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The time scales of continental rifting: Implications for global processes

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ABSTRACT

The rifting cycle initiates with stress buildup, release as earthquakes and/or magma intrusions/eruptions, and visco-elastic rebound, multiple episodes of which combine to produce the observed, time-averaged rift zone architecture. The aim of our synthesis of current research initiatives into continental rifting-to-rupture processes is to quantify the time and length scales of faulting and magmatism that produce the time-averaged rift structures imaged in failed rifts and passive margins worldwide. We compare and contrast seismic and geodetic strain patterns during discrete, intense rifting episodes in magmatic and amagmatic sectors of the East African rift zone that span early- to late-stage rifting. We also examine the longer term rifting cycle and its relation to changing far-field extension directions with examples from the Rio Grande rift zone and other cratonic rifts. Over time periods of millions of years, periods of rotating regional stress fields are marked by a lull in magmatic activity and a temporary halt to tectonic rift opening. Admittedly, rifting cycle comparisons are biased by the short time scale of global seismic and geodetic measurements, which span a small fraction of the 10^2–10^5 year rifting cycle. Within rift sectors with upper crustal magma chambers beneath the central rift valley (e.g., Main Ethiopian, Afar, and Eastern or Gregory rifts) seismic energy release accounts for a small fraction of the deformation; most of the strain is accommodated by magma intrusion and slow-slip. Magma intrusion processes appear to decrease the time period between rifting episodes, effectively accelerating the rift to rupture process. Thus, the inter-seismic period in rift zones with crustal magma reservoirs is strongly dependent upon the

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magma replenishment cycle. This comparison also demonstrates that intense rifting events, both magmatic and amagmatic, produce the long-term fault displacements and maintain the along-axis rift architecture through repeated episodes. The magmatic events in particular accommodate centuries of inter-seismic strain, implying that inter-seismic-plate opening rates in late stage rifts should be extrapolated to the past with caution.

1. INTRODUCTION

Earth's plates stretch and heat at mid-ocean-ridge systems and within the interiors of plates at continental rift zones (e.g., Fig. 1A). The extension occurs in response to far-field plate forces, such as slab pull and ridge push, as well as tractions at the lithosphere-asthenosphere boundary induced by mantle flow. Failed rift zones and associated volcanic rocks scar continental landmasses, documenting the pervasive role of rifting in the chemical and structural evolution of continents since Archean time (e.g., Burke, 1980; Schulte and Mooney, 2005). Some continental rifts may stretch and heat to the point of plate rupture and the onset of subaerial or shallow seafloor spreading (Esedo et al., 2012). The faulted, intruded, and heated continental lithosphere on either side of the spreading center loses heat and subsides as the deformation shifts to a narrow ridge flanked by the widening, subsiding zone of oceanic lithosphere. The post-rift sedimentary strata deposited near or below sea level on the two passive margins record the detailed history of vertical crustal movements after breakup, but these thick sedimentary sequences also mask the syn-rift faults, intrusions, and sediments informing breakup processes (e.g., Steckler and Watts, 1982; Hutchinson et al., 1982; Leroy et al., 2010; Paton, 2012).

Observations of active rifting and incipient seafloor-spreading processes in the Red Sea–Gulf of Aden–East African rift zone, Rio Grande rift, Baikal rift, and the Salton Trough provide new insights, and constrain predictive models of the continental to oceanic rifting process, including the initiation and maintenance

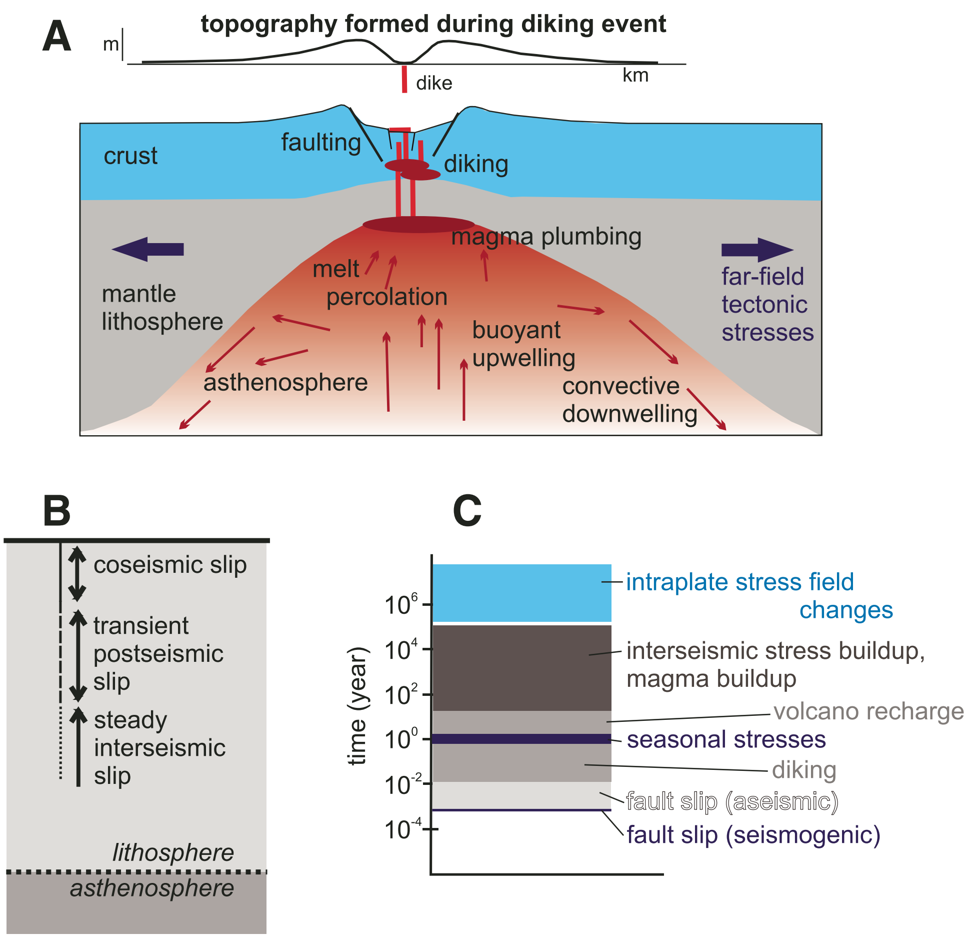


Figure 1. (A) Crustal deformation and upper mantle processes related to an advanced stage of rifting. A thin-mantle lithospheric lid underlies a zone of stretched and thinned crust. This zone is intruded by magmatic dikes and sills; magmatism results from decompression melting. Convection in the asthenosphere results in upwelling below the rift and downwelling adjacent to the rift zone. (B) Geometry of elastic half space and lithosphere-asthenosphere coupling models of strike-slip faulting, from Thatcher (1983). (C) The time scales of rifting. Changes in the intraplate stress field occur on a millions-of-years time scale; magmatic processes such as plumbing, magma buildup in chambers, and volcano recharge generally occur on time scales from hundreds to thousands of years; diking events are of very short duration (hours to weeks); seismogenic fault slip occurs in seconds. Rift opening is an episodic rather than continuous process.

of along-axis segmentation at time scales of hours to millennia. These models, in turn, facilitate the deconvolution of large-scale plateau uplift, flood volcanism, and tectonics from the decadal to millennial-scale climate system (e.g., Courtillot et al., 1999; Spiegel et al., 2007; Sepulchre et al., 2006), as well as the evaluation of mineral, hydrocarbon, and water resources contained in rift basins (e.g., Morley, 1995; Tiercelin, 2009) and along passive margins (e.g., Bois et al., 1982; Brooks, 1990).

Although both continental and oceanic rifts are zones of lithospheric extension, the greater thickness and inherent heterogeneity of continental crust and mantle lithospheric rocks lead to greater complexity in terms of deformation and magmatism; a wide variety of rift zone dimensions, deformation styles, and degrees of magmatic involvement are imaged worldwide (e.g., Buck, 1991; Schlische, 1993; Agar and Klitgord, 1995; Shillington et al., 2009). The majority of successfully rifted margins show evidence for magmatism prior to plate rupture (e.g., Coffin and Eldholm, 1994; Menzies et al., 2002; White et al., 2008; Bronner et al., 2011). Yet, thermo-mechanical models of evolution in heterogeneous continental lithosphere are largely 2D (e.g., Keen, 1985; Dunbar and Sawyer, 1989; Lavier and Manatschal, 2006; Schmeling, 2010; Huismans and Beaumont, 2011), and only a few address the effects of melting in rifting processes (e.g., van Wijk et al., 2001; Bialas et al., 2010; Schmeling and Wallner, 2012).

These rifting models have been developed largely from interpretations of time-averaged deformation patterns, yet many of the observed surface and subsurface structures are achieved through discrete, intense rifting episodes separated by 10^2–10^5 a (e.g., Machette et al., 1991; Wright et al., 2006; Rowland et al., 2007; Wei et al., 2011) (Fig. 1A). The fault-related rifting cycle commences with tectonic stress buildup over decades to centuries (inter-seismic period), followed by seismogenic failure along existing or new faults over periods of seconds (co-seismic), and culminating in relaxation and ductile creep in the lower crust and/or mantle lithosphere for years to decades (post-seismic) (e.g., Savage and Prescott, 1978; King et al., 1988a, 1988b; Foulger et al., 1992; Bürgmann and Dresen, 2008; Freed et al., 2012). Where pressurized crustal magma reservoirs underlie rift basins under extensional stress, magma can intrude the lithosphere as dikes oriented perpendicular to the direction of extension. The buoyant magma pressure adds to the tectonic stress, facilitating dike intrusion at lower stress than fault slip (e.g., Rubin and Pollard, 1988; Buck, 2004; Behn et al., 2006) (Fig. 1A). In rifts with active magmatism, the rifting cycle is modulated by the magma supply cycle, which may have time periods of years to centuries (e.g., Tait et al., 1989; Ebinger et al., 2010; Druitt et al., 2012).

Yet, the short, intense events marking the onset of the rifting cycle are documented in only a few areas worldwide. Unlike intense deformation events at subduction and transform zones, continental rift zone earthquakes almost exclusively initiate in the crust and are rarely larger than M_w 7, meaning that the number of well-determined earthquake source mechanisms is relatively small (e.g., Sibson, 1982; Frohlich and Apperson, 1992; Yang and Chen, 2010; Craig et al., 2011). For example, one empirical seismic moment release versus normal-fault dimension relation is based on analyses of seven normal faulting earthquakes with well-documented surface fault ruptures (Wesnousky, 2008). Likewise, the historic record is very short compared to the long time scales of the earthquake cycle (10^2–10^5 a).

Dike intrusions are the mechanism of oceanic crustal formation, yet their passage through oceanic and continental crust is difficult to detect: small magnitude earthquakes that are rarely detected on global seismic networks (e.g., Einarsson and Brandsdóttir, 1980; Delaney et al., 1998; Rubin et al., 1998; Dziak et al., 2004; Tolstoy et al., 2006). The repeat time of magma intrusions lacking eruptions is even less well constrained, since their unambiguous identification relies on measuring the pattern of surface deformation (e.g., Fialko and Simons, 2001; Biggs et al., 2009; Grandin et al., 2009; Pagli et al., 2012). The advent of space geodetic methods during the last ~15–20 years are now revealing critical constraints on the frequency and distribution of these sometimes seismically "quiet" events, as outlined in Section 3 below.

The aim of our synthesis of current research initiatives into continental rifting processes is to characterize the time and length scales of faulting and magmatism that produce the time-averaged rift structures imaged worldwide. Our syntheses are intended to inform the development and parameterization of 4D numerical models of heterogeneous continental lithosphere, and in the design of geodetic, rock mechanics, and seismic experiments. These results, in turn, inform our understanding of lithospheric rheology, lithosphere-asthenosphere interactions, geohazards, epithermal mineralization associated with magmatism, geothermal and hydrocarbon energy, and sedimentary and water resource assessment (e.g., John, 2001; Bois et al., 1982; Stein et al., 1991; Assumpção, 1998; Twichell et al., 2009).

2. BACKGROUND

2.1. Review of Rifting Processes

Continental rift zones are sites of lithospheric stretching. The crustal stretching is achieved through normal faulting that thins the brittle crust; a weak lower crust may thin via viscous flow (e.g., Lavier and Manatschal, 2006). Comparatively dense mantle lithospheric rocks rise upward to replace the thinning crust, enhancing subsidence in the fault-bounded basin (e.g., McKenzie, 1978; Weissel and Karner, 1989) (Fig. 1A). The mantle lithosphere also necks and thins. Replacement of mantle lithosphere by hotter asthenosphere transfers heat to the lithosphere beneath the stretched zone, reducing rock density, leading to a regional, time-dependent uplift over time scales of tens of millions of years (e.g., Şengör and Burke, 1978; Keen, 1985). The passive mantle upwelling may be enhanced by anomalously hot mantle upwellings or plumes and/or small-scale convection induced by the steep gradients at the transition between thinned and unthinned lithosphere (e.g., Buck, 1986; King and Anderson, 1998; van Wijk et al., 2001, 2008) (Fig. 1A). In all of these cases, the distribution of plate boundary deformation and magmatism also may be influenced by preexisting heterogeneities in lithospheric thickness, strength, and composition; strain and magmatism preferentially localize to pre-rift tectonic boundaries and deviate around deeply rooted Archean cratons (e.g., Dunbar and Sawyer, 1989; Petit and Ebinger, 2000; Corti et al., 2007).

Where the mantle advectively rises beneath thinning lithosphere to pressures equivalent to ~90 km subsurface, basaltic melts may be generated by adiabatic decompression melting. The volume and distribution of melt produced in space and in time from plate stretching will depend on the volatile content of the mantle and the vertical component of velocity, which depends on the geometry and rate of thinning (e.g., Gallagher and Hawkesworth, 1994; Tackley and Stephenson, 1993; Bown and White, 1995). Considerable melt volumes can be produced along steep gradients in the mantle lithosphere, either at the boundaries of the thinned and unthinned lithosphere, and/or at preexisting lithospheric thickness variations, such as the edge of cratonic roots (e.g., King and Anderson, 1998; King and Ritsema, 2000). Where rifts initiate above or near elevated mantle temperatures or a mantle plume, the anomalously high temperatures facilitate melting at deeper levels, producing large volumes of flood volcanism and magma intrusion at high eruption rates (e.g., White and McKenzie, 1989; Coffin and Eldholm, 1994). Prolonged and voluminous magmatism may occur over millions of years, depending upon the rate of plate movement over the hot, volatile-rich plume source and plume interactions with heterogeneous lithosphere (e.g., Self et al., 1997; Ebinger and Sleep, 1998). The feedbacks between heat transfer to the plate from the asthenospheric melt, and its percolation pathways through the plate may also thin and soften the plate in a bottom-up process (e.g., Tackley and Stephenson, 1993; Buck, 2004; Bialas et al., 2010). Small volume melts, such as kimberlites, carbonatites, and hyper-alkalic rocks, may be generated at deeper levels, where mantle lithosphere is enriched in volatiles, or where the upwelling mantle brings volatiles to the base of the lithosphere, causing metasomatism (e.g., Lloyd et al., 1987; Chesley et al., 1999; Fischer et al., 2009). Melt extraction beneath and at the base of the plate, and melt retention within the thinning plate, alter the plate rheology (e.g., Choblet and Parmentier, 2001; Bialas et al., 2010; Schmeling and Wallner, 2012), and they permanently change its composition and density structure (e.g., Tackley and Stephenson, 1993; Rooney et al., 2012a).

The stretching and heating with or without magmatism moderate the time-space distribution of surface deformation and volcanism, as well as rates and planform of subsidence and uplift patterns (e.g., Şengör and Burke, 1978; Newman and White, 1999; Buck, 2004; Bialas et al., 2010). After the cessation of rifting, the heat gained by the plate dissipates, and the thinned plate beneath a failed rift or passive margin subsides (e.g., Sleep, 1971; McKenzie, 1978). New geophysical and geochemical data and numerical models are providing insights into the bottom-up lithosphere-asthenosphere interactions controlling the sources and pathways of magmatism beneath rifts (e.g., van Wijk et al., 2008; Bialas et al., 2010; Keir et al., 2011; Rychert et al., 2012; Schmeling and Wallner, 2012), as well as the evolving composition of the mantle lithosphere (e.g., Graham et al., 2006; Blichert-Toft et al., 2005; Rooney et al., 2012a).

Several continental rift zones on Earth are seismically and volcanically active, meaning that the time and length scales of the process can be deduced from direct observation. One class is orogenic rifts within late-stage collisional belts where the crust has been thickened and heated by thrust faulting and/or magmatism, leading to a broad zone of extensional deformation within weak lithosphere (e.g., Buck, 1991; Brun, 1999; Lavier and Manatschal, 2006). Active examples are the Basin and Range Province that extended thickened crust (e.g., Hamilton, 1987; Buck, 1991; Kreemer and Hammond, 2007), the Gulf of California, where rifting initiated immediately after arc volcanism (e.g., Stock and Hodges, 1990; Dorsey and Umhoefer, 2012), and the incipient rift zones of the Apennines (e.g., Jolivet et al., 1998). A second class is cratonic rifts underlain by relatively strong crust and thick lithosphere far from subduction or collisional zones, leading to strain localization in a single rift zone with narrow, deep basins (e.g., Weissel and Karner, 1989; Buck, 1991). Examples include the Baikal rift in central Asia, the Rio Grande rift in central North America, and the East African Rift System.

2.2. Along-Axis Segmentation

Although textbook images and models of the continental rupture process stem from two-dimensional transects of passive continental margins and rift zones, active and ancient continental rift zones are fundamentally three-dimensional. From inception, rift zones show regular along-axis structural segmentation into basins bounded on one or both sides by large offset border faults, with sedimentary strata recording individual horizontal and vertical movements (e.g., Bosworth, 1985; Rosendahl et al., 1986; Lambiase, 1990; Scholz et al., 1990). The border faults have the largest dimensions of the fault populations, and they are flanked by broad uplifts that may rise 3 km above the surrounding regional elevations (e.g., Fig. 2). This along-axis segmentation effectively partitions stratigraphic sequences and strongly influences erosion and drainage patterns in rifts (e.g., Gawthorpe and Leeder, 1997; Mats et al., 2000; Densmore et al., 2004; Dorsey and Umhoefer, 2012). Initially discrete border-fault segments interact and are mechanically connected through transfer faults and relay ramps oriented obliquely to the strikes of border faults (e.g., Larsen, 1988; Morley and Nelson, 1990; Walsh and Watterson, 1991; Peacock and Sanderson, 1994). Rift segments may grow or link through along-axis propagation and interaction of faults and magmatic plumbing systems (e.g., Ebinger et al., 1989; Densmore et al., 2004; San'kov et al., 2000, 2009; Beutel et al., 2010; Allken et al., 2012). Transfer fault zone geometries may change, or be abandoned, during linkage (e.g., Morley and Nelson, 1990; Schlische, 1993; Cartwright et al., 1995; Densmore et al., 2007). The role of magmatism in segment linkage during the early rifting stages remains poorly understood owing to the general lack of detailed subsurface data (e.g., Ebinger et al., 1989; Rowland and Sibson, 2001), but clear patterns are seen in mature rift zones, as outlined below.

Although major faults bounding and linking basins may reactivate preexisting fault zones or crustal fabrics (e.g., Sibson, 1985; Versfelt and Rosendahl, 1989; Smith and Mosley, 1993; Bellahsen and Daniel, 2005), border fault, rift flank, and basin dimensions increase with increasing strength of the lithosphere (e.g., Weissel and Karner, 1989; Ebinger et al., 1999). Rebound of the shoulders, and bending of the hanging wall at a wavelength that scales with plate strength, are the elastic response of the plate to unloading of the footwall by repeated single episodes of fault slip, or many cycles of slip to produce deep basins (e.g., Vening-Meinesz, 1941; King et al., 1988a, 1988b; Weissel and Karner, 1989) (Fig. 2). The mechanical strength, in turn, depends on lithospheric structure, composition, geothermal gradient, volatile content, as well as on the rates of the applied processes (e.g., Watts and Burov, 2003; Freed et al., 2012). Where the lower continental crust is quartzo-feldspathic or with elevated geotherms, extensional stress is accommodated by creep rather than by faulting, and the lower crust deforms aseismically (e.g., Sibson, 1982; Kusznir and Park, 1986; Chen and Molnar, 1983; Wells and Coppersmith, 1994; Niemi et al., 2004) (e.g., Fig. 1B). Normal faults in the Basin and Range Province and other collapsing orogens, and along magma-poor margins, have produced soles into the weak lower crust (e.g., LePichon and Sibuet, 1981; Buck, 1986; John and Howard, 1995; Abers, 1991).

In cratonic rift segments lacking evidence of volcanism, the largest earthquakes initiate in the lower crust, indicating that the lower crust can support significant stress, and that the border faults reach the lower, probably gabbroic, crust (e.g., Shudofsky et al., 1987; Yang and Chen, 2010; Craig et al., 2011). Local seismicity studies show no gap in seismicity within the mid-crust,

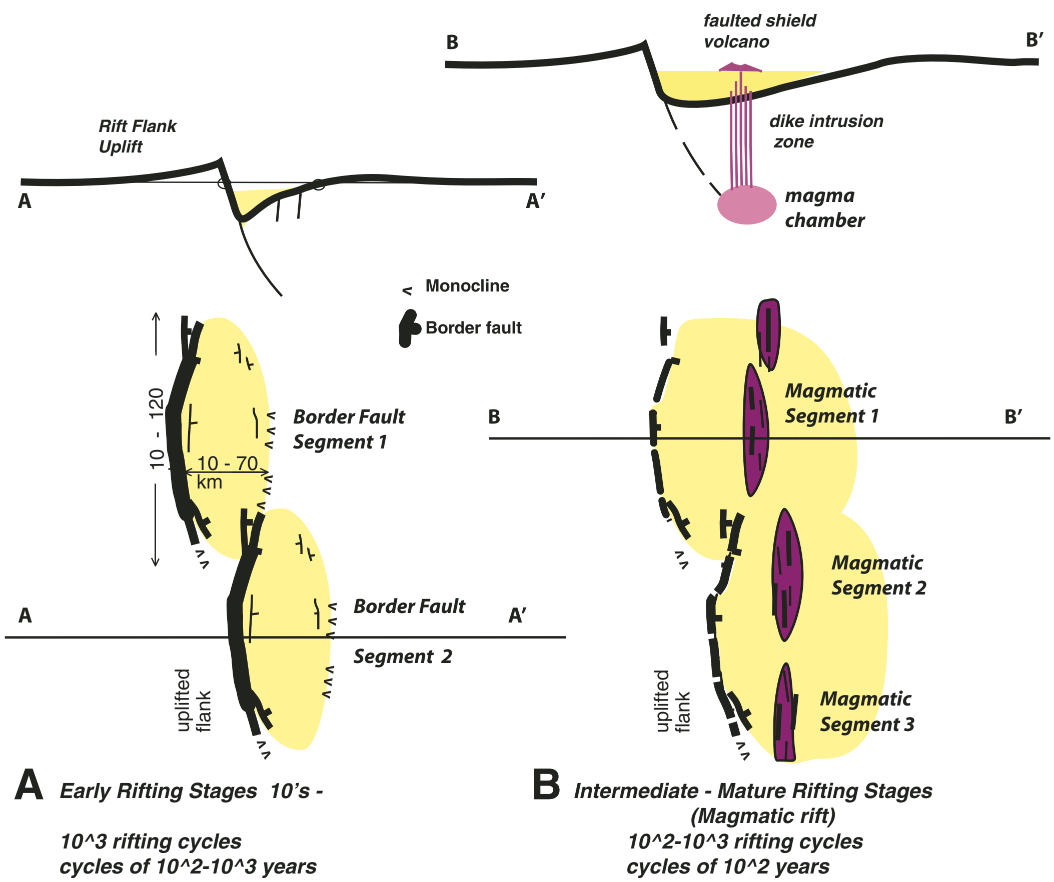


Figure 2. Cross sections (top) and plan form drawings of along-axis continental rift segmentation. (A) Early to middle-stage amagmatic or weakly magmatic rift zones are bounded along one or both sides by border faults, the longest fault with the maximum displacement in each basin segments. Strain patterns in discrete basins are linked via relay ramps and/or faults with a wide variety of orientations and transient geometries. (B) Along-axis segmentation in mature magma-rich rift zones. Strain has localized to the zones of magma intrusion fed from shallow crustal reservoirs, and distributed along the length of magmatic segments via frequent dike events. Border faults are largely inactive: the interseismic cycle is long. The magma feeding system largely controls the rates of rift opening. Adapted from Ebinger et al. (1999).

consistent with a low geothermal gradient and mafic lower crustal rheology (e.g., Camelbeeck and Iranga, 1996; Déverchère et al., 2001; Mulibo and Nyblade, 2009; Albaric et al., 2010). Cratonic rifts with eruptive centers within basins show seismicity restricted to the upper 15 km (e.g., Ibs-von Seht et al., 2001; Ayele, 1995; Keir et al., 2006), consistent with source depths of the largest earthquakes (e.g., Yang and Chen, 2010; Craig et al., 2011). Clusters of earthquakes in the upper mantle beneath the northern sector of the Western rift, East Africa, are interpreted as brittle failure in response to locally high stress rates caused by transport of magmatic fluids through the mantle lithosphere (Lindenfeld and Rumpker, 2011), consistent with metasomatized mantle xenoliths found in nearby Quaternary eruptive fields (e.g., Lloyd et al., 1987; Furman, 1995).

Observations from amagmatic and magmatic margins demonstrate clear differences in the rift to rupture process. Where magma is absent until the onset of seafloor spreading, rift segments widen to as much as 5 times their original widths, drastically reducing plate strength (e.g., Hopper et al., 2004; Reston and Pérez-Gussinyé, 2007; Lizarralde et al., 2007; Péron-Pinvidic and Manatschal, 2010; van Avendonk et al., 2009). The presence of magma during rifting decreases the amount of stretching required to achieve plate rupture, as can be seen in the much narrower widths of conjugate magmatic margins worldwide (e.g., Coffin and Eldholm, 1994; Menzies et al., 2002; Shillington et al., 2009; Leroy et al., 2010). As rifting progresses to seafloor spreading, strain localizes to aligned chains of eruptive centers and smaller faults within the central basin that are fed from crustal reservoirs beneath each segment, and the border faults become inactive (e.g., Hayward and Ebinger, 1996; Keranen et al., 2004; Keir et al., 2006; Beutel et al., 2010) (Fig. 2B). This new, shorter magmatic segmentation persists over periods of ~1 m.y. or more, suggesting that segment centers receive an enhanced melt supply, as along slow-spreading ridges (e.g., Keir et al., 2006; d'Acremont et al., 2005; Keir et al., 2009; Belachew et al., 2011). The length scales of faults and eruptive centers, however, may be longer in rift zones dominated by transtension (e.g., Scrutton, 1982; Axen, 1995; Umhoefer et al., 2002).

Post-rift subsidence patterns indicate that extension accommodated by brittle faulting is less than the total thinning, suggesting that the upper crustal and lower crustal stretching are decoupled (e.g., Karner et al., 2003; Davis and Kusznir, 2004). Along magma-poor margins, the discrepancy between strain accommodated by faulting in the upper crust and deeper levels of the lithosphere may be explained by magma intrusion, which also accommodates plate opening (e.g., Buck, 2004; Péron-Pinvidic and Manatschal, 2010), or by mantle lithosphere detachments (e.g., Esedo et al., 2012). Alternatively, thermo-mechanical models of rheologically layered continental lithosphere indicate that the discrepancy can be explained by concave downward faults that exhume lower crust and/or upper mantle (e.g., Lavier and Manatschal, 2006).

These along-axis segmentation patterns are fundamentally different from the ~300 km separation between transform faults (first-order spacing) and magma feeding systems of a length scale of ~50 km (second-order spacing) observed along mid-ocean-ridge systems worldwide (e.g., MacDonald et al., 1991; MacDonald, 1998; Briais and Rabinowicz, 2002). Numerical models of mantle flow beneath slow-spreading ridges that take into account viscosity changes accompanying melt extraction develop upwellings with spacings of ~70 km, similar to the second-order magmatic segmentation observed at spreading ridges (e.g., Parmentier and Phipps-Morgan, 1990; Magde et al., 2000; Choblet and Parmentier, 2001). The spacing of the first-order, transform boundary segmentation is strongly influenced by thermal stresses in the cooling lithosphere, although details of the 3D process and time-length scales of hydrothermal cooling remain to be explored (e.g., Haxby and Parmentier, 1988; Sandwell and Fialko, 2004; Choi and Gurnis, 2008; Coumou et al., 2008). Thus, oceanic transforms are a dynamic process inherent to the formation of new columns of oceanic lithosphere at ridges, whereas obliquely trending transfer faults in continental rifts are transient features linking finite-length normal faults that accommodate most of the strain, at least until magma intrusion becomes equally important.

2.3. The Rifting Cycle

Satellite geodetic, seismic, and remote sensing studies of continental rift zones document segment extension through seismogenic slip along faults and/or magmatic intrusions (e.g., Machette, 1986; Wright et al., 2006; Rowland et al., 2007; Grandin et al., 2009; Berglund et al., 2012). These periods of intense activity, however, represent only one part of the rifting cycle of stress buildup, release, and visco-elastic rebound, multiple episodes of which combine to produce the observed, time-averaged rift zone architecture (Fig. 1A). Elastic rebound and the slower isostatic compensation of density changes that accompany repeated slip and magma intrusion and extrusion events produce the characteristic along-axis segmentation of rift basin and flank morphology.

The moment magnitude, M_0, of an earthquake scales as the slip area of the fault zone according to $M_0 = \mu LWD$, where μ is frictional coefficient, L and W are the strike length and down-dip width of the slip plane, respectively, and D is the displacement (Aki, 1966). Rarely are all of L, W, and D known, and we rely on empirical relations derived from comparisons of earthquake magnitude scaling and fault length–displacement relations (e.g., Cowie and Scholz, 1992a, 1992b; Wells and Coppersmith, 1994; Wesnousky, 2008). The length-displacement ratios of the relatively few well-imaged faults can be fit by power-law relations (e.g., Cowie and Scholz, 1992a, 1992b; Walsh and Watterson, 1991), whereas Ackermann et al. (2001) suggest that border faults that penetrate the brittle layer thickness evolve to an exponential relationship. The catalogue of well-determined fault lengths and displacements includes faults in sedimentary, metamorphic, and igneous rocks with Young's moduli varying by an order of magnitude. Different scaling laws may apply to faults that initiate in different rock rheologies, including faults above dike intrusions where fluids are involved, as discussed in Section 3. The growing catalogues of earthquake source mechanisms, co- and post-seismic deformation patterns, surface deformation and gas emissions during volcanic eruptions, and volcano inflation-deflation signatures are enabling separation of faulting and magma movement processes, and the lithospheric response to these strains, over time scales of seconds to millennia, as outlined in Sections 3, 4, and 5.

Given our focus on the discrete time steps of deformation producing the observed rift architecture, we review current progress into rift evolution over time scales of the rifting cycle (hours to centuries) and over the 30–100 km length scales of individual rift segments, using examples from magmatic and amagmatic rifting episodes. We then examine the evolution of rifts from the initial growth and linkage of border faults spanning perhaps 10^2–10^3 rifting cycles. We take this approach to enable assessment of contributions from multiple processes producing the range of rift architecture outlined above. Examples from magma-rich and magma-poor segments of the East African, Red Sea, and Gulf of Aden rift zones in Ethiopia, and the weakly magmatic Rio Grande rift, illustrate the time and length scales of episodic rifting processes.

3. EAST AFRICAN RIFT ZONE

The Red Sea, Gulf of Aden and Main Ethiopian rifts that meet in the Afar triple junction transect the Ethiopia-Yemen Plateau, whereas the Eastern, Western, and Southwestern rift systems transect the broad East and Central African Plateaus (Figs. 3, 4). The depression between the two plateaus marks a failed Mesozoic rift system, allowing the possibility that the plateaus are actually one uplift extending from southern Africa to the Red Sea, the African superplume province (e.g., Nyblade and Robinson, 1994; Ritsema et al., 1998; Simmons et al., 2007). The

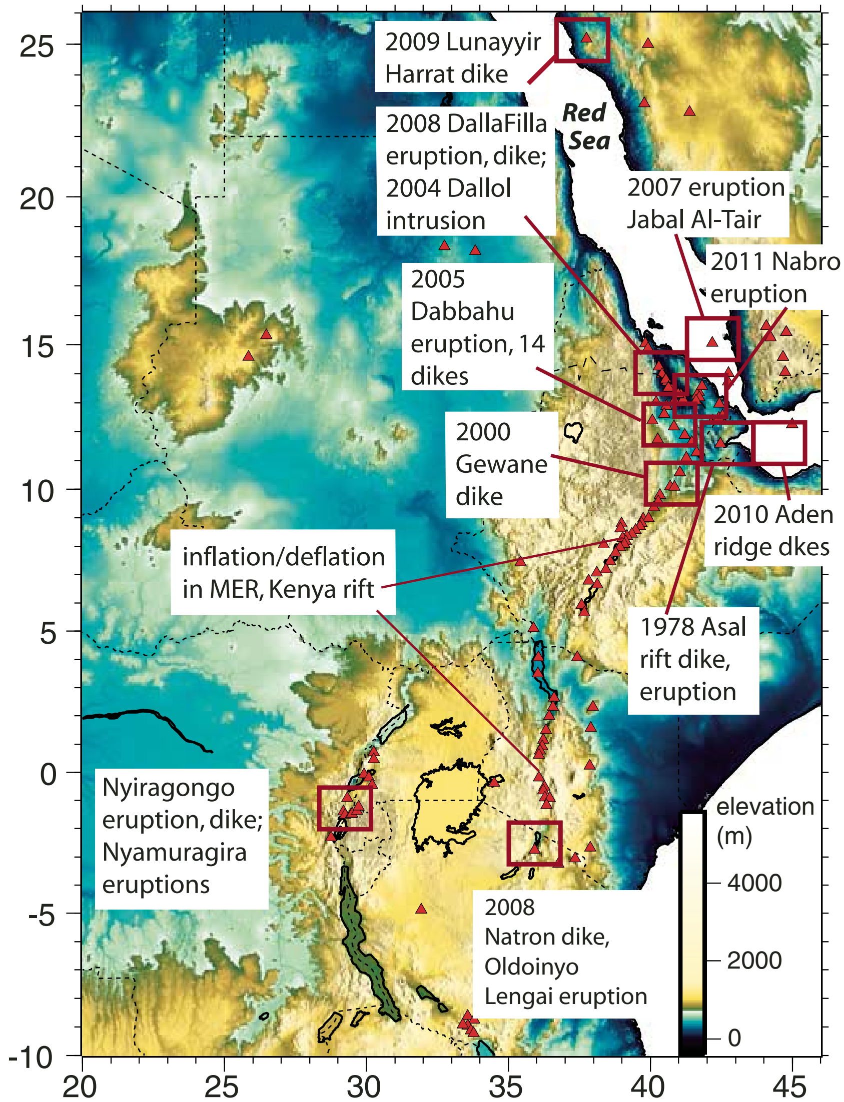
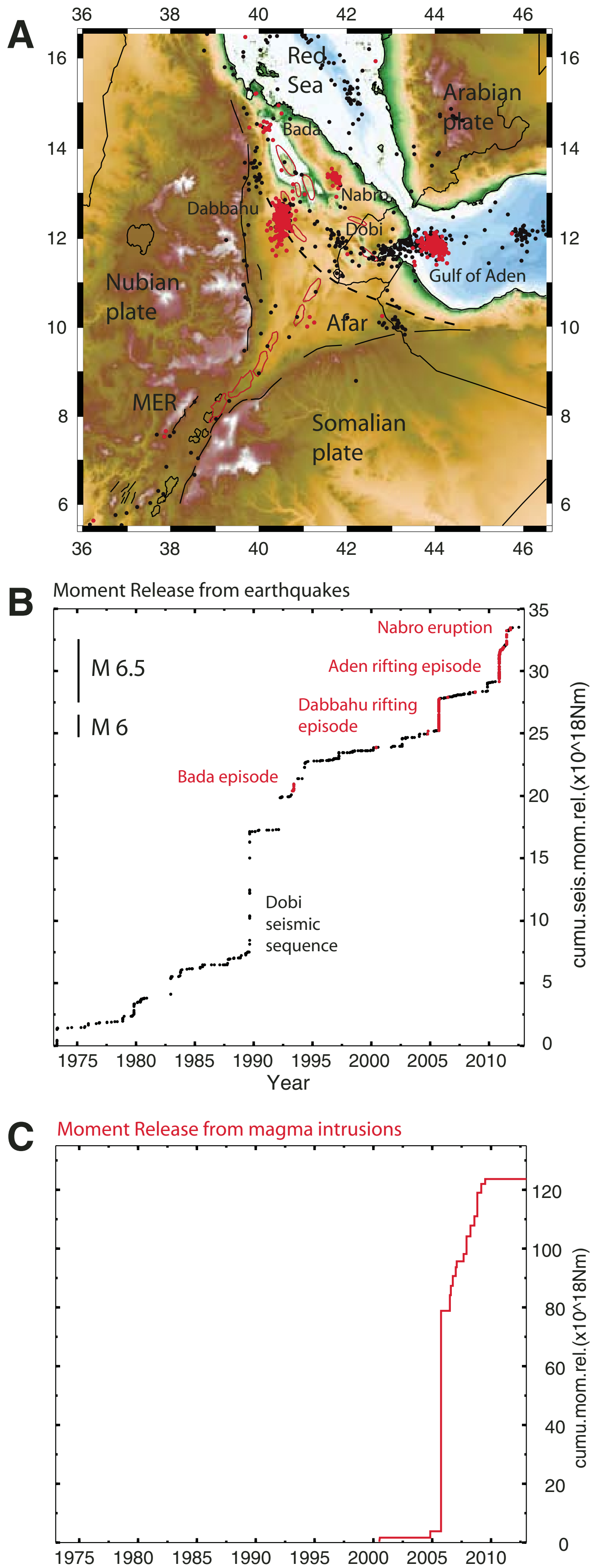


Figure 3. Compilation of confirmed magma intrusion and volcanic eruption events within the East African Rift System. Holocene volcanoes from the Global Volcanism program (www.si.edu). From north to south: 2009 Lunayyir Harrat dike intrusion (Pallister et al., 2010); 2007 Jabal Al-Tair eruption (Carn); 2008 DallaFilla rifting episode (Pagli et al., 2012); 2011 Nabro eruption (Oppenheimer et al., 2011); 2004 Dallol dike intrusion (Nobile et al., 2012); 2005–2011 Dabbahu rifting episode (e.g., Yirgu and Ayele, 2006; Ayele et al., 2007; Belachew et al., 2011); 2010 Gulf of Aden submarine rifting episode (Shuler and Nettles, 2012; Ahmed et al., 2012); 1978 Asal rifting episode (Abdallah et al., 1979); 2000 northern Main Ethiopian dike intrusion (Keir et al., 2011); central Main Ethiopian Rift magma inflation episodes (Biggs et al., 2011); Eastern rift magma inflation episodes (Biggs et al., 2009); 2002 Nyiragongo eruption and dike intrusion (e.g., Tedesco et al., 2007; Wauthier et al., 2012); multiple Nyiragongo and Nyamuragira eruptions.



geochemistry of Eocene–Recent eruptive volcanic products points to a mantle plume origin for the Ethiopia-Yemen flood basalt sequences, but the signal is less clear within the East African Plateau region south of Ethiopia (e.g., Pik et al., 2006; Furman, 1995; Chakrabarti et al., 2009; Rooney et al., 2012b). The earliest basaltic volcanism in the East African Rift System occurred in southwestern Ethiopia and northernmost Kenya between 45 and 39 Ma (e.g., Morley et al., 1992; Ebinger et al., 1993) (Fig. 3). During this same period, kimberlites were emplaced in Archean lithosphere between the Western and Eastern rift (Harrison et al., 2001), and flanking the Southwestern rift (Batumike et al., 2007). Their widespread distribution suggests that heating and local metasomatism preceded any surface expression of rifting across the broad, uplifted plateaus. Extension postdated or was concurrent with the massive outpouring of basaltic and felsic lavas in the future sites of the Red Sea, easternmost Gulf of Aden, and the central Ethiopian Plateau in the Afar Depression (e.g., Baker, 1986; Hofmann et al., 1997). In the southern Red Sea, the highest magma production rates at ca. 30 Ma coincided with the onset of rifting (Wolfenden et al., 2005).

The highly extended crust between the two plateaus was rifted during Mesozoic time, and again at ca. 25 Ma, making this the site of the earliest known extension south of the Afar Depression (e.g., Morley et al., 1992). Magmatism commenced at ca. 16 Ma in the northernmost part of the Eastern rift, roughly coeval with the initiation of rift fault systems (e.g., Baker, 1986). In the Western rift, basin development in the northern sector had begun by ca. 14.5 Ma (Roller et al., 2010), and magmatism had begun by ca. 12 Ma (Kampunzu et al., 1998). Carbonatitic volcanism at 25 Ma in the Western rift region suggests that plume-lithosphere interactions may have commenced earlier (Roberts et al., 2012).

Figure 4. (A) Distribution of seismicity in the Afar triple junction region reported by the National Earthquake Information Center since 1973. Red dots represent earthquakes associated with magma-intrusion episodes, whereas black dots represent those with no documented magmatic activity. Rift segments and volcanoes that have undergone major deformation episodes are labeled (e.g., Nabro and Dobi). MER—Main Ethiopian Rift. (B) Cumulative seismic moment release since 1973, with color coding and labeling as used in the top panel. Note that the most significant seismic moment release occurred during the 1989 Dobi sequence where 21 earthquakes of magnitude 4.5–6.2 over a period of two days ruptured a rift sector with no recent volcanoes. Major episodes of faulting induced by magma intrusion include the Dabbahu and western Gulf of Aden rifting episodes. (C) Estimate of the cumulative geodetic moment release reported from space geodetic measurements of discrete intrusions. These include the May 2000 dike in the northern Main Ethiopian Rift (Keir et al., 2011), the October–November 2004 dike near Dallol in northern Afar (Nobile et al., 2012), the multiple intrusions during the 2005–2010 Dabbahu rifting episode (e.g., Keir et al., 2009; Grandin et al., 2009, 2011; Hamling et al., 2009; Belachew et al., 2011). Note that intrusions with no constraints on the amount of rift opening, such as the November 2010 western Gulf of Aden rifting episode and the June 2011 intrusion and resultant eruption at Nabro volcano, are not included.

4. TIME SCALES OF SECONDS TO YEARS: ACTIVE DEFORMATION IN MAGMA-RICH AND MAGMA-POOR CRATONIC RIFTS

Among the historically active continental rifts, the East African Rift zone is arguably the most active, with seven M_w >7 earthquakes since 1973, 22 documented magmatic intrusions, and at least 136 historical eruptions before 1994 (Simkin and Siebert, 1994) (Figs. 3–5). At least 20 eruptions have occurred at Nyiragongo volcano since 1884, and at least 42 at nearby Nyamuragira since 1880; 19 eruptions have occurred at Oldoinyo Lengai in the Eastern rift since 1880; and 7 have occurred at Erta'Ale volcano, including a continuous lava lake since 1967 (Global Volcanism Program, www.si.edu; Mavonga et al., 2007; Shuler and Ekström, 2009; Wauthier et al., 2012). Below we provide a brief context for rifting in East Africa, and then summarize the magmatic and amagmatic rifting cycles in Afar. We can evaluate strain patterns from analyses of historic records of earthquake activity, which span about a century of the 10^2–10^5 a rifting cycle. Here we summarize earthquake activity from 1973 to 2012, using the U.S. Geological Survey's National Earthquake Information Center (NEIC) catalogue (Fig. 4). This catalogue is complete to about M_w 4.5 (e.g., Midzi et al., 1999; Ayele, 1995). The intrusions are constrained by field observations and Interferometric Synthetic Aperture Radar (InSAR) measurements, from which we obtain estimates of the total geodetic moment release, a measure of strain over the deforming volume (e.g., Thatcher, 1984; Ward, 2002).

4.1. Tectono-Magmatic Rifting Cycles: Mature Rift Zones

The Afar triple junction zone is the most advanced sector within the East African Rift System, and parts of the western Gulf of Aden and southern Red Sea are sites of incipient seafloor spreading (e.g., Abdallah et al., 1979; Hayward and Ebinger, 1996; Manighetti et al., 1997; Doubre et al., 2007; Keir et al., 2009). More than 30 m.y. of magmatism and extension above or near a mantle plume led to crustal thinning from ~40 km to ~25 km across the broad zone, and ~18 km beneath magmatic segments (Berckhemer et al., 1975; Tiberi et al., 2005; Dugda et al., 2007; Stuart et al., 2006; Hammond et al., 2011). Stretching and advection have variably thinned (e.g., Bastow et al., 2008; Hammond et al., 2013; Bastow and Keir, 2011) or removed the mantle lithosphere (Rychert et al., 2012) beneath the southern Red Sea rift in Afar. Geochemical data indicate temporal changes in mantle lithospheric thickness from onset of flood magmatism to the present day (Vidal et al., 1991; Rooney et al., 2012a).

The Afar depression is the site of historic volcanic eruptions from 13 volcanic complexes (DallaFilla, Erta'Ale, Dubbi, Nabro, Manda-Inakir, Dama'Ale, Ardoukoba, Alayta, Dabbahu, Manda-Hararo, Fantale, Kone) (e.g., Abdallah et al., 1979; Gouin, 1979; Audin et al., 1990; Wiart and Oppenheimer, 2000; Lahitte et al., 2003; Yirgu and Ayele, 2006; Ferguson et al., 2011; Oppenheimer et al., 2011; Pagli et al., 2012) and numerous large dike intrusions (Figs. 3, 4). Nearly all of these active volcanoes and dike intrusions lie along ~60-km-long magmatic segments, which are the loci of active deformation (e.g., Barberi and Varet, 1977; Hayward and Ebinger, 1996; Manighetti et al., 1997) (Fig. 4). The magmatic segmentation postdates the ca. 1–3 Ma Stratoid basalts, which covered much of the Afar depression (e.g., Barberi and Varet, 1977; Kidane et al., 2003; Lahitte et al., 2003) and may have initiated as recently as ca. 70 ka (Williams et al., 2009; Medynski et al., 2013). Strain also occurs in a higher relief, narrow graben in the complex zone between the Red Sea and Gulf of Aden rifts; some or all of these graben may have been magmatic segments at 10^5 y. BP (Lahitte et al., 2003).

Since 1973, the cumulative seismic moment for the Main Ethiopian Rift, Afar, and nearby Gulf of Aden to 46.5°E, and the Red Sea to 16.5°N has totaled 3.3×10^{19} Nm, equivalent to a single M_w 7 earthquake (Fig. 4). The earthquake records reveal two types of intense rifting events: repeated dike intrusion and faulting events spanning days to weeks in magmatic segments (e.g., Abdallah et al., 1979; Ayele et al., 2009; Pagli et al., 2012; Shuler and Nettles, 2012), and intense faulting events spanning hours in segments lacking Holocene eruptive centers (Sigmundsson, 1992; Jacques et al., 1999, 2011), as outlined below. Magma intrusion and/or extrusion with no distinguishable extension signal occurred during the November 2010 overspill of Erta'Ale volcano (Field et al., 2012). Likewise, there was no detectable extension associated with inflation and deflation cycles of Corbetti, Aluto, Bora, and Haledebi volcanoes in the Main Ethiopian Rift (Biggs et al., 2011), or during the September 2007 eruption of Jabal Al-Tair in the southern Red Sea (Figs. 3, 4).

4.2.1. Magmatic Cycle

The first major rifting cycle documented by space-based geodesy was the 2005 Dabbahu rifting episode (Yirgu and Ayele, 2006) (Figs. 3–5). Analyses of InSAR data demonstrate that a 65-km-long, subaerial segment of the southern Red Sea rift widened by up to 8 m (Wright et al., 2006; Ayele et al., 2009; Grandin et al., 2009). This episode continued for at least 6 more years, with at least 13 additional ~1–3-m-wide dike injections and fissural eruptions that distributed strain along all or parts of the segment's length (e.g., Hamling et al., 2009; Ebinger et al., 2010; Wright et al., 2012). The 14 dike intrusions were sourced from a magma chamber in the mid- to lower crust at the center of the ~65-km-long segment (e.g., Hamling et al., 2009; Grandin et al., 2010; Belachew et al., 2011), although some magma was sourced from Dabbahu volcano at the northern end of the segment during the first, and volumetrically largest, intrusion (e.g., Wright et al., 2006; Grandin et al., 2009). Field measurements and comparative imagery document up to 3–5 m of surface displacement across a 3–5-km-wide zone, and >3 m of uplift on either side of the dike intrusion (Wright et al., 2006; Rowland et al., 2007; Grandin et al., 2009) (Figs. 1, 5).

Relating the time scales of intense magmatic-tectonic rifting episodes to the rifting processes outlined in Section 2, dike intrusions and faulting above the dikes caused opening along

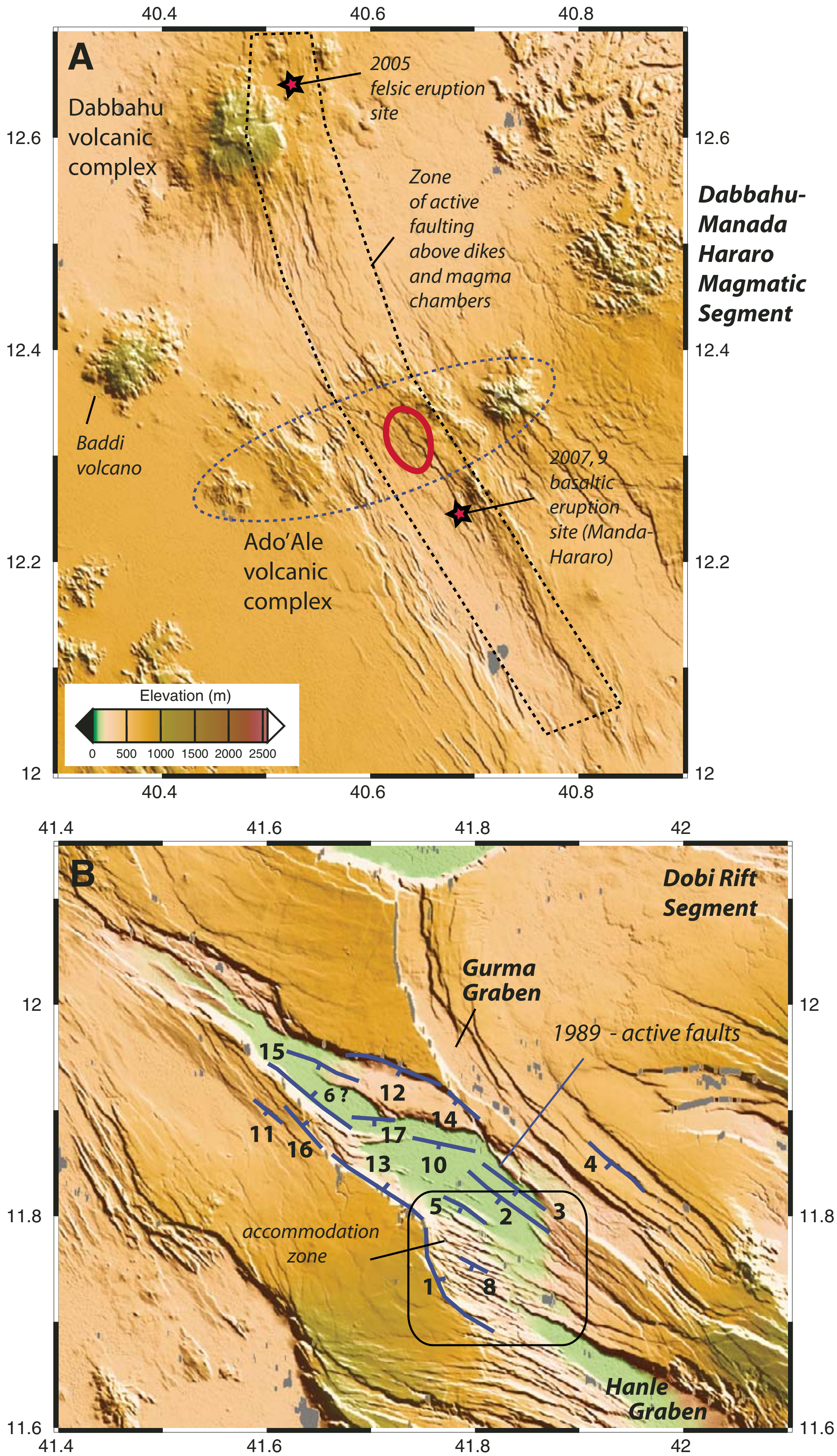


Figure 5. Comparison of digital elevation models (DEMs) of the magmatically active Dabbahu rift segment, and the magma-poor Dobi rift segment (see Fig. 6 for locations). Both rifts formed after the Stratoid sequence was erupted at 1–3 Ma, and they encompass the segments with the greatest seismic moment release (Fig. 4). Cosmogenic isotopic ages indicate that the Dabbahu segment initiated after ca. 70 ka (Williams et al., 2009; Medynski et al., 2013), and with two distinct sources beneath Dabbahu volcano and the mid-segment (Medynski et al., 2013). (A) The Dabbahu magmatic segment is characterized by low topographic relief with no large offset faults. The dotted blue line encompasses the zone of surface faulting and fissuring associated with the 2005–2011 Dabbahu rifting episode (after Rowland et al., 2007; Grandin et al., 2009). Fourteen dikes were sourced from a magma chamber in the mid- to lower crust, and the first (largest volume) dike was also sourced from Dabbahu volcano, at the northern end of the segment. The red oval outlines an aseismic feeder zone for the dikes. Stars indicate sites of fissural eruptions in 2005, 2007, and 2009. Black ellipse encloses the Addo'Ale Volcanic Complex, a chain of rifted eruptive centers. The 50 ka date for the base of the Baddi complex indicates that the magmatic segment has been centrally fed over tens to hundreds of rifting cycles (Williams et al., 2009). (B) Comparatively high topographic region, with several >10-km-long, large offset faults, and highly rotated fault blocks in the transfer fault zone at the southern end of the segment, as enclosed in the black rectangle. Blue lines are fault ruptures associated with the 1989 earthquake sequence, with numbers referring to each of the teleseismic earthquakes (after Jacques et al., 2011).

all or parts of the magmatic segment over a period of hours to days. During these intense "episodes," nearly all of the rift widening was accommodated by dike intrusions, slip along shallow faults above the dikes (Barisin et al., 2009; Hamling et al., 2009; Belachew et al., 2011, 2013), and mantle relaxation (e.g., Nooner et al., 2009) account for the remainder of the deformation. The cumulative opening during the first, largest dike intrusion is more than 50 times the plate opening rate of 15–20 mm/a estimated from GPS measurements (Vigny et al., 2006; McClusky et al., 2010). Thus the intense magma intrusion-extrusion cycles are primary contributors to the observed along-axis rift architecture, and the distribution of magma chamber(s) controls the along-axis segmentation.

4.2.2. Amagmatic Rifting Cycle

The 1989 Dobi earthquake swarm included 21 earthquakes of 4.5 < M <6.2 over a period of 2 days, but there was no evidence of magma involvement (Sigmundsson, 1992; Noir et al., 1997; Jacques et al., 1999, 2011) (Figs. 3–5). The first and largest earthquake event initiated along the western border fault, and it was followed by events that progressed in a counterclockwise fashion around the basin margins, with an overall NW propagation direction (Sigmundsson, 1992; Jacques et al., 2011). Several of the earthquakes produced slip along fan-shaped faults within the accommodation zone between the Dobi graben and the Hanle graben to the southeast (e.g., Tesfaye et al., 2008; Jacques et al., 2011) (Figs. 4, 5). Based on surface displacements mapped soon after the 1989 episode, and analyses of focal mechanisms, Jacques et al. (2011) estimate rupture lengths of ~15 km, and fault displacements of 0.35 m deduced from changes in topography in the absence of erosion (Fig. 5).

4.2.3. Comparison of Magmatic and Amagmatic Cycles

In both the magmatic and amagmatic rifting episodes, strain was distributed along the entire length of mature rift segments in highly extended crust transitional between oceanic and continental, providing insights into the maintenance of along-axis segmentation. The longest fault is the western border fault to the Dobi graben, with a length of 14 km and a maximum displacement of 620 m (Baker, 2008) (Fig. 5). In contrast, the longest fault in the Dabbahu segment is 7 km, and the largest displacement is ~100 m, excluding faulted volcanic constructs (Hayward and Ebinger, 1996; Rowland et al., 2007; Baker, 2008).

Field measurements of co-seismic slip show that the comparatively large magnitude (M_L ≤6.2), Dobi graben events were accompanied by ~0.35 m surface displacement along fault segments of a length of ≤15 km (Jacques et al., 2011), whereas the Dabbahu dike intrusion events were accompanied by 3–5 m vertical offsets and with an M_L <5.5 (Rowland et al., 2007; Ayele et al., 2007). Both showed normal slip along steeply dipping planes, although many of the dike-induced earthquakes had significant non–double couple solutions, interpreted as dilatation along normal faults above the dikes (Belachew et al., 2013). Comparing fault lengths and displacements with empirical magnitude-displacement relations, the 0.35 m slip is similar to that observed for other M_w 6.2 earthquakes worldwide (e.g., Wells and Coppersmith, 1994; Wesnousky, 2008). The 3–5 m slips for the dike-induced faulting in the Dabbahu magmatic segment (M_w <5.4) are clearly outliers, even considering that slip may have accrued over multiple earthquakes within the migrating swarms (e.g., Rowland et al., 2007; Ayele et al., 2009). Source mechanisms of the largest earthquakes (M_L <5) in subsequent large volume dike intrusions indicate that dikes induce rupture of short (~2 km) and large displacement (>0.5 m) faults above the dikes (Belachew et al., 2013). Work in progress with LiDAR data over faults that formed above dike intrusions promises to inform the length-displacement scaling relations.

A comparison of rift architecture in the two segments suggests that the 1989 and 2005 rifting episodes are characteristic of deformation styles in each basin (Fig. 5). That is, repeated magmatic rifting events characterized by local uplift and largely aseismic faulting in narrow zones above dikes sum to produce the low relief characteristic of magmatic segments, whereas repeated seismogenic slip along longer faults bounding and linking basins, and that rotate strata to dips >30°, produce the high relief rift segments. Decades of monitoring the inter-seismic cycle in the Asal rift indicate that faults slip at rates of ~4×10^{-3} m/a (Doubre and Peltzer, 2007).

Since the first geodetic measurements of magma intrusion in 2000, >75% of the seismic moment release in the Afar region has occurred via intrusion-induced earthquakes (Fig. 3). These include the May 2000 dike in the northern Main Ethiopian Rift (Keir et al., 2011), the October–November 2004 Dallol dike in northern Afar (Nobile et al., 2012), the multiple intrusions during the 2005–2010 Dabbahu rifting episode (see Section 3.2.1), the November 2010 western Gulf of Aden rifting episode (Shuler and Nettles, 2012; Ahmed et al., 2012), and the June 2011 eruption of Nabro volcano (Oppenheimer et al., 2011). Based on limited local seismicity patterns and repeat satellite imagery, these intrusions were all emplaced during time periods ranging from ~4 h to ~2 weeks. The total geodetic moment estimated using models of the InSAR data is 1.21×10^{20} Nm, an order of magnitude more than the 1.0×10^{19} Nm seismic moment release during the same time period (Fig. 4). This comparison demonstrates that catastrophic events, both magmatic and amagmatic, contribute significantly to long-term fault displacements, and maintain the along-axis rift architecture.

Although the repeat time between intense rifting cycles remains at the limits of surface dating methodology, historic records suggest that repeat periods are ~10^2 a. Two additional rifting episodes occurred prior to 2000: the 1978 Asal rifting episode (e.g., Abdallah et al., 1979) and the Alayta rifting episode during a 20-month period in 1906–1907 (Gouin, 1979). Additional rift opening events associated with earthquakes less than the M_w ≥3.5–4 threshold of the NEIC database include the November 2008 eruption of Alu-DallaFilla in northern Afar, accompanied by a maximum of ~4.6 m of rift opening near the center of the dike (Pagli et al., 2012). Thus, four of eight magmatic segments in the Red Sea and Gulf of Aden rift zones have undergone intense activity over the last century. The question remains whether the Dabbahu episode represents the maximum opening, or if discrete rifting episodes involve even larger magma volumes.

The only regional constraints on post-seismic processes are from the first years of the Dabbahu rifting episode (Nooner et al., 2009) and ~25 years of InSAR, GPS, and seismicity study in the Asal rift in Djibouti after a magma intrusion and faulting episode affected a 60-km-long rift segment in 1978 (Abdallah et al., 1979; Ruegg and Kasser, 1987; Ruegg et al., 1993; Cattin et al., 2005; Doubre and Peltzer, 2007). These studies quantify displacements on three major faults in the rift. Displacement rates varied from 4×10^{-3} m.y.$^{-1}$ during the interseismic cycle to 2.5×10^{-2} m.y.$^{-1}$ via ~month-long bursts of microseismicity at ~year-long intervals (Doubre and Peltzer, 2007). Time-averaged strain rate estimated from fault reconstructions of the elongate, rifted Fieale caldera is $1.7–2.4 \times 10^{-2}$ m.y.$^{-1}$ since ca. 150 ka, similar to the modern opening estimated from GPS networks (Vigny et al., 2006; McClusky et al., 2010). Summarizing, the intense magma intrusion events, combined with the viscous relaxation to the rapid opening, are several orders of magnitude more than the time-averaged opening rates estimated from seafloor-spreading anomalies, and regional GPS networks and fault slips are much greater than predicted by empirical earthquake scaling relations.

4.3. Tectono-Magmatic Rifting Cycles: Juvenile to Middle-Aged Rift Zones

South of the Ethiopian Plateau and the Eocene-Oligocene flood basalt province, faulting and magmatism have localized in Proterozoic-Precambrian orogenic belts around the deeply rooted Tanzania craton (Figs. 3, 6). These rift systems are segmented along their length into 60- to 120-km-long border fault systems, and the rift architecture shows large variations, particularly in the degree of magma involvement. Here we contrast constraints on short-term deformation processes in the relatively magma-rich Eastern rift with those of the amagmatic Tanganyika rift in the Western rift (Fig. 6). No geodetic profiles span these rift zones, and rift opening rates are estimated at ~3 mm/a using rigid block models of sparse continuous GPS data (Stamps et al., 2008).

4.3.1. Magmatic Rifting Cycle

Nearly all of the seismic energy release in the admittedly short time period of the NEIC catalogue is associated with a 3-month-long sequence of faulting, intrusion, and subsequent eruption of the carbonatitic volcano, Oldoinyo Lengai, in northeastern Tanzania in 2007 (e.g., Baer et al., 2008; Calais et al., 2008; Albaric et al., 2010; Vaughan et al., 2008). One month prior to the first earthquake in the swarm (15 July 2007, M_w 5.4), satellite imagery detected thermal anomalies and increased eruptive activity in the Oldoinyo Lengai crater (Vaughan et al., 2008) (Fig. 6). The largest earthquake in the swarm was the 17 July 2007 M_w 5.9 earthquake beneath the Natron basin, in the same location, within standard errors. Between 12 and 17 July, swarms of earthquakes migrated upward and 10–15 km northeastward (Albaric et al., 2010), during a period of slow fault slip detected in GPS data (Calais et al., 2008). Waveform modeling of the earthquake swarms indicates slip along NE-striking faults, but the slip planes remain debated (e.g., Baer et al., 2008; Calais et al., 2008; Biggs et al., 2009). A second InSAR scene captured uplift, indicating a dike intrusion along the southern flank of the Gelai shield volcano. Calais et al. (2008) modeled the intrusion as a 2-m-wide dike intrusion beneath the ~8-km-long zone of open fissures and normal faults. Nearly 55 days later, an M_w 5.4 earthquake and an M_w 5.5 earthquake with source depths <9 km occurred beneath a nearby volcano, Oldoinyo Lengai, which then entered into an explosive eruptive stage that continued another nine months (e.g., Vaughan et al., 2008; Albaric et al., 2010). This example demonstrates both the complex interplay between magma and faulting in the accommodation of rift deformation and the stress coupling between pressurized magma chambers and fault zone unloading and loading.

4.3.2. Amagmatic Rifting Cycle

Using the admittedly short NEIC record between 1973 and 2012, the estimated seismic moment release for the Tanganyika rift is 3.2×10^{19} Nm, the same order of magnitude as moment release in an equal area that encloses the Afar triple junction, where extensional velocity is 4 times that estimated for the Tanganyika rift (Figs. 4, 6). This window spans ~7 rift segments, increasing the probability that at least one segment has undergone a rifting episode characteristic of that rift sector. Seismic moment release is dominated by two relatively large earthquakes and their aftershocks: in 2 October 2000 the M_w 6.5 earthquake at a depth of 39 km beneath the central Tanganyika rift, and the 5 December 2005 the M_w 6.8 earthquake at a depth of 16 km beneath the southern basin (Craig et al., 2011). Aftershocks of the 5 December 2005 mainshock had source depths as deep as 27 km (Craig et al., 2011), consistent with long, steep border fault systems that penetrate to lower crustal levels in relatively thick, strong lithosphere. The deep extent of border faults indicates that these basins have achieved their maximum length. Like most of the Western rift earthquakes, there is no surface deformation associated with sub-lacustrine earthquakes, and they are not in the global-fault-length versus displacement comparisons.

4.3.3. Comparison of Magmatic and Amagmatic Rifting Cycles

Given the very slow strain rates of the Eastern and Western rift zones (≤3 mm/a), the rifting cycle comparisons are clearly biased by the short time scale of space-based geodetic and seismic observations, which span a small fraction of the rifting cycle. As was seen in the Afar comparisons, the magmatically active Eastern rift episodes demonstrate that seismic energy release, a measure of brittle failure, accounts for a small fraction of the deformation; most of the strain during discrete, rifting episodes in magmatically active rift segments is accommodated by magma intrusion and slow-slip. Unfortunately, too little information on

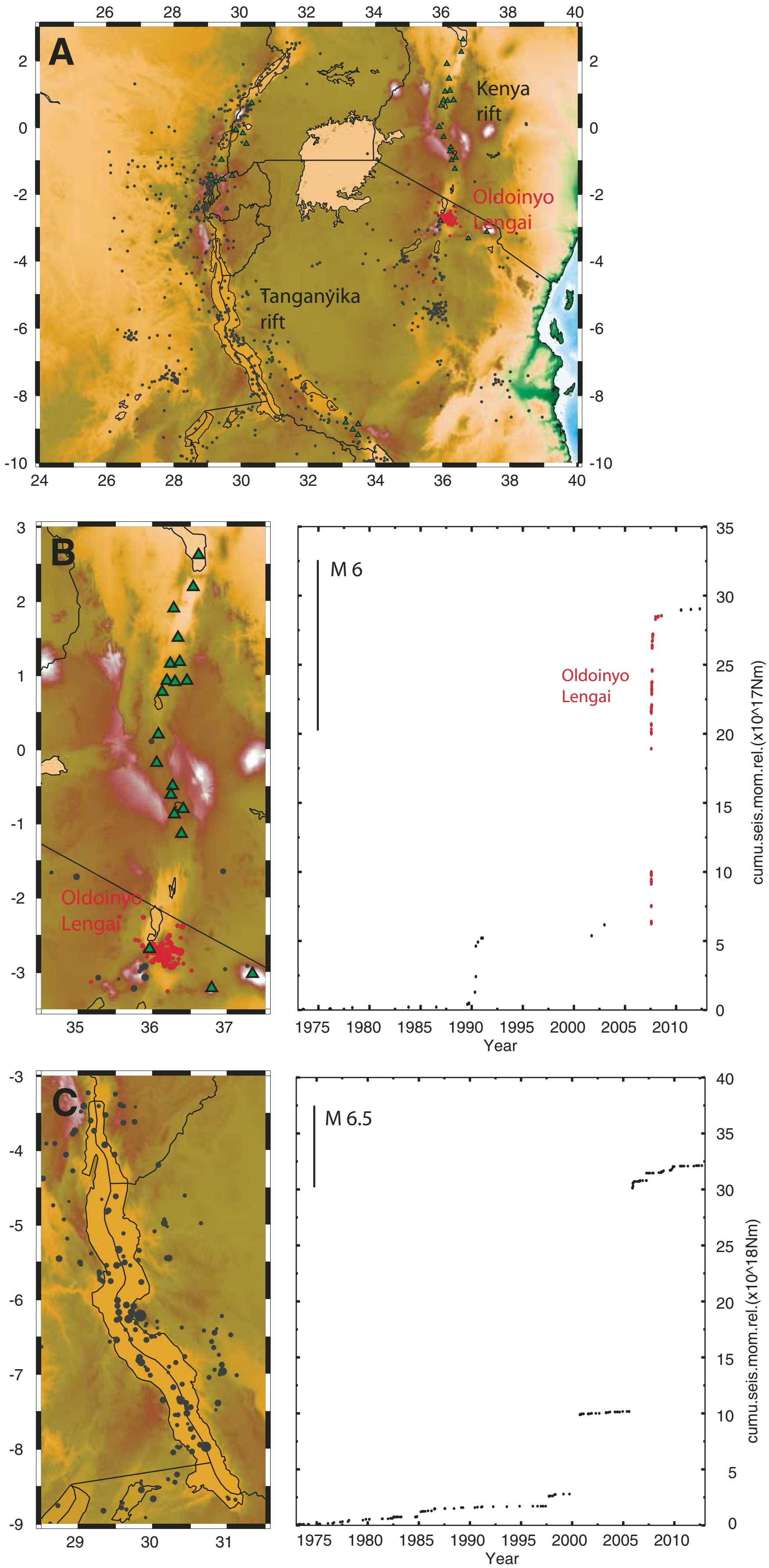


Figure 6. Comparison of seismic moment release in magmatic and amagmatic rift sectors. Note that vertical scaling differs by one order of magnitude for 6B and 6C. (A) Distribution of seismicity in East Africa reported by the NEIC since 1973, and distribution of Holocene volcanoes as green triangles. Red dots are earthquakes associated with magma intrusion episodes, whereas black dots are those with no associated magmatic activity. Note that the Eastern rift is relatively aseismic but has numerous volcanoes within the central rift valley. In contrast, the Tanganyika rift is seismically active and magma-poor. (B, C) Maps of NEIC seismicity and Holocene volcanoes in the Tanganyika and Kenya rifts, with the release of the seismic energy through time displayed on the rift. Seismic moment release is dominated by two relatively large earthquakes and their aftershocks (Craig et al., 2011). The total seismic moment release for the Eastern rift in Kenya and northern Tanzania is an order of magnitude less at 2.9 × 10^{18} Nm. The Eastern rift seismic moment release is dominated by a 2-month-long sequence in 2007 of faulting, intrusion, and subsequent eruption of the carbonatitic volcano Oldoinyo Lengai in northeastern Tanzania (e.g., Calais et al., 2008; Albaric et al., 2010).

the sub-lacustrine Tanganyika earthquakes is available to compare with the empirical fault-length displacement ratios. Independent data from the Malawi rift zone show that entire 100-km-long fault segments can rupture in single or coupled earthquake events (e.g., Jackson and Blenkinsop, 1997).

5. EPISODIC RIFTING ON THE MILLIONS OF YEARS TIME SCALE: OBSERVATIONS FROM THE RIO GRANDE RIFT

The Rio Grande rift zone extends over a distance of >1000 km from Leadville, Colorado, to Chihuahua, Mexico (Fig. 7). The northern part of the rift separates the Colorado Plateau on the west from the Great Plains on the east. It consists of a series of narrow rift basins trending approximately N-S. These basins resulted from a recent late Miocene–Holocene extensional phase (Morgan et al., 1986). An early event of crustal extension occurred during middle Oligocene–early Miocene time. Basins formed during this extensional phase were much broader than the younger basins (Baldridge et al., 1994), and they trended northwestward in contrast to the young N-S basins. The exact age of initiation of extension for the Rio Grande rift is debated. If the oldest erupted basalts and silicic volcanic rocks are taken as an indication for the onset of extension, then extension is estimated to have started in the Oligocene (Ingersoll, 2001). The earliest rift-related sedimentary rocks have younger ages, however, ranging from late Oligocene to early Miocene (Keller and Cather, 1994; Ingersoll, 2001).

The Rio Grande rift has not been widening significantly in historic times. GPS measurements suggest a current opening rate of ~1 mm/a (Savage et al., 1980; Berglund et al., 2012). Large Holocene earthquakes have been documented to occur on several of its border faults, however, including the Pajarito fault of the Española Basin. Here, a Holocene earthquake occurred as recently as ca. 1.5 ka, and many other large Holocene earthquakes have been documented along the rift (McCalpin, 2005; Section 6). Such observations from currently active rifts point toward an episodic pattern of continental rifting: Rifts are actively widening and lengthening during short (magmatic and/or fault controlled) events, followed by a phase of relative quiescence.

Rifting in the Rio Grande rift region has been episodic on the millions-of-years time scale, with separate rifting events from the late Cenozoic to the Quaternary. Rifting has been described as being multi-phased (Morgan et al., 1986; Ingersoll, 2001; Smith et al., 2002); a first period of extension occurred from ca. 27–20 Ma, and a second period occurred beginning at 10 Ma until recent times. Magmatic activity reflects the episodic rift behavior; the time in between the rapid rifting phases was characterized by minimal magmatism (Morgan et al., 1986; Baldridge et al., 1991). A compilation of igneous rock ages from New Mexico

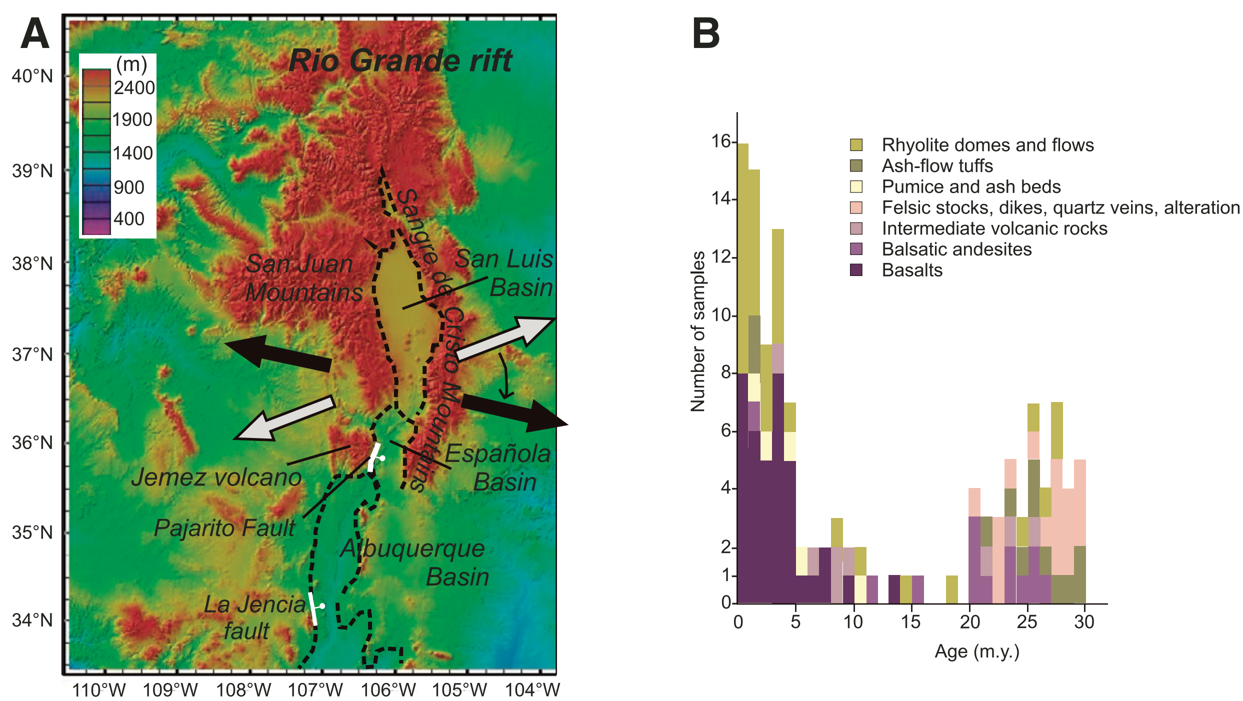


Figure 7. (A) N-S–oriented Neogene basins of the northern Rio Grande rift (outlined by dashed lines), Pajarito and La Jencia normal faults (see text for discussion), indicated by white lines, and least principle stress directions from Zoback et al. (1981). Paleo-stress field indicated by white arrows, recent stress field indicated by black arrows. (B) Histogram of igneous rock ages from Oligocene present in New Mexico from Chapin and Seager (1975). The period of tectonic quiescence in the Rio Grande rift (estimated between ca. 20 and 10 Ma) coincides with a period of lesser igneous activity in New Mexico. Ages based on K/Ar and fission track dates.

(Fig. 7; Chapin and Seager, 1975) shows that the Oligocene–early Miocene phase of magmatism was followed by a period of relative quiescence between ca. 20 and 10 Ma. Then magmatic activity increased again ca. 10 Ma, and the rift has been magmatically active since. The composition of igneous rocks was different during the earlier and most recent rift phases (Perry et al., 1988), possibly reflecting a change in depth of melting and thinning of the lithosphere, as well as metasomatism (Perkins et al., 2006). Oligocene–early Miocene rifting was characterized by basaltic andesites and by intermediate and felsic compositions; the most recent rift phase had a large basaltic component (Fig. 7), indicating asthenospheric melting and thermal erosion of the mantle lithosphere (Perry et al., 1988).

Paleostress data are available that give information on the orientation of past stress fields in the southern Rio Grande rift (e.g., Zoback et al., 1981). These data indicate a ~45° clockwise rotation of the extension direction since ca. 10 Ma in the southern part of the rift (Zoback et al., 1981). WSW-ENE extension, resulting in northwestward-trending basins, occurred between about the middle Oligocene to the early Miocene (ca. 30–20 Ma) in the Rio Grande rift region (Fig. 8) (Baldridge et al., 1991, 1994). These basins have been overprinted by the most recent rifting phase. During this most recent active phase of extension (beginning ca. 10 Ma), extension had rotated to about E-W. This extension direction was orthogonal to the N-S–trending Rio Grande rift basins, which form the characteristic narrow segments of the rift. This least-principal stress orientation has persisted to Holocene times.

The magmatically active phases of extension in the Rio Grande rift appear coincident with changes in the lithospheric stress field. The stable E-W, least-principal stress orientation since the late Miocene coincides with the most recent rift phase during which the N-S–trending narrow basins of the rift were formed. During the earlier stable WSW-ENE extension direction, the NW-trending basins developed. The rift was in a relatively inactive phase with little associated magmatism during rotation of the lithospheric stress field before ca. 10 Ma (Figs. 7, 8).

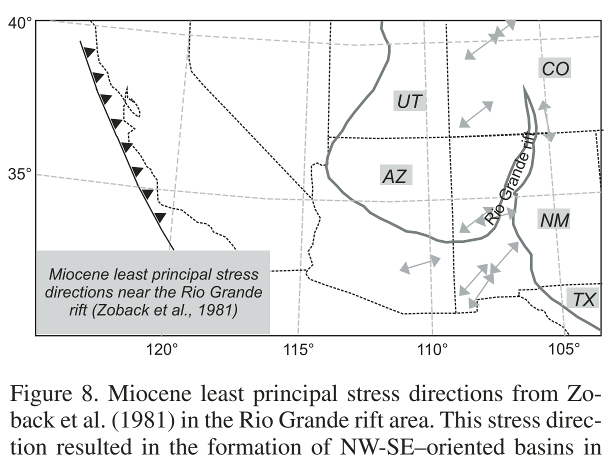


Figure 8. Miocene least principal stress directions from Zoback et al. (1981) in the Rio Grande rift area. This stress direction resulted in the formation of NW-SE–oriented basins in the Rio Grande rift, which were more recently overprinted by the current narrow basins. Dark-gray line outlines the Colorado Plateau and Rio Grande rift tectonic provinces. Triangular line indicates subduction-zone plate boundary. AZ—Arizona; CO—Colorado; NM—New Mexico; UT—Utah; TX—Texas.

Changes in the regional stress field in the western USA on these time scales are related to changes in plate boundary forces (e.g., Zoback et al., 1981). Late Cretaceous–early Tertiary compression related to Farallon plate subduction was followed by extensional tectonics. The exact onset of extension is debated, but a broad zone of backarc-related extension with a direction perpendicular to the subduction trench developed in the western USA between ca. 30 and 20 Ma. The least principal stress direction at this time became generally WSW-ENE and was responsible for formation of the NW-trending Rio Grande rift basins. With the development of the San Andreas Fault system the least principal stress orientation changed to E-W, and this rotation was responsible for formation of the NS-oriented narrow segments of the rift.

The Rio Grande rift is not unique in its faceted rifting history. Many continental rifts (and rifted margins) have undergone periods of active rifting, alternating with phases of less activity over time scales of millions to tens of millions of years. On the eastern margins of the North Atlantic (Norwegian margin), multiple rifting episodes were recorded before rupture occurred (e.g., Lundin and Doré, 1997; Reemst and Cloetingh, 2000). Here, they resulted in a series of rift basins formed between the Permo-Triassic and the breakup at the Paleocene–Eocene transition. The orientation of the basin axes did not change significantly between rifting episodes, which suggests that the orientation of the stress field must have been constant. In contrast to the Rio Grande rift, not all of the basins on the Norwegian margin overlap (Reemst and Cloetingh, 2000). The episodic rifting on the Norwegian margins has been explained by changes in plate boundary forces and in the lithospheric stress field (e.g., Bukovics and Ziegler, 1985; Reemst and Cloetingh, 2000), and by extension in a constant stress field with a migration in the locus of rifting (van Wijk and Cloetingh, 2002).

Also in the Baikal rift the direction of the paleostress field has changed during rifting. Rifting in the Baikal rift started ca. 30 Ma, resulting in a series of narrow and elongated basins in the central and southwestern parts of the rift and narrow but shorter basins in the north (Delvaux et al., 1997; van der Beek, 1997; Corti et al., 2011). The orientation of the stress field in the late Oligocene–Miocene that resulted in the onset of rifting was generally transtensional, with localized areas of transpression (Delvaux et al., 1997). This orientation changed in the late Miocene–early Pliocene during a ~5 m.y. transitional phase. By 3 Ma, when the stress field had rotated to facilitate pure extension in the rift, the recent active rifting phase started (Delvaux et al., 1997). Pure extension resulted in a change in fault kinematics in the central part of the Baikal rift, while the southern part underwent strike-slip motion. The change in stress orientation has been attributed to a change in the intraplate stress field related to the India-Eurasia collision zone and subduction in the western Pacific (Corti et al., 2011), although some rotation may have been related to evolving basin linkage zones. Stress field rotations also may have affected parts of the East African Rift System as rifting propagated along-axis, and Africa moved relative to the mantle upwelling (e.g., Strecker et al., 1990; Mortimer et al., 2007).

To summarize, on the long time scale of 10^6–10^7 a, rift activity shows a strong relationship to the intraplate stress field resulting from plate boundary forces (Fig. 8). Rifts open during phases when the stress field is stable, and are in a state of quiescence characterized by very low interseismic strains (order of ~1–5 mm/a) and limited tectonic-magmatic activity when the stress field rotates.

6. PALEO-SEISMICITY INDICATIONS FOR TECTONIC EXTENSION PATTERNS ON THE 100,000 YEARS TIME SCALE

Current seismic activity along the Rio Grande rift zone is low, with the exception of the Socorro magma body (Fig. 9; Sanford et al., 2002). In the past 40 years, earthquakes have been small (M <4.5) and shallow (upper crust), and most were associated with the Socorro magma body (Fig. 8; Sanford et al., 2002). Also, historically, earthquakes have been sparse. In the last ~140 years or so (the recorded history of New Mexico), few magnitude 5 earthquakes have been reported in the Rio Grande rift. They had an estimated maximum intensity of VI–VIII (modified Mercalli scale) and occurred near the Socorro magma intrusion (Sanford et al., 1995). The Rio Grande rift is not discernible by seismicity (Fig. 8); instead, earthquake trends tend to cross the rift zone in several places. The Jemez Lineament and the Socorro Fracture Zone are SW-NE–trending linear zones of comparatively high seismicity levels. These zones correspond to Proterozoic plate boundaries and lineaments and are not directly related to Miocene rifting (Karlstrom and Humphreys, 1998). The present-day extensional strain rate across the Rio Grande rift is extremely small (Berglund et al., 2012) and is distributed over a wide region from the Great Plains to the Colorado Plateau. This lack of localized deformation, combined with the limited recent earthquake activity, suggests that the rift is currently undergoing a period of tectonic quiescence.

The rift has not always been seismically inactive during the Holocene and Pleistocene. Detailed studies of two large border faults of the rift, the Pajarito fault of the Española basin (Fig. 7) and the La Jencia fault farther south in the central part of the Rio Grande rift, show a series of fairly recent movement along most of the segments of both faults (Machette, 1986; McCalpin, 2005). On the La Jencia fault, five to six events have been found for the last 33 ka (Machette, 1986). Some of them may have been as strong as M_L 6–7. Prior to this latest Pleistocene phase of activity, however, a phase of tectonic quiescence occurred during a >0.5 m.y. period. On the Pajarito fault zone, recurrence intervals are longer and variable, probably between 10 and ca. 60 ka (McCalpin, 2005). Also on this fault zone Holocene faulting events have been found, probably between ca. 1.4 and 11 ka (McCalpin, 2005). Slip rates on the faults have been highly variable. On the La Jencia fault, slip rates have varied from the late Miocene to the early Pliocene, when they were relatively large (~100 m/m.y.) to the middle and late Quaternary, when they were much slower (~5–20 m/m.y.; Machette and McGimsey, 1983). Slip rates on the Pajarito fault during the last ca. 100 ka were estimated to range between ca. 0.004 and 0.10 m/ka; various seismic cycles yielded different slip rates (McCalpin, 2005). Other faults in the rift show similar patterns, suggesting that the recurrence interval of some of the faults may be of the order of 100,000 years or so (Machette, 1986). Many Quaternary faults in the Rio Grande rift, however, have a Holocene recurrence interval of several to many hundreds of years (Machette et al., 1998), which may explain why no large earthquakes have been detected historically.

To summarize, although the Rio Grande rift is currently inactive, and has historically not undergone any large earthquakes, it underwent several more seismically active rifting periods during the Holocene and Pleistocene. The earthquake recurrence interval and slip rates vary strongly over time, but there are indications of a Pleistocene period of tectonic quiescence on the La Jencia fault in the central Rio Grande rift. The recurrence interval is similar to that found on some other faults in the western USA, including the Owens Valley Fault (Bacon and Pezzopane, 2007). This high degree of variability in earthquake recurrence is common on the time scale of the recurrence interval (<100,000 years) (Wallace, 1987; Coppersmith, 1989; Sieh et al., 1989; Machette et al., 1991). Nicol et al. (2005) found that these short-term fluctuations in recurrence frequency are not related to the far-field stresses but to fault interactions. In contrast, the recurrence intervals on geological time scales (100,000–1 m.y.) are related to the long-term tectonic loading rate (Nicol et al., 2005). The observed ~0.5-m.y. period of relative quiescence on the La Jencia fault may thus indicate an adjustment in regional strain rates.

7. RIFTING PERIODICITY: FROM SECONDS TO GEOLOGICAL TIME SCALES

Although rifting of continental lithosphere is commonly described as a continuous process, this overview of magmatic and amagmatic rifting cycles summarizes evidence that rift opening is discontinuous, on both geological and daily to millennial time scales (Fig. 10). Owing to the short record of seismic and geodetic monitoring relative to the rifting cycle, we can only generalize in terms of opening rates during discrete events versus the inter-seismic plate motions. At the longest time scales, variations in the direction of the least principal stress direction, which forms and loads faults, result in slow opening with minimal magmatism, as is seen in the Rio Grande rift. During periods of stable plate configurations, over time periods of millions of years, rates match well with extrapolations of modern plate opening (e.g., Vigny et al., 2006; McClusky et al., 2010; Bennett et al., 2003). Observations from the Baikal rift and Rio Grande rift suggest that episodes of stress field rotation on geologic time scales appear to coincide with tectonic quiescence of continental rifts.

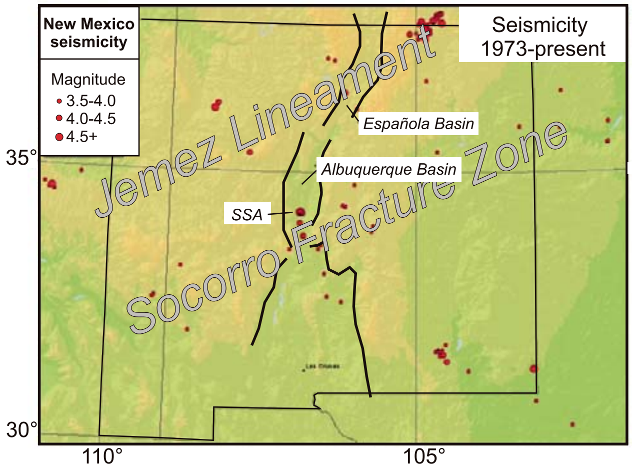


Figure 9. Seismicity in the Rio Grande rift and surrounding areas (source: USGS, http://earthquake.usgs.gov/earthquakes/states). Sparse seismic activity in the rift primarily is located around the Socorro Seismic Anomaly (SSA), and is induced by emplacement of the Socorro magma body. On longer time scales interpreted from trenching, two seismic trends crossing the Rio Grande rift can be recognized: the Jemez Lineament and the Socorro Fracture Zone. The rift is currently seismically and tectonically inactive.

At the time scales of the seismic and magmatic cycles, the East African Rift examples show variations related to stage of rifting, as well as the presence or absence of magma at crustal levels. These processes are inherently 3D, given that earthquakes rupture finite-length fault segments, and dike intrusions reach distances controlled by excess pressure within magma chambers. During the initial and early stages of rifting, interseismic plate opening is very slow, and at the detection limits of satellite geodesy. The plate is thick and comparatively strong, and the stress accumulation required to cause fault slip may take 10^3–10^5 a, based on sparse paleo-trenching studies (e.g., Zielke and Strecker, 2009; Machette et al., 1998). All or large parts of long border-fault systems that develop very early in basin evolution slip in large magnitude earthquakes (e.g., Kinabo et al., 2008; Craig et al., 2011), and intrabasinal faulting may also be seismogenic (e.g., Biggs et al., 2011). The interseismic cycle is barely detectable with GPS, suggesting that nearly all strain occurs during rifting episodes. Although magma intrusion and storage in the mantle lithosphere

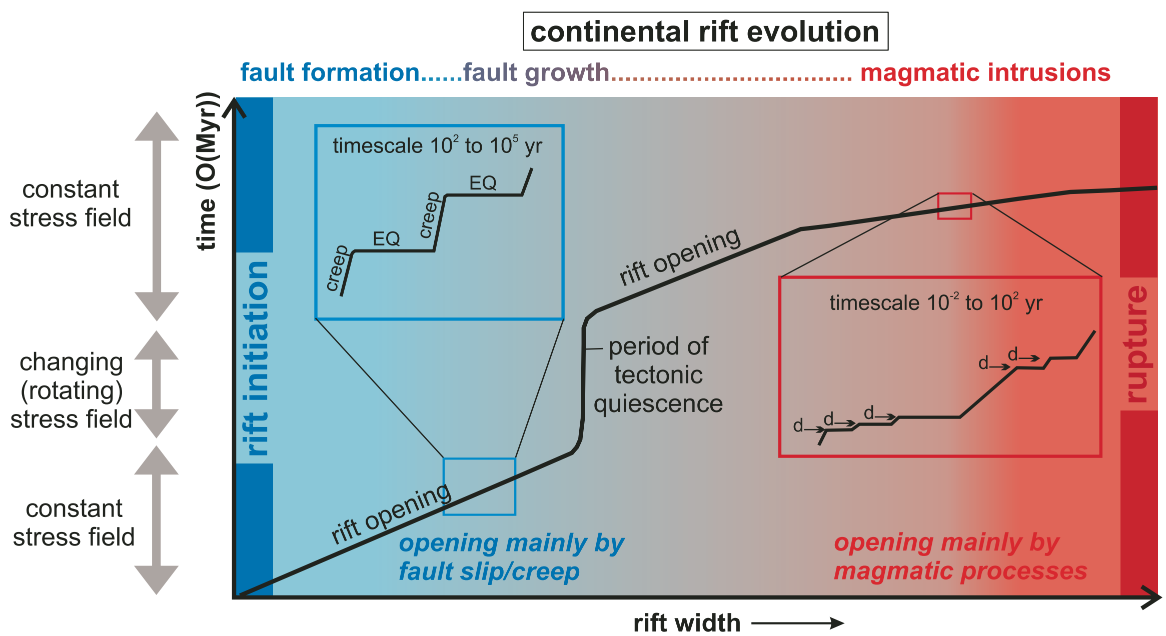


Figure 10. Schematic evolution of continental rifts, showing a time range of processes, as historically observed. During periods of constant far-field extensional stress the rift opens; periods of stress-field rotation coincide with tectonic quiescence. During the first stages of rifting, rift opening occurs primarily through slip and creep of faults as a result of tectonic loading. At levels deeper than the seismogenic layer, stretching is accommodated by ductile deformation. Rift opening accelerates when magma reaches crustal levels. Magmatic processes dominate (strain, plate weakening) toward rupture. Rift opening by dike injections is stepwise and much faster than opening by faulting alone. EQ—earthquake; d—dike intrusion.

may also be rate-controlling factors, we have few constraints on the time and length scales of these processes in early stage rifts. Volatile degassing from metasomatized mantle lithosphere may also contribute to plate weakening during early stage rifting (e.g., Vauchez et al., 2005; Fischer et al., 2009).

As magma rises to shallower levels, crustal magma chambers develop within the central rift basin. Strain accommodated by magma intrusion increases, and seismic energy decreases, as seen in the comparison of rifting episodes (e.g., Figs. 4, 6). Although the historical record is short, the number of volcanoes showing magma replenishment over the past two decades (Biggs et al., 2009, 2011) and the Natron rifting event suggest that strain accommodation by dike intrusion reduces the repeat time between rifting episodes. As a consequence, the interval between border fault slip episodes may increase, as strain localizes to the zones of magma intrusion.

As rifting progresses to seafloor spreading, the rate of opening is controlled by the magma replenishment–eruption cycle (e.g., Figs. 1, 4). These intense periods of magma intrusion and faulting above the dikes occur over time periods of hours to days, and at opening rates several orders of magnitude faster than the time-averaged ones. For decades after the large-volume dike intrusions, opening occurs at rates much faster than the secular plate opening rates, as the visco-elastic mantle relaxes in response to the change in stress (e.g., Cattin et al., 2005; Nooner et al., 2009). These examples demonstrate that the inter-seismic period in rift zones with crustal magma reservoirs is strongly dependent upon the magma replenishment cycle, controlling the state of stress within the reservoir.

Taking these cycles back to the geological process, when magma is abundant, extension by the dike intrusion will be effective in relieving far-field extensional stress, and stress will rarely accumulate to levels high enough to cause slip along border or intrabasinal faults. Since the magmatic intrusions locally transfer heat and rheologically alter the surrounding crust and sediments, fault and magmatic localization to weakened sites of earlier intrusions are expected, accelerating opening rates (Bialas et al., 2010). Further, at this late stage of rifting the regional stress field may change as a result of thermal or mechanical erosion of the mantle lithosphere, which may result in uplift (e.g., Rychert et al., 2012; Esedo et al., 2012), increasing tensional forces that facilitate rupture, or by buoyant uplift as a result of asthenospheric convection below the rift (Buck, 1986; Esedo et al., 2012). On the geological time scale, it is this change in tensional forces, combined with crustal alteration from magma intrusions, that facilitates continental breakup (Fig. 10).

8. IMPLICATIONS AND FUTURE DIRECTIONS

The recognition that entire 50–100-km-long rift segments at all stages of development undergo detectable deformation along their lengths during both magmatic and amagmatic rifting events motivates a reevaluation of time-averaged rates of rifting processes. The magmatic events in particular accommodate centuries of inter-seismic strain, implying that inter-seismic plate opening rates in late stage rifts should be extrapolated to the past with caution. A major obstacle to understanding long-term patterns of fault growth and magma intrusion is sampling and dating the sedimentary and volcanic rocks these processes disrupt. Paleotrenching and seismic profiling of lake basins, combined with dating of long core samples, offer an ideal opportunity to decipher the repeat time periods of major earthquakes. In regions where subsurface sampling is costly or complicated, improved measurements of surface fault displacements using field measurements and high resolution digital elevation models (DEMs), supplemented by comprehensive dating of fault-cut flows, and cosmonuclide analysis on exposed fault scarps, are required. Comprehensive mapping and dating programs of volcanic complexes are required to understand temporal variations in eruption volumes and styles.

Our understanding of decadal scale variations in faulting is primarily derived from the record of large, poorly located earthquakes in catalogues from global seismic networks. While these data provide a good first-order record of co-seismic deformation, local seismicity and geodetic studies are required to quantify the strain accommodated largely aseismically by magma intrusion, as well as the time scales and rates of post-seismic deformation in cratonic lithosphere. Data from local seismic and geodetic networks will also facilitate accurate imaging of the variation in strain release on active fault planes where depth constraints are crucial for understanding rheological variability in the crust and transition from brittle faulting to ductile creep. Recent and ongoing multidisciplinary projects are beginning to provide three-dimensional constraints on the distribution of strain and magmatism in the lower crust within active rifts needed to develop the next generation of rift models (e.g., Keranen et al., 2004; Thybo and Nielsen, 2009; Hammond et al., 2011; Delph et al., 2011). Systematic measurements of volatile degassing in rift zones will help constrain their flux through stretching lithosphere, and inform models and experiments of mantle lithospheric deformation where partial pressures of CO_2 are elevated (e.g., Fischer et al., 2009; Head et al., 2011).

9. CONCLUSIONS

Over time periods of millions of years, periods of rotating regional stress fields are marked by a lull in magmatic activity and a temporary halt to tectonic rift opening. Over time periods of decades to millennia, the presence or absence of magma at crustal levels produces very different strain patterns. Within rift sectors with upper crustal magma chambers beneath the central rift valley (Eastern, Main Ethiopian, Red Sea, and Gulf of Aden rifts), seismic energy release accounts for a small fraction of the deformation. In these mature rift sectors, most of the strain is accommodated by magma intrusion and slow-slip. Magma intrusion processes appear to decrease the time period between rifting episodes, effectively accelerating the rift to rupture process. Thus, the inter-seismic period in rift zones with crustal magma reservoirs is strongly dependent upon the magma replenishment cycle. Both magmatic and amagmatic rifting events can occur within the same rift sector, but the time-averaged rift morphology indicates that the magmatic and amagmatic cycles are representative of strain patterns in each rift segment. Thus, both magmatic and amagmatic rifting events produce the long-term fault displacements and maintain the along-axis rift architecture through repeated episodes of faulting, magma intrusion, and post-seismic–intrusion relaxation in the mantle lithosphere. Given the short time scale of seismic and geodetic measurements relative to the 10^2–10^5 a rifting cycle, the rates from the historical examples may not span the full range of rift activity, but they motivate reevaluation of time-averaged rates of rifting processes determined from basin subsidence patterns and inter-seismic geodetic measurements, and geohazard mitigation programs.

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