Roadmap to continental rupture: Is obliquity the route to success?

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INTRODUCTION

Continental rift systems progress to plate rupture and seafloor spreading, or fail. The separation of Africa and South America is no exception: ruptured Mesozoic rift systems define their irregular coastlines, with basins overlying thinned crust of failed rift sectors. Heine and Brune (2014, p. 211 in this issue of *Geology*) simulate the opening rates and spatial distribution of Mesozoic extensional basins using numerical models of rheologically layered continental lithosphere. The match between time-averaged deformation patterns and continent-scale predictive models leads them to propose new answers to old questions: Why do some rift sectors rupture, and others fail? What enables strain localization, and thus increased extensional velocities, in some sectors, whereas others stall or cease deformation? Heine and Brune propose that rift sectors with extensional fault systems oblique to opening directions preferentially rupture. The explanation is counter-intuitive: elastic deformation transitions to plastic yielding at a lower plate boundary force when extension is oblique than when fault systems are orthogonal (Brune et al., 2012). A fundamental implication is that obliquely opening sectors do so more quickly.

This study, and that by Heine et al. (2013) are the first to synthesize a wealth of data to simulate continental rifting over spatial scales large enough to include multiple rift sectors, and time scales spanning the ~40 m.y. period of stretching and thinning leading to seafloor spreading. The kinematic models include significant internal plate deformation prior to rupture, the onset of seafloor spreading, and acceleration of the South America plate at the initiation of Andean subduction. The plate-scale comparison adds a new pathway, rift obliquity, to successful plate rupture.

RIFTING TO RUPTURE PROCESSES

As rifting progresses, the thick (125–250 km) continental lithosphere heats and thins, and the zone of deformation broadens (Fig. 1). Initially, strain in the elastic upper crust occurs across <100-km-wide basins. The length scales and along-axis arrangement of fault and magmatic systems impart a third dimension: along-axis

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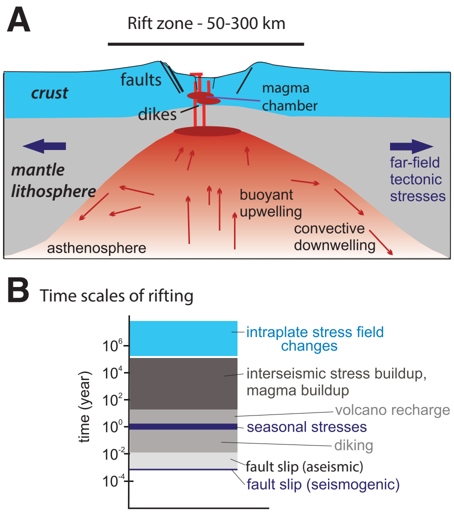

Figure 1. A: A mature rift zone. A thin mantle lithospheric lid underlies a zone of stretched, thinned crust, possibly intruded by magmatic dikes and sills; magmatism results from decompression melting. Asthenospheric convection results in upwelling below the rift, downwelling adjacent to it . B: Time scales of rifting. Changes in the intraplate stress field occur over millions of years; magmatic processes occur on time scales from hundreds to thousands of years; diking events from hours to weeks; and seismogenic fault slip in seconds. (after Ebinger et al., 2013).

segmentation. Accommodation of strain in the lower crust–upper mantle remains debated because of uncertainties in extrapolation from laboratory data to the low rates of extension processes (~10^{-14} s^{-1}), and heterogeneity of continental crust. The timing and rate of rifting processes control the rheology of the lithosphere and rift kinematics.

As stretching continues, the zone of active deformation narrows, and outboard fault systems become inactive. Small-scale convection may facilitate the removal of mantle lithosphere and heating (e.g., van Wijk et al., 2008). The asthenosphere rises to replace the thinned crust and mantle, leading to decompression melting. At rupture, most of the plate boundary deformation is accommodated by intrusion of magma and formation of oceanic lithosphere. The stretched, heated continental crust and any remaining mantle lose heat and subside as the plate boundary shifts to the new spreading center. If forces are not sufficient and/or melt is unavailable, the rift fails, subsides, and is filled with sediments.

Several components remain controversial: How can rifting initiate in cratonic lithosphere that is mechanically stronger than the plate forces ? What controls the strain localization enabling plate rupture?

These questions have led to generations of models, launched cruises along passive continental margins, and fuelled expeditions to remote areas. Subsurface imaging of rifted continental margins separated by ocean basins and failed rifts provides information on the geometry of crustal stretching. The rates of sediment deposition provide constraints on the rates of stretching and heating/cooling processes over millions of years. Studies of active rift zones establish initial conditions and provide rates, kinematics, and dynamics of rifting processes. We do not, however, know whether any one rift zone will form a new ocean basin. Rift geodynamicists turn to plate kinematics to predict plate driving forces millions of years in the future, similar to Heine and Brune's (2014) hindcast modeling of Africa–South America separation.

Studies of active processes enable scientists to understand and calibrate time-averaged deformation. Some deformation in rift zones is achieved through discrete, intense rifting episodes. The far-field and dynamic stresses may build in the elastic crust over 10^1-10^3 yr, to be released in seconds during earthquake rupture. Fault slip may accommodate decades to centuries of the longer-term plate opening. Geologically instantaneous deformation is accommodated by viscous relaxation and ductile creep in the lower crust and/or mantle. The rifting cycle may be modulated by the magma supply at time scales of years to centuries; large magma intrusions may accommodate centuries of plate opening (e.g., Ebinger et al., 2013). The loads on the plates caused by fault slip, ductile deformation, and magmatism are compensated by flexural isostasy over 10^3-10^5 yr. The large-scale lateral density changes due to crust and mantle lithospheric necking, sedimentary basin formation, and magmatism cause vertical and horizontal movements.

Rifts and passive margins are not created equal, with large differences in geometry, kinematics, and time-space evolution. Differences relate to a few key parameters. Thermal and hydration states, and the composition and thickness of the continental lithosphere largely determine where and how strain occurs in response to far-field and dynamic forces (e.g., Kusznir and

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Park, 1987; Lowry and Pérez-Gussinyé, 2011). The magnitude, direction, and rate of the plate-driving forces also influence rift evolution.

An understanding of the physical properties from rift inception to rupture in three dimensions (3-D) remains elusive, however. Geodynamicists probe model space to find the critical processes controlling rift evolution, and to refine and enhance models of time-averaged rifting processes. We identify factors influencing the initiation of rifting through continental rupture: (1) the presence or absence of magma and volatiles, (2) lithospheric heterogeneities, and (3) the obliquity of rift structures relative to the plate-driving forces. We use time-averaged extensional velocity as one measure of the rifting effectiveness, although this is an oversimplification of the actual rifting cycle.

RIFT INITIATION AND RUPTURE: KEY FACTORS

Magma Transport in the Lithosphere

Although most rifted margins experience magmatism prior to breakup, inclusion of mass, heat, and volatile transfer adds considerable complexity to rheologically stratified plate models, so that magmatism usually is not addressed. A new generation of models, however, shows that the presence of magma facilitates rift initiation, and strain accommodation by magma intrusion has been observed to be important during late-stage rifting. Magma may be generated when rifts form above upwelling mantle, or lithospheric thinning is pronounced. Buoyant magma transfers heat and volatiles to the thinning plate as it rises. Magma intrusion locally increases heating, facilitating and/or accelerating strain localization. If melt is generated in the upwelling asthenosphere, dike intrusion can occur at ~10% of the stress required to cause slip on a crustal fault system (e.g., Bialas et al., 2010). Degassing volatiles further decrease mineral strength and enhance strain localization (e.g., Mackwell et al., 1985; Hirth and Kohlstedt, 1996). Thus, magma may make rifting possible in strong cold lithosphere.

Models including the generation and rise of magma with plate thinning predict progressively shallower zones of magma intrusion, accommodating a larger part of plate opening (e.g., Behn et al., 2006; Bialas et al., 2010). The shape of the lithosphere-asthenosphere boundary may localize magma ponding.

Preexisting Heterogeneities and Rift Obliquity

In homogenous isotropic materials, brittle failure initiates along faults perpendicular to the extension direction; along fault planes dipping ~65° from horizontal. On the 100–1000 km scale of rift zones, crustal rocks are not homogenous, with strain fabrics imposed by earlier orogenic or rifting episodes. The distribution of plate boundary deformation and magmatism are influenced by heterogeneities in lithospheric thickness, strength, and composition. Strain and magmatism preferentially localize to pre-rift tectonic boundaries (e.g., Dunbar and Sawyer, 1989; King, 2005). The deep keels of Archean cratons may deflect the flow of mantle upward and laterally around the keels, localizing strain and magmatism at craton margins (e.g., King, 2005), as in the seismically and volcanically active Western and Eastern rift systems of East Africa In many plate-scale simulations, homogenous plate models are seeded with a weak zone to examine sensitivity to rate and state changes.

Strain during rift initiation may localize to an orogenic zone where crust and mantle are comparatively weak and/or more volatile-rich. The geometry of this preexisting weak zone may be oblique to the orientation of extensional forces. During the 10–50 m.y. evolution from rift initiation to rupture, changes in plate configurations may well occur. Thus, individual rift sectors may initiate orthogonal or oblique to extension, or may experience changes in the stress field.

Observations from active rift zones show relationships between extensional velocity and rift obliquity. On the long time scale of 10^6–10^7 yr, rift activity shows a strong relationship to the intraplate stress field resulting from plate boundary forces. Rifts open when the stress field is stable, and are characterized by low interseismic strains (on the order of ~1–5 mm/yr) and limited magmatic activity when the stress field rotates (e.g., Ebinger et al., 2013). Through comparisons of tectonic stress fields, rift structures, and opening rates, the rift community can test the new hypothesis.

FUTURE DIRECTIONS IN RIFT RESEARCH

As longer time series of geodetic and seismicity data become available, and continental lithospheric mapping continues to probe rift zones, we begin to ask: What does time-averaged deformation mean in terms of the physics of continental rifting? Plate-scale models of continental rifting processes make it possible to separate the effects of large-scale plateau uplift, flood volcanism, and tectonics from those of climate change. Rift zones and passive continental margins contain vast water, mineral, and energy reserves. Predictive models of rift basin evolution will guide geothermal energy initiatives, and may lead to more sustainable methods in the extractive industry.

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