensive magmatism (Fig. 1) accompanied the continental break-up of large portions of the Atlantic margins (e.g., Mutter et al., 1982; Eldholm et al., 1989). Propagating rift systems formed the segmented margins (e.g., Planke et al., 1991; Clemson et al., 1997; Franke et al., 2007; Tsikalas et al., 2005) of the southern South Atlantic and the northern North Atlantic (Fig. 1). Segmentation is recorded in the southern South Atlantic in seaward-dipping reflector (SDR) distribution and magnetic anomalies (e.g., Clemson et al., 1997; Franke et al., 2007; Koopmann et al., 2014a). For the North Atlantic between Norway and Greenland, segmentation has also been observed (e.g., Planke et al., 1991; Tsikalas et al., 2005; Elliott and Parson, 2008), although some segment boundaries (or “transfer zones”) are still under discussion (Olesen et al., 2007). Segmentation resulted from inherited crustal zones of stiffness or weakness that guided the rift at these boundaries (e.g., Rosendahl, 1987).

Offshore volcanics are a defining feature of volcanic margins (e.g., Mutter et al., 1982), and their distribution is commonly linked to onshore-emplaced large igneous provinces (LIPs) and the arrival of a mantle plume (White and McKenzie, 1989). A strong correlation between SDR volumes and rift segmentation, together with three-dimensional (3-D) numerical modeling results, lead us to challenge this view. Here, we demonstrate that sequentially active rift segments may create shallow, pressure-driven, rift-parallel flow across segment boundaries, thereby guiding along-strike magmatism.

VOLCANISM AND BREAK-UP IN THE SOUTH AND NORTH ATLANTIC

Continental extension lasted from the Late Triassic (ca. 210 Ma; e.g., Macdonald et al., 2003) until the South Atlantic opened in the Early Cretaceous at ca. 134 Ma. Volcanism related to break-up of the South Atlan-

tic and emplacement of the Paraná-Etendeka LIP in Brazil and Namibia (Fig. 1) peaked between 134 Ma and 129 Ma (dPeate, 1997). Opening of the southern segment of the South Atlantic occurred from south to north (e.g., Austin and Uchupi, 1982) and is interpreted as a successive unzipping of rift zones (e.g., Jackson et al., 2000; Franke et al., 2007). Accordingly, SDRs were emplaced from south to north: magnetic anomaly M9 is the oldest anomaly seaward of the SDRs off Cape Town, South Africa, while 2000 km northward, the oldest magnetic anomaly is M0 (Koopmann et al., 2014b), a difference of ~10 m.y. Seafloor spreading north of the Florianópolis fracture zone (FZ), Brazil (Fig. 1), initiated at ca. 112 Ma (e.g., Torsvik et al., 2009; Moulin et al., 2010; Heine et al., 2013), resulting in an interval of ~20 m.y. between break-up ages of southern (134 Ma) and northern (112 Ma) segments (Franke, 2013). The Paraná-Etendeka LIP is spatially linked to this FZ (Fig. 1).

Continental break-up in the North Atlantic ended an ~350-m.y.-long predominately extensional period (e.g., Ziegler, 1988). During break-up, two sublinear rift segments in the south and north of the opening oceanic basin were connected by a sinuous segment in the middle (Larsen et al., 1994). The North Atlantic LIP in East Greenland (Fig. 1) erupted at 56–55 Ma (e.g., Eldholm et al., 1989; Storey et al., 2007). The third of three episodes emplacing plateau basalts in the center of the North Atlantic LIP in East Greenland (Larsen and Watt, 1985) at ca. 55 Ma correlates with oceanic crust formation. This indicates rift delay during the two previous episodes. Geometrical difficulties in reconstruction of the southwest Vøring (offshore Norway) and northeast Greenland margins indicate discontinuous, complex rifting (Olesen et al., 2007). Oceanic spreading

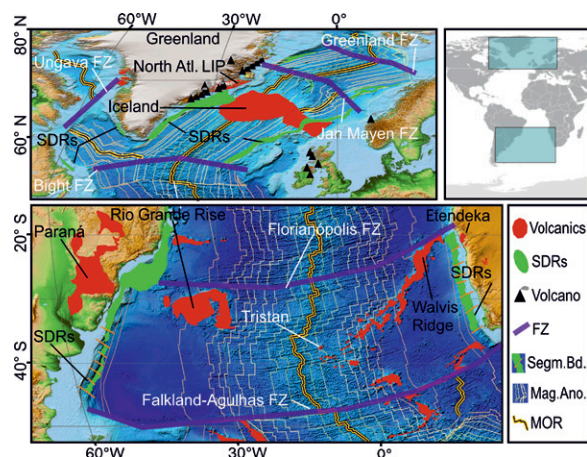


Figure 1. ETOPO1 (www.ngdc.noaa.gov/mgg/global/global.html) map of the South and North Atlantic (Atl.), with offshore and onshore large igneous provinces (LIPs), magnetic anomalies, and other structural elements. Margin segments are defined by structural variations across segment boundaries. Structures are from Bryan et al. (2010), Courtillot et al. (2003), Franke et al. (2007), and Koopmann et al. (2014a). SDR—seaward dipping reflector; FZ—fracture zone; MOR—mid-oceanic ridge; Mag. Ano.—magnetic anomaly; Segm. Bd.—segment boundary.

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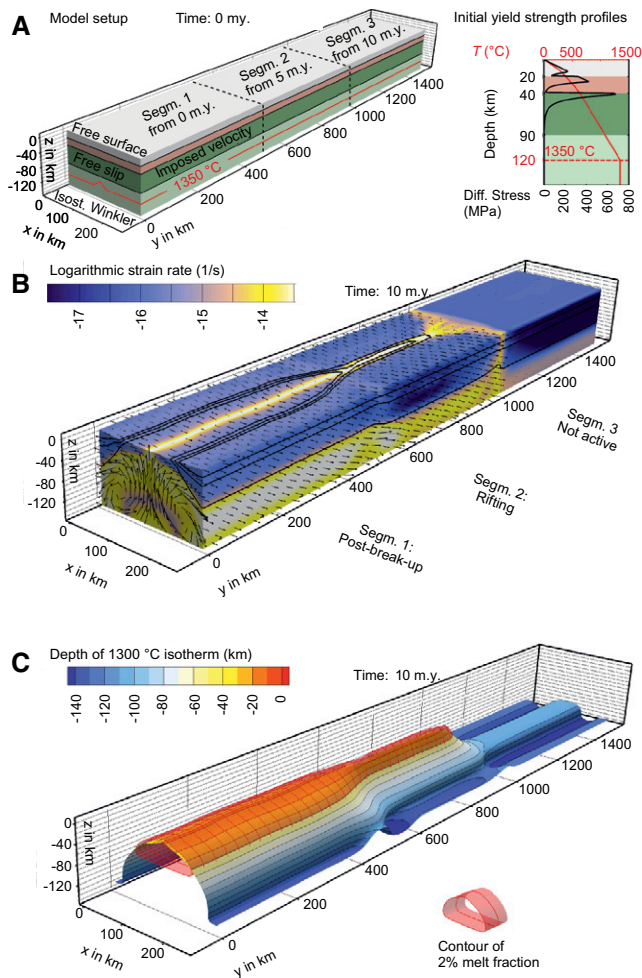


Figure 2. A: Model with sequentially activated segments (left), and initial yield strengths for numerical model (right); weak zone in model center (see 1350 °C isotherm at left). Segm.—segment; Isost.—isostatic. B: Logarithmic strain rate plot shows strain rate and velocities at 10 m.y. Black lines are material boundaries. Surfacing of lowermost layer indicates break-up. Diff.—differential. C: Lithosphere-asthenosphere boundary (1300 °C isotherm) and generation of melt at 10 m.y. See the Data Repository (see footnote 1) for entire model animation (Item DR1).

anomalies C24A, C23, and C22 terminate against the northeast Greenland continental slope between the Greenland FZ and Jan Mayen FZ (Hinz et al., 1991). Spreading propagated southward toward the proposed Iceland hot-spot location, from the Greenland FZ at 54.2 Ma to the proto-Jan Mayen FZ at 50 Ma (Lundin, 2002). Overlapping spreading centers on the Greenland-Iceland Rise persisted to 20 Ma, with northward propagation along proto-Reykjanes and proto-Kolbeinsey ridges (Lundin, 2002). The Jan Mayen FZ was an oceanic transform connecting the two ridges until abortion of the Aegir Ridge between ca. 33 Ma and ca. 22 Ma (C13n–C6B; Gaina et al., 2007; Gernigon et al., 2012).

In conclusion, initiation of rifting occurred in the South and North Atlantic distal from the proposed hot-spot locations (Tristan and Iceland, respectively) with a rift migration toward the hot-spot position, and there is consistent delay in break-up at the proposed location of hot-spots and the location of abundant volcanism.

NUMERICAL MODEL SETUP

We use the numerical code SLIM3D (Popov and Sobolev, 2008) to explore the consequences of imposed rift delay at transfer zones on

melt generation. This elasto-visco-plastic, finite-element code has been successfully applied in a number of 3-D rift experiments (e.g., Brune et al., 2013; Brune, 2014). Our model is 1500 km long, 250 km wide, and 150 km deep, comprising laterally homogeneous crust and mantle layers (Fig. 2A). Full extension velocity is 20 mm/yr in the *x* direction, comparable to the extension rate of the southern South Atlantic rift system (Heine et al., 2013). To test the effect of rift segmentation and delay, we impose sequential extension of three rift segments: rifting starts in the first 500-km-long segment by imposing half of the extension velocity at each model side while the remaining 1000 km of the model is not activated. After 5 m.y., the second segment is activated with the same extension velocity, and only the third segment remains inactive. From 10 m.y. on, all segments extend with a velocity of 20 mm/yr. Transfer faults between the segments are prescribed at model start by reducing the friction coefficient to 0.05. The top boundary has a free surface, the boundaries facing the *y* direction feature a free-slip condition, while the bottom boundary allows inflow and outflow of material through a dynamically tracked Winkler boundary condition (Popov and Sobolev, 2008). A more detailed model description is given in the GSA Data Repository¹. Decompression melting of mantle material is incorporated in a post-processing step. A peridotite batch melting model using the formulation of Katz et al. (2003) and the consumption of lattice energy (Brune et al., 2014) are considered. For simplicity, we neglect the effects of melt-induced viscosity reduction (Bürgmann and Dresen, 2008) and strengthening of residual mantle material (Hirth and Kohlstedt, 1996). We assume that melt is extracted vertically and all magma is emplaced at the surface. Previous numerical models provided insight on melt dynamics in 2-D rift settings (e.g., Armitage et al., 2010; Schmeling, 2010), whereas the presented model is among the first to describe 3-D rift dynamics with melt generation.

NUMERICAL MODEL RESULTS

In agreement with observations from the North and South Atlantic, and as imposed by the model's boundary conditions, break-up in the model occurs via rift propagation (Figs. 2B and 2C). In the same manner, the area affected by decompression melting (between 65 km and 15 km depth beneath the rift center; Fig. 3A) propagates along strike.

Figure 3 depicts a key aspect of the model by illustrating how rift delay controls the flow beneath the rift axis. During the first 5 m.y., extension of segment 1 is accompanied by dominantly vertical flow, until it comprises a reservoir of hot, low-viscosity mantle material. Following activation of the next segment at 5 m.y., parts of this low-viscosity mantle are pulled across the first segment boundary into segment 2. This repeats at 10 m.y., when opening of segment 3 increases rift-parallel flow across the second segment boundary.

This rift-parallel flow is caused by a lateral pressure gradient between sequentially opening segments. The along-strike flow of hot material near the segment boundary leads to elevated temperature and thus decompression melting if compared to the segment interior. This takes place without enhancing crustal thinning near the transfer zones and generates peaks in overall pre-break-up melt volumes (Fig. 4).

Rift-parallel flow and concomitant magmatic peaks near rift propagation barriers are robust features taking place in alternative model scenarios with different delay times (Fig. 4C) and also with different resolutions and model depths (see the Data Repository). If no delay is imposed in our model, the rift system behaves in a 2-D manner, and localized melt generation does not occur. An increase of the rift delay time results in a larger, hotter, and less viscous reservoir of mantle material within the opening segment upon activation of the next segment. This generates faster shall-

¹GSA Data Repository item 2014368, distribution of seaward-dipping reflectors, numerical model description, alternative setups and model robustness, and five animations of the model run, is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

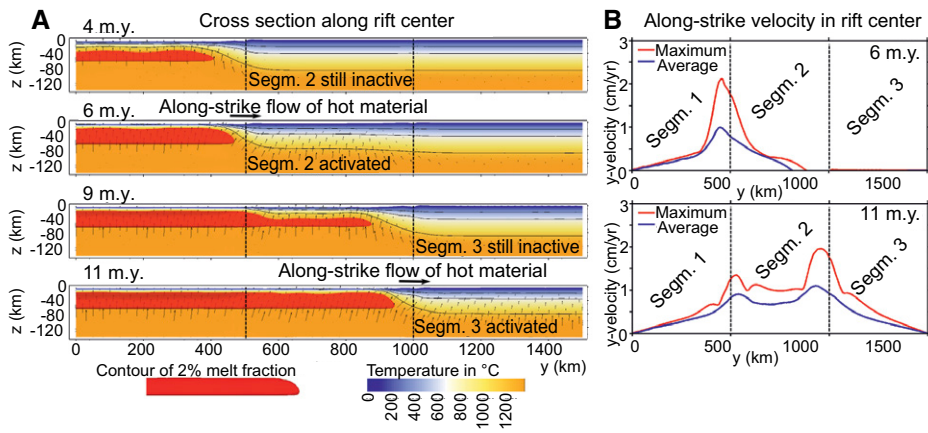


Figure 3. A: Temperature, melt fraction, and velocity in cross section along rift center. At 4 m.y., dominantly upward flow in segment (Segm.) 1; at 6 m.y., along-strike flow at activated segment boundary; at 9 m.y., dominantly upwards flow in segment 2; at 11 m.y., lateral flow toward segment 3. Melt contour arrival accompanies along-strike flow at 6 and 11 m.y. **B:** Velocity field beneath rift center. While the boundary condition at y-facing model sides allows only vertical flow, transfer faults in model interior show a distinct rift-parallel velocity component at 6 m.y. (between segments 1 and 2) and at 11 m.y. (between segments 2 and 3). See the Data Repository (see footnote 1) for animations (AnimationDR2–AnimationDR5) underlining the influence of rift-delay duration.

low mantle flow across the segment boundary, and enhanced melt generation near the transfer zone (Fig. 4C).

DISCUSSION

Assembled observations and modeling results indicate a close relationship between transfer zones and excess volcanism during break-up. This relates to previous analog modeling by Corti et al. (2002) who suggested that magma distribution at depth controls strain localization during transfer zone formation and that, vice versa, magma migrates toward imposed transfer zones (Corti et al., 2004). While these studies focused on magma migration, our study, in contrast, provides an explanation for enhanced magma generation near transfer zones. Note that both the previous analog experiments and our numerical models consistently feature rift-parallel flow of either melt or mantle, respectively.

The exclusive role of elevated temperatures has been questioned previously by Armitage et al. (2009) who found strong evidence that inherited structures from previous extensional periods were needed to explain volcanic rifting in the North Atlantic and that the mantle thermal structure alone was unable to do so. Here we provide additional evidence that geological structures and dynamic rift evolution can control locally enhanced magmatism.

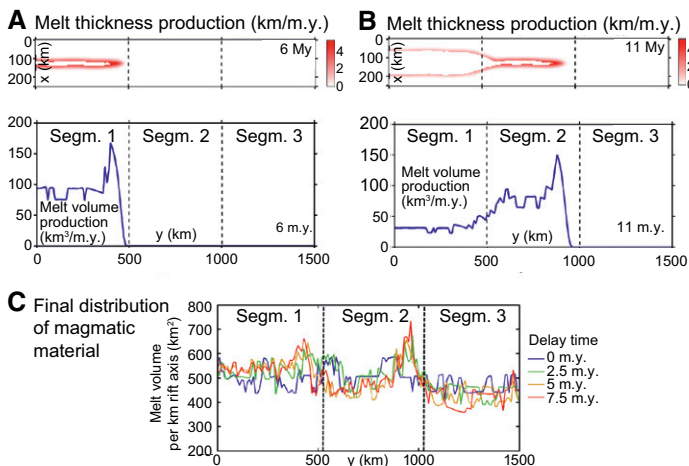


Figure 4. A, B: Upper panels: Production of igneous crustal thickness over time, at 6 m.y. and 11 m.y. Lower panels: Igneous crust production integrated along the x-axis comparable to total melt emplacement at continental margins, at 6 m.y. and 11 m.y. Melt production peaks after activation of segment (Segm.) 2 at 6 m.y. Viscous coupling forces peak production off imposed segment boundary at 500 km. Melt production peaks a second time at newly formed segment boundary at 11 m.y. **C:** Melt production indicates the importance of rift-delay duration. See the Data Repository (see footnote 1) for model animations (AnimationDR2–AnimationDR5) of different rift-delay durations.

Moreover, rift-parallel flow of mantle material has the potential to elucidate the increase in along-strike magmatic volumes and the localization of melt at segment boundaries (Franke, 2013), creating a distinctive triangle-shape for mapped SDRs (see Fig. DR1 in the Data Repository). Most material migration occurs during or right after incipient break-up at the proto–transform fault (the larger part of the triangle), whereas rift-parallel material flow may be responsible for the thinning of the triangle.

If rift segmentation affects offshore volcanism, to what extent could this process have contributed to the emplacement of huge onshore volcanic provinces? Interestingly, the Atlantic Ocean LIPs are spatially linked to the transform fault systems with the largest delay in rift propagation. During unidirectional rift propagation in the South Atlantic (Fig. 1), the most prominent propagation barrier was the proto–Florianópolis FZ, with remarkable offset (150 km; Elliott et al., 2009) and delay in break-up of 5–20 m.y. with respect to the oceanic crust further south. At the complex Jan Mayen FZ system (Fig. 1), an even longer period of rift delay between C25 and C5 can be derived from magnetic anomalies (e.g., Gernigon et al., 2012) and an offset of up to 320 km. This allows the speculation that rift delay and the along-strike flow at segment boundaries drive portions of offshore and onshore volcanic provinces to their location.

CONCLUSIONS

Volumes of offshore magmatic material vary along the volcanic margins of the North and South Atlantic. These margins are segmented, with transfer zones between rift segments indicating rift propagation delay. Numerical modeling suggests a causal link between segment boundaries and magmatic volume variations. Our results explain observations made in natural systems by significant rift-parallel mantle flow across segment boundaries and subsequent melt generation at the tip of each opening rift segment. For specific rift configurations, delay at inherited structures can play an important part in enhancing and localizing volcanic activity.

ACKNOWLEDGMENTS

We acknowledge funding by the German Research Foundation within priority program SPPI375 for projects NE 1193-I-1 (Koopmann) and FR 2119/2-2, and by ERC for Marie-Curie project 326115 (Brune). We thank Editor J. Spotila and reviewers J.J. Armitage, G. Corti, and M. Bonini for their helpful comments and thorough reviews, which greatly benefited this manuscript.

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Manuscript received 16 July 2014

Revised manuscript received 11 September 2014

Manuscript accepted 24 September 2014

Printed in USA