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How diking affects the tectonomagmatic evolution of slow spreading plate boundaries: Overview and model

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ABSTRACT

Recent diking episodes along slow spreading boundaries included the generation of normal faults, showing that extension is accommodated, on a scale of a few years or less, by both magma intrusion and fault movement. Here we aim to define how diking may affect the overall rift structure on the longer term (≥100 yr). We first summarize the main features of the transient diking episodes as obtained from geological, geophysical, geodetic, and modeling studies. We then put these episodes into a broader context, considering the overall longer term shallow and deep structure of the plate boundaries. The synthesis of the data shows that in Iceland crustal extension at depth largely occurs by means of dikes, with negligible normal faulting; faults focus toward the surface (<1 km depth), forming dike-induced grabens commonly propagating downward; along-rift diking episodes (1 in ~200 yr) may induce all the observed surface deformation. The close similarities with transitional (Afar Rift) and magmatic continental rifts (East African Rift System) suggest that repeated diking induces most of the surface deformation along slowly spreading (≤2 cm/yr) magmatic plate boundaries. The frequency of diking and the induced strain may not allow extension to be accommodated amagmatically, through creep or seismic or aseismic faulting. This implies that a diking episode locks any amagmatic faulting until the strain is released (centuries), when the subsequent diking episode occurs, with the cumulative result of controlling and shaping the evolution of slow spreading magmatic plate boundaries. This process appears independent of the stage of magmatic rifting.

INTRODUCTION

Slow spreading magmatic plate boundaries (herein referred to as boundaries) are characterized by the drifting motion of the two interacting plates from each other and accompanied by active volcanism at the surface and magma emplacement at depth. In the past decades, several diking episodes (multiple intrusions lasting a few years) or events (single intrusion lasting a few weeks) have occurred along the continental, transitional, and oceanic portions of these boundaries: these include Krafla, Iceland (1975–1984), Asal-Ghoubbet, Afar (1978–1979), Dallol, Afar (2004), Dabbahu, Afar (2005–2009), Lake Natron, Tanzania (2007), Harrat Lunayyir, Red Sea (2009), and Bardarbunga, Iceland (2014) (i.e., Jacques et al., 1996; Ruegg et al., 1979; Tarantola et al., 1980; Buck et al., 2006; Wright et al., 2006; Calais et al., 2008; Biggs et al., 2009; Pallister et al., 2010; Nobile et al., 2012; Gudmundsson et al., 2014; Sigmundsson et al., 2015). In these cases, diking occurred along magmatic systems, consisting of rift-parallel zones of focused magmatic and tectonic activity along the boundary (Gudmundsson, 1995a; Ebinger and Casey, 2001; Acocella, 2014). These diking episodes usually generated (or reactivated) normal faults at the surface, forming graben structures a few hundreds of meters to a few kilometers wide, with slip along the faults between a few tens of centimeters and ~3 m (Sigurdsson, 1980; Rubin and Pollard, 1988; Rowland et al., 2007; Ruch et al., 2015; Trippanera et al., 2015a). The formation, or reactivation, of these normal faults shows how a part of the rift structure observed at the surface is dike induced, at least on the short term (<10 yr) (e.g., Buck, 2006; Ebinger et al., 2010; Acocella, 2014). Accordingly, the evolution of these boundaries appears largely episodic and magmatic, at least for the best-known Icelandic case; this case shows an intrusive frequency of one every a few hundreds of years that is expected to induce repeated faulting at the surface (Fig. 1; Sigmundsson, 2006; Ebinger et al., 2010).

Less is known about the longer term evolution (≥100 yr) of these boundaries and, in particular, if the short-term events may entirely account for the cumulative deformation along rift zones, or if other processes (as amagmatic faulting or creep) should be considered (Fig. 1). Available data on the development of major graben structures within the magmatic systems suggest periods of activity of ~100 ka or slightly less, as at Fantale (Main Ethiopian Rift, MER; Williams et al., 2004), Dabbahu (Afar; Rowland et al., 2007; Medynski et al., 2013), and Thingvellir (Bull et al., 2005): for example, several magmatic cycles, each lasting 20–40 ka and resulting in periods of high and low magma supply rate, have been distinguished at Dabbahu (Medynski et al., 2013). The cumulative surface deformation developed at these representative locations in these time spans consists of graben-like structures usually <4–5 km wide, with conjugate sets of normal faults, each with vertical displacement of ~10 m (Gudmundsson, 1987; Rowland et al., 2007; Trippanera et al., 2015a). Considering a frequency of diking episodes of ~100 yr (at least for the best-known Icelandic case), with slip values along normal faults of a few meters over periods of 10,000 yr, gives a theoretical cumulative fault displacement of 100 m, quite in agreement with the observed fault scarps, several tens of meters high, along the boundaries; any discrepancy between the theoretical and observed values may be related

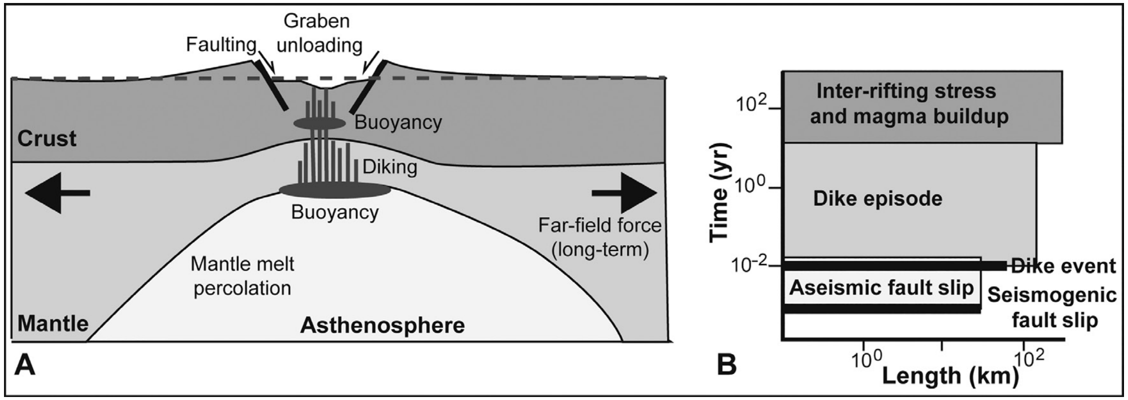

Figure 1. (A) Schematic section view of a divergent plate boundary, and the main mechanisms possibly controlling rifting. (B) Summary of length and time scales of extensional strain processes (modified after Ebinger et al., 2010).

to erosion (on the fault footwall) or deposition (on the hanging wall). This suggests that, in principle, all the surface deformation along the boundaries may be dike induced. However, regional extensional tectonics, in the form of creep or amagmatic faulting, may also play a longer term role between the diking episodes occurring on the shorter term (Fig. 1B). Unfortunately, very limited, mostly geodetic, information is available to test this possible contribution of amagmatic faulting or fault creep (e.g., Gudmundsson, 2013). In several cases, as at the Asal or Krafla magmatic segments, the detected geodetic non-cointrusive extension, slightly higher than the expected one from global plate motion models, is related to the decade-long postrifting relaxation of the crust (Hofton and Foulger, 1996; Doubre and Peltzer, 2007; Vigny et al., 2007; Arnadottir et al., 2009). In other cases, as along the magmatically inactive Reykjanes Peninsula, Iceland, the amount of extension varies from 0 (between 1993 and 1998) to ~7 mm/yr (between 2000 and 2006) (Hreinsdottir et al., 2001; Keiding et al., 2008). Similar variations have been obtained for the MER, where geodetic measurements between 1976 and 1992 suggest a widening rate of ~1.1 ± 2.2 mm/yr (Asfaw et al., 1992), an extension of 4.5 ± 1 mm/yr between 1969 and 1997 (Bilham et al., 1999), and an extension of 8 mm/yr between 1992 and 2010 (Kogan et al., 2012). Compared to the 1976–1992 measurements, the higher 1969–1997 extension rate may have been affected by the postemplacement relaxation of a dike in the central MER in 1993 (Bendick et al., 2006). In general, the variability of these data, coupled with extension rates at times significantly lower than the predicted ones and the lack of important amagmatic faulting events, suggests a limited role of the regional tectonic structures in maintaining the extension along temporarily inactive magmatic systems. However, these general considerations are based on short-term information (the global positioning system data, GPS) and therefore more evidence is needed to further test whether amagmatic faulting or creep may still play any role in shaping the boundaries on the longer term.

This study aims at better understanding the longer term evolution of slow spreading magmatic plate boundaries, as well as bridging the gap between the better known shorter term and the less known longer term evolution. For this, we first summarize the main structural features of the transient diking events observed on the shorter term, as obtained from available geological, geophysical, geodetic, and modeling studies. We then put these events into a broader context, considering the overall longer term shallow and deep structure of the boundaries. The synthesis of the collected data allows us to propose an original model, essentially magma driven, for the evolution of the slow spreading magmatic boundaries; the magma may be related to plumes (Iceland and Afar), deglaciation (Iceland), adiabatic melting (MER) or any combination of these factors.

TRANSIENT DIKING EPISODES OR EVENTS

Here we consider and summarize the main structural features of the best-known diking episodes or events that have been occurring on the short term (<10 yr) along the boundaries, presented in chronological order. The definition of the structural features associated with these episodes is the first step in understanding the possible longer term evolution of the cumulative deformation observed along the boundaries.

The 1783 Lakagigar eruption provides an interesting historical and well-constrained example of major diking episode feeding an eruptive fissure on a previously undeformed area. The 27-km-long fissure is in the Grimsvotn magmatic system (East Volcanic Zone of Iceland), and has an overall extension rate of 10 mm/yr (Fig. 2; Perlt et al., 2008). The fissure formed in 1783–1784, when a dike or dikes propagated northeastward, creating 10 en echelon fissures and feeding one of the largest historical basaltic lava flows (~15 km^3) (Thordarson and Self, 1993; Andrew and Gudmundsson, 2008). The northeast-southwest–trending fissure (Figs. 3A, 3B) is partly centered within a 150–450-m-wide graben bordered by northeast-southwest–trending normal faults with mean dip of 75° (Fig. 3). In correspondence to the preexisting Laki hill, the fissure becomes interrupted, but the border faults climb the hill

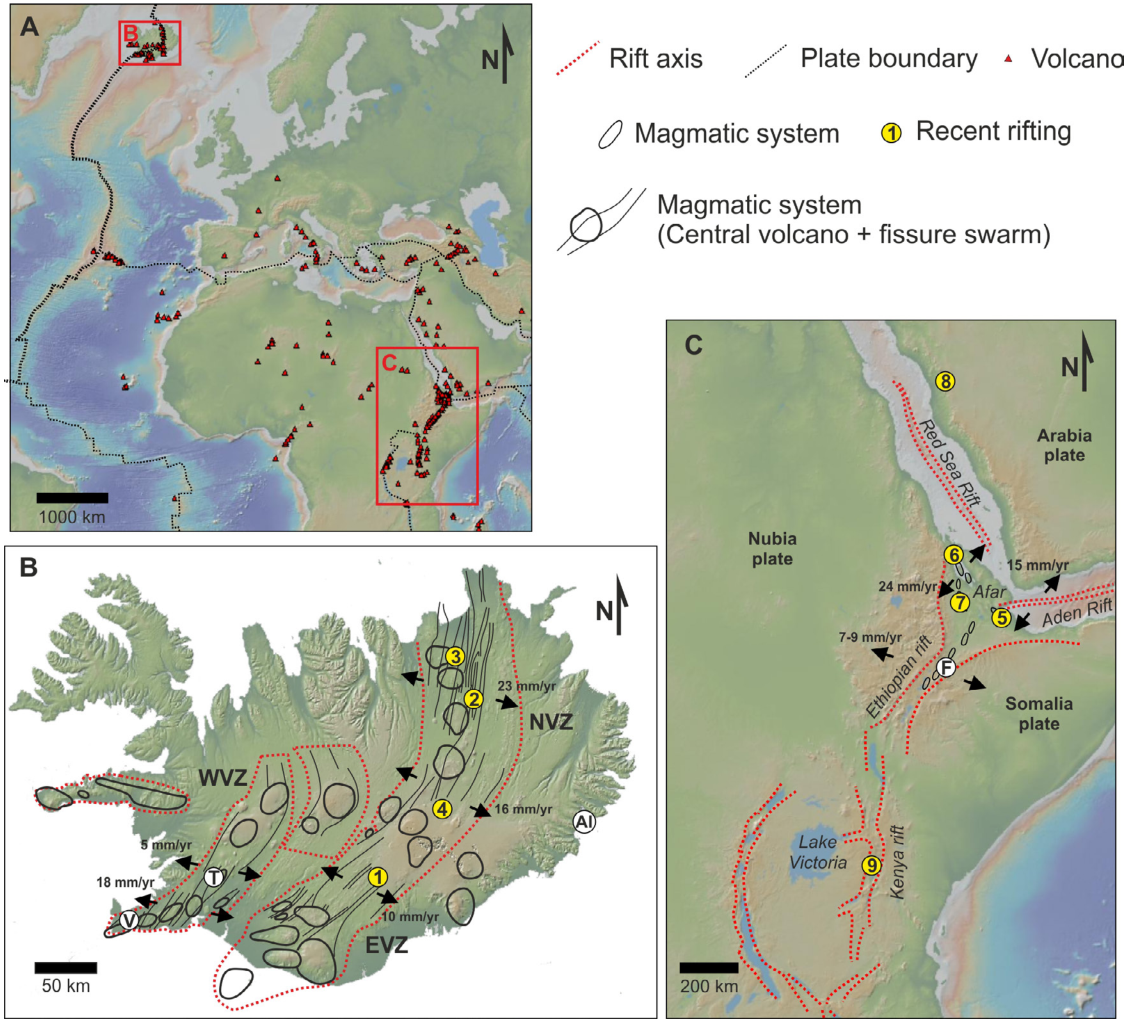

Figure 2. (A) Locations of the considered portions of slow and ultraslow spreading plate boundaries. (B) Oceanic Mid-Atlantic Ridge of Iceland. WVZ—West Volcanic Zone; EVZ—East Volcanic Zone; NVZ—North Volcanic Zone; V—Vogar fissure swarm; T—Thingvellir fissure swarm; Al—Alftafjordur magmatic system. (C) Continental East African Rift System. F—Fantale magmatic system. In B and C, locations of each recent rifting event or episode are indicated by numbers within yellow circles. Black arrows indicate the plate spreading directions (Perlt et al., 2008, for Iceland; Manighetti et al., 2001). 1—Lakagigar 1783–1784; 2—Sveinagjá 1872–1875; 3—Krafla 1975–1984; 4—Bardarbunga 2014; 5—Asal-Ghoubbet 1978–1979; 6—Dallol 2004; 7—Dabbahu 2005–2010; 8—Harrat Lunayyir 2009; 9—Natron 2007.

(Fig. 3D), diverging from each other, becoming as much as 700–800 m apart. The mean vertical throw of the border open normal faults along the fissure is ~3 m, with a mean extension component of ~1.4 m between the footwall and the hanging wall. At times, the deformation on the border faults is locally partitioned into several minor parallel normal faults, resulting in a widespread and less defined border. The eruptive fissure usually is in the middle of the graben, but may also locally coincide with the border fault (Fig. 3B). This implies that the border fault formed before the dike reached the surface and was subsequently reactivated by the rising dike. In addition, available field data suggest that the normal faults propagated from the surface downward (Trippanera et al., 2015a).

In Iceland, the Krafla magmatic system underwent a rifting episode between 1975 and 1984. The magma chamber below the Krafla caldera fed ~20 dikes propagating over a distance of ~80 km, inducing surface fracturing and faulting. Dike-induced faults formed a general but discontinuous graben structure; this was narrower (<1 km) in the caldera area, enlarged to ~2 km in the central portion of the rift, and widened as much as 4–5 km at the northern end, where the resulting asymmetric structure resembled a half-graben with major

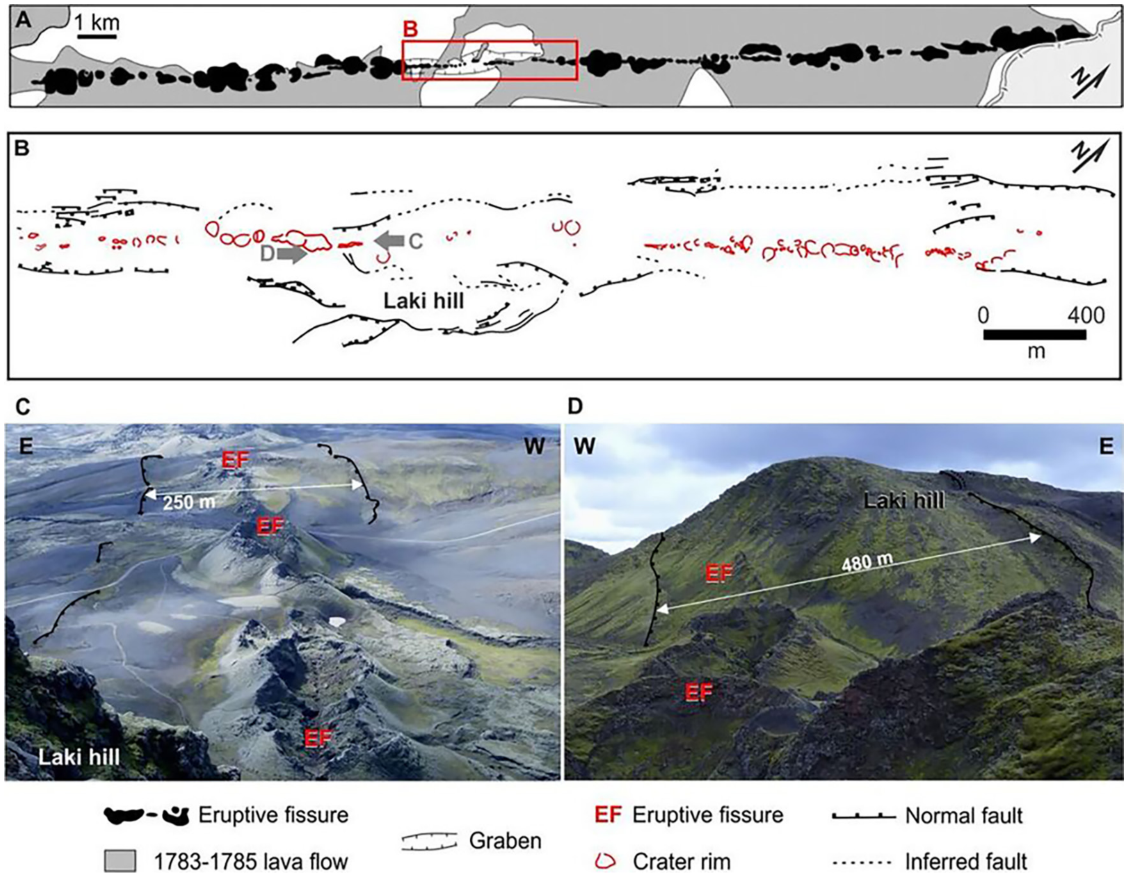

Figure 3. (A) Map of crater rows forming the Lakagigar eruptive fissure, southeast Iceland (from Thordarson and Self, 1993). (B) Field structural map of the Lakagigar eruptive fissure's central portion (location in A). (C) View of the Lakagigar eruptive fissure, located within a graben, from Laki Hill (location in B, gray arrow). (D) View of the Laki hill (location in B, gray arrow); where the eruptive fissure vanishes, the graben faults become more distant.

faults to the east. The northward widening of the graben results from the deepening of the intruded dikes, the depths of which range from 0 km in the caldera region to 3 km at the northern end of the rift (Sigurdsson, 1979; Gudmundsson, 1995b; Hollingsworth et al., 2012; 2013). Faulting reached 2 m of vertical displacement and maximum cumulative widening of ~9 m at 10–12 km north of Krafla, decreasing to 4–5 m to the northern end, suggesting no correlation between rift opening and the distance of dike propagation (Bjornsson et al., 1977; Sigurdsson, 1980; Tryggvason, 1984, 1994; Buck et al., 2006; Hollingsworth et al., 2013). A total of ~250 Mm^3 of lava erupted from the fissures, and ~1–2 km^3 intruded the crust as dikes (Tryggvason, 1984; Hollingsworth et al., 2013). The postrifting excessive spreading rate along the Krafla magmatic system has decreased from >30 mm/yr, observed by GPS in 1987–1992, to 23 mm/yr between 1993 and 2004, slightly higher than the predicted rate of plate spreading (18–20 mm/yr; Arnadottir et al., 2009).

The Asal Rift is the first onland segment of the Aden Ridge in eastern Afar (Ethiopia), and has a spreading rate of nearly 2 cm/yr (McClusky et al., 2010). Asal underwent a rifting episode in November 1978, with the emplacement of a 4.5-km-long and 2.2-m-thick dike accompanied by widespread seismicity; at the same time, an ~8-km-long and ~4-m-thick dike was emplaced below the offshore Ghoubbet Rift, ~15 km to the southeast (Tarantola et al., 1979; Jacques et al., 1996). On the onshore Asal portion, the eruption of basaltic lavas was associated with surface fracturing forming extension fractures as much as 1 m wide and steep normal faults with vertical displacement of as much as 80 cm. The latter created an ~2-km-wide and ~8-km-long graben-like structure, widening southeastward, responsible for an extension of 2 m and subsidence of 70 cm (Tarantola et al., 1979, 1980). Between 1978 and 1986, the Asal Rift opened at a fast rate, mainly magmatically and aseismically; in the ~20 yr after 1986, the opening rate and seismicity decreased, even though still showing sustained mass input and postrifting unsteady opening, higher than the large-scale Arabia-Somalia motion, and suggesting transient variations (Ballu et al., 2003; Cattin et al., 2005; Doubre et al., 2007a; Vigny et al., 2007). The Asal Rift faults are in a critical state of failure and respond instantly to small pressure changes in fluid-filled fractures; this is consistent with the evidence that more recent activity below the Fieale central volcano (Asal Rift) is mainly magma induced or assisted (Doubre et al., 2007b; Doubre and Peltzer, 2007). Repeated dike-induced rifting events at Asal, as in 1978, are inferred to be responsible for the formation and upward growth of normal faults above the dikes; surface faulting is thus compatible with magmatic intrusions, rather than amagmatic faulting (Pinzuti et al., 2010).

Dallol provides an example of smaller surface deformation due to a moderate diking event in 2004, along the axis of a volcanically inactive portion of the onshore Red Sea Rift (northern Afar; Fig. 2). Here the Red Sea Rift consists of a focused (<10 km wide) and ~100-km-long northwest-southeast–trending area of basaltic volcanism and associated fracturing along the Erta Ale Range; this opens at ~10 mm/yr and, in addition to the >100 yr active Erta Ale lava lakes, it has been recently characterized by shallow magma emplacement at Gada Ale and eruptive activity at Dalafilla volcanoes (e.g., Amelung et al., 2000; McClusky et al., 2010; Pagli et al., 2012; Wright et al., 2012). To the north, the Erta Ale Range terminates into a narrow northwest-southeast–trending salt plain, nearly 100 km long and 120 m below sea level: here volcanism is lacking and only the Dallol area, in the center of the salt plain, shows widespread hydrothermal activity located at a 2-km-wide dome formed by the salt deposits. The Dallol rift segment underwent an along-rift dike intrusion in October 2004, triggering a moment magnitude, M_w, 5.5 earthquake on the faults on the western flank of the rift. The northwest-southeast–trending dike, with maximum thickness of ~4 m, propagated parallel to the rift axis south of Dallol for ~7 km. The surface deformation consisted of a northwest-southeast–trending depression, ~4 km wide and a few tens of centimeters deep, without evident surface faulting, even though diking induced nearly 2 m of slip at depth on a nearby rift border fault (Nobile et al., 2012). The cointrusive deflation in the Dallol area suggests that the diking episode originated from a ~2.4-km-deep magma chamber, which has not produced any volcanics (Nobile et al., 2012). Despite the limited development of a graben structure at the surface, Dallol testifies to the birth of immature magmatic segments in a protovolcanic stage, with a null extrusive/intrusive ratio.

A subsequent rifting episode occurred a few hundreds of kilometers to the south in Afar, at Dabbahu (northern Manda Hararo Rift), from 2005 to 2010. The major intrusive event in 2005 emplaced a 60-km-long and 8-m-thick dike. Dike intrusion was accompanied by a minor eruption, >M5 earthquakes, and surface deformation, both elastic and anelastic, creating normal faults and extension fractures, and reactivating preexisting normal faults with displacement of as much as 3 m over a 2.5-km-wide area immediately to the east of the most depressed rift axis (Wright et al., 2006; Ayele et al., 2007; Rowland et al., 2007; Grandin et al., 2009). The 2005 intrusion consisted of 2 laterally propagating dikes from below the Ado'Ale volcanic complex, at a depth between 2.5 and 6 km (Ebinger et al., 2008; Ayele et al., 2009). The fault reactivation suggests that most of the normal faults along the rift axis may result from diking and thus magma may be responsible for the axial relief of a magmatic segment (Rowland et al., 2007); 13 more dike intrusions, for a total volume >3 km^3 of magma, occurred between 2005 and May 2010, sourced from the center of the Dabbahu–Manda Hararo Rift segment (Ebinger et al., 2010; Grandin et al., 2009; Wright et al., 2012). The diking events are followed by a decadal-scale period with extension rates faster than the secular divergent plate motion, mainly resulting from stress relaxation in a viscoelastic upper mantle (Nooner et al., 2009; Pagli et al., 2014).

In July–August 2007, a seismomagmatic crisis in the Natron basin (Tanzania) was accompanied by the first diking event ever captured geodetically in a continental rift. The younger than 5 Ma Natron basin is near the southern termination of the eastern branch of the East African Rift System, separating the Somalian and Nubian plates with current far-field extension rates of 3–4 mm/yr (Stamps et al., 2008). The Natron seismomagmatic crisis began on 12 July 2007 below the southern toe of Oldoinyo Gelai volcano, at depths between 4 and 12 km, and then migrated as much as 10 km to the northeast between 14 and 17 July. Seismic activity continued in July and August and then decayed rapidly in September. No eruption or venting was recorded at Oldoinyo Gelai, but the crisis was accompanied by renewed activity at the nearby Oldoinyo Lengai carbonatitic volcano. Hypocenters delineate a steep plane dipping to the northwest, striking N33°E, associated with a normal fault between 5 and 10 km deep, and slip of 0.5 m from 12 to 17 July. Of the total moment, ~98% must have been dissipated aseismically, and this is the first slow slip event recorded on a normal fault in a continental rift (Calais et al., 2008). Faulting may have been induced by the pressurization of a deep-seated magma chamber below Oldoinyo Lengai (Baer et al., 2008). Field observations in early August and October revealed 2 main north-northeast–south-southwest fracture strands forming a graben 2–3 km wide since July 21; displacements are larger along the middle eastern strand, with an average of 12 cm of opening and 35 cm of vertical offset. Graben formation is consistent with the lateral emplacement of a 7-km-long and ~2-m-thick dike (Baer et al., 2008; Calais et al., 2008; Biggs et al., 2009). Modeling results suggest that such a dike opening was triggered by static stress changes associated with normal fault slip. Therefore, the 2007 Lake Natron rifting episode provides a snapshot of strain partitioning between faulting and magma intrusion processes needed to link discrete rifting episodes with the time-averaged deformation (Calais et al., 2008).

An unusually energetic swarm of >30,000 earthquakes took place in April–June 2009 beneath Harrat Lunayyir, on the Red Sea side of Saudi Arabia. During the peak seismic activity on 19 May, reaching M5.4, a northwest-southeast–trending 8-km-long surface rupture resulted in 45 cm of tensional opening at the surface. This rupture, located in a broad area of deformation, with ~40 cm of uplift and >1 m of east-west extension, mainly consisted of a northwest-southeast–trending graben, bound on the southwest by a main fault and on the northeast by several discontinuities; the location of the graben coincides with that of the main cluster of epicenters from the earthquake swarm.

This deformation field was best modeled by the intrusion of an ~10-km-long, northwest-southeast–trending dike, with a top at <2 km depth and volume of ~0.13 km^3 (Pallister et al., 2010). The graben width decreases from the northwest (6–7 km) to the southeast (<2 km), consistently with the southeast decrease in depth of the dike.

The 2014 Bardarbunga (Iceland) diking episode is the most recent and probably best constrained geophysically. The Bardarbunga magmatic system is in the East Volcanic Zone of Iceland, with an overall spreading rate of 16 mm/yr (Fig. 2; Perlt et al., 2008). The propagation of a ~50-km-long and <5-m-thick dike was seismically and geodetically detected within 2 weeks in August 2014. The dike first had a northwest-southeast direction, radial to the Bardarbunga volcano, then abruptly twisted to become aligned to the north-northeast–south-southwest rift zone; the propagation of the dike was mainly episodic, with an overall discontinuous velocity. Past the Vatnajokull ice cap, at Holuhraun, the dike formed a north-northeast–south-southwest–trending graben, 0.8–1 km wide and ~6 km long; 2 eruptive fissures formed within the graben on 29 and 31 August, feeding a lava flow producing ~1.6 km^3 of lava until February 2015 (Gudmundsson et al., 2014; Sigmundsson et al., 2015). Both fissures are close to older fissures related to the 1910 diking episode, when a lava flow possibly buried a preexisting graben and the associated border faults, forming a flat area: these preexisting border faults have been probably reactivated by this new intrusion during the formation of the graben. Contemporaneously to the propagation and eruption of the dike, the collapse of the Bardarbunga caldera has been detected below the Vatnajokull ice cap, suggesting a lateral magma transfer from the central volcano. An original field survey, carried out in October 2014, defined the structure of the graben along the two border faults immediately to the north of the 29 August eruptive fissure. Both the north-northeast–south-southwest–trending normal faults bordering the 0.8-km-wide graben show a dilational component, with flat footwall and 20° ± 5° inward-tilted hanging wall (Fig. 4). The faults have a mean vertical throw of 6 ± 1 m and the tensile component between the footwall and the hanging wall is ~3 m on each fault. However, the hanging wall is locally split into 2 or more blocks, with a cumulative opening of as much as 7 m. Minor extension fractures, with mean opening of 0.5 ± 0.2 m, are also present on the footwall in a 30–40-m-wide zone parallel to the fault. These fractures, mainly striking north-northeast–south-southwest, merge along strike together and with the main fault (Fig. 4C).

These rifting episodes have shown how dike emplacement, on the short term (<10 yr) and commonly along the axis of the boundaries, creates simple graben structures at the surface. These grabens are usually several hundreds of meters to a few kilometers wide, with normal faults with slip of a few meters, newly formed or reactivated, and with a dilational component. In some cases, as at Dallol, the depth and the opening of the dike induce only minor deformation at the surface, whereas preexisting fractures may be reactivated at depth. A consistent dike-induced surface deformation pattern has been found in the other major diking episodes, which have been occurring in historical time, as at Eldgja (A.D. 933–941; Thordarson et al., 2001) or Sveinagja, in Iceland (1872–1875; Sigurdsson and Sparks, 1978; Gudmundsson and Backstrom, 1991; Tentler, 2005). This indicates that in these rifting episodes extension is accommodated by diking at a depth of a few kilometers and normal faulting in the shallowest part of the crust.

LONGER TERM CUMULATIVE DEFORMATION

Here we consider and summarize the cumulative deformation resulting from the main structural features of portions of magmatic systems (referred to as fissure swarms) formed on the longer term (≥100 yr) along the boundaries. These swarms are considered both at the surface, in the case of active systems, and at depth of a very few kilometers, in the case of eroded extinct systems.

The active portions of magmatic systems studied at the surface are located along the oceanic ridge of Iceland, the transitional crust of Afar, and the continental Main Ethiopian Rift. Unlike the cases previously described herein, the structure of these swarms presents a much higher variability. Rather than providing a long and detailed description of the many fissure swarms, we group all the best-known cases into two main representative categories. These categories display specific and distinct structural features, providing a reference to explain the architecture of most fissure swarms (Trippanera et al., 2015a).

A first group of more symmetric fissure swarms may be exemplified by the Fantale case (Ethiopia). Fantale probably provides the best structural exposures of the Main Ethiopian Rift, whose north-northeast–south-southwest–trending central portion is characterized by active and clear surface deformation with mean extension rate of ~7 mm/yr (Fig. 2C; Fernandes et al., 2004). Here the dominant volcano is the polygenic edifice of Fantale, hosting an east-west–trending summit caldera, whose last historical eruption produced a lava field and a series of rift-parallel cinder and spatter cones in 1820 (Williams et al., 2004). The fissure swarm to the south of the caldera has an overall north-northeast–south-southwest orientation and is characterized by extensional fractures and normal faults that dissect the 168 ± 38 ka welded tuff originating from Fantale. This deformation, developed during the past 7 ka (Williams et al., 2004), is focused in a 3-km-wide and 15-km-long area, characterized by a main graben located between the southern base of the volcano to the lake shoreline farther to the south (Fig. 5A). The 2.5-km-wide graben is delimited by two main normal border faults oriented north-northeast–south-southwest, with constant dilational component of a few meters between the footwall and the locally tilted hanging wall (Fig. 5B). The throw on the border faults reaches 10–15 m on the western and 15–20 m on the eastern border. The mean tilt of the hanging wall is ~30°, but it may locally reach 88°. Within the graben, several extension fractures with openings of 1–2 m are present.

The structure of the fissure swarm south of Fantale can be thus exemplified by an overall graben-like symmetric structure. Even though associated with larger displacements of the normal faults, this structure is very similar to the ones observed (see discussion of transient diking episodes) and produced during the diking episodes or events. Similar polygenic (formed during repeated events) fissure swarms with structures consistent with those of

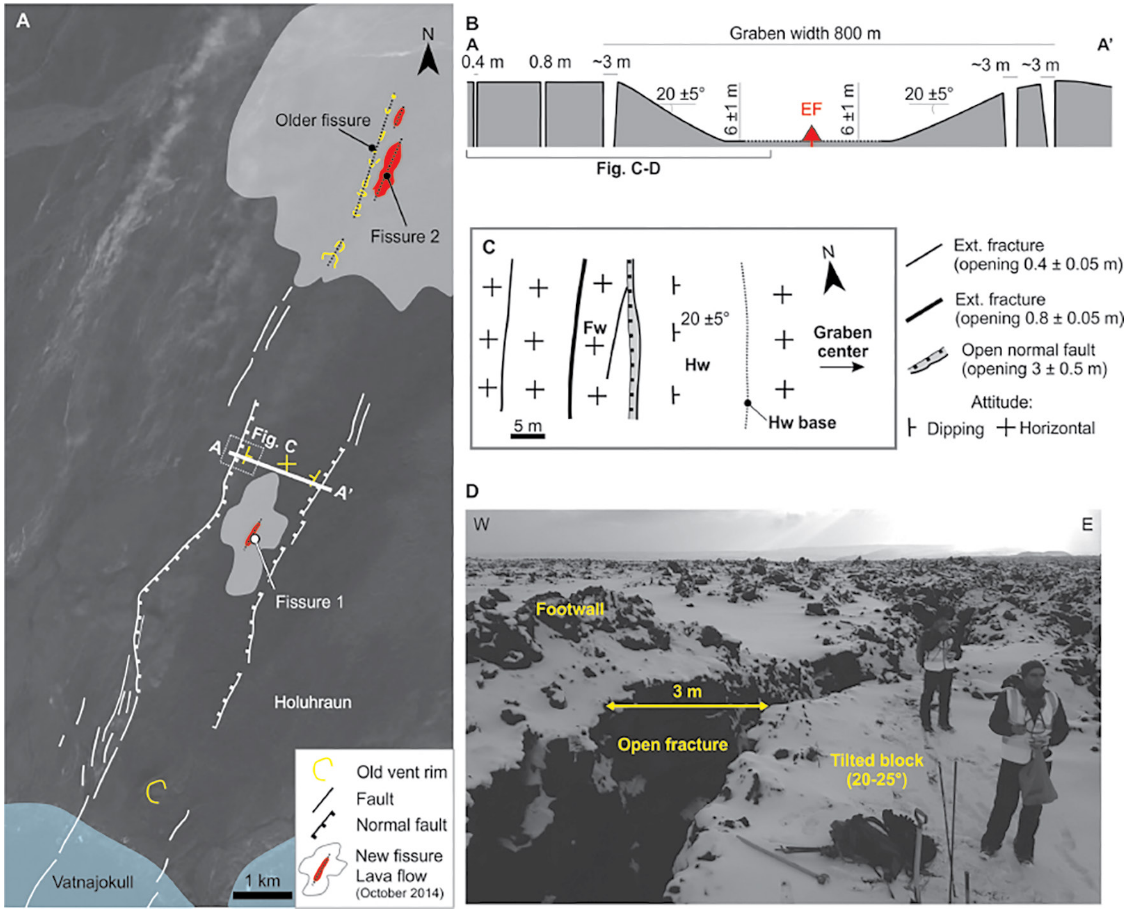

Figure 4. (A) Structural map of the Holuhraun area, along the northern 2014 Bardarbunga fissure, Iceland (location in Fig. 2). (B) Schematic structure of the new-formed graben border faults, along the line A-A′ (not to scale; location in A). EF—eruptive fissure. (C) Detailed field map of portion of the western boundary normal fault displaying a horizontal footwall (Fw), a tilted hanging wall (Hw), and a tensile area between (location in A and B). Ext.—extensional. (D) Photo of area in C.

Fantale are found in the Asal and Dabbahu magmatic segments (Afar; De Chabalier and Avouac, 1994; Ebinger et al., 2010, and references therein) and at Thingvellir (Iceland). The latter is an ~5-km-wide graben delimited by several inward-dipping conjugate normal faults with as much as 40 m of subsidence and 70 m of horizontal extension in the past 9 ka (Gudmundsson, 1987; Saemundsson, 1992; Bull et al., 2005).

A second group of less symmetric fissure swarms may be exemplified by Krafla, one of the most active magmatic systems in Iceland. The 80-km-long and as much as 10-km-wide Krafla magmatic system strikes north-northeast–south-southwest and has a mean extension rate of 23 mm/yr (Fig. 6; Bjornsson et al., 1977; Perlt et al., 2008). In its most active portion, Krafla hosts the ~9-km-wide active Krafla caldera, where most of the 35 Holocene eruptions of the magmatic system occurred (Bjornsson et al., 1977). These eruptions included several rifting episodes, as the 2–1.5 ka basaltic fissure eruptions in Kelduhverfi and spatter cones at Gjastykki, the 1724–1729 Myvatn fires, and the most recent rifting episode in 1975–1984 (Opheim and Gudmundsson, 1989). The central portion of the Krafla magmatic system does not consist of a simple and single graben; rather, there is a 2–3 km, wider deformation zone, partly asymmetric, with several extension fractures and normal faults within, defining 30–300-m-wide minor nested grabens (Fig. 6). The faults are usually vertical at the surface, with a tensile area between the footwall and the hanging wall; the largest fault, defining the western part of the deformation zone, has a maximum throw of 42 m and a maximum opening of 28 m (Opheim and Gudmundsson, 1989; Angelier et al., 1997). The opening of the extension fractures is a few meters, usually <1 m, smaller than the mean width of the tensile area of the normal faults.

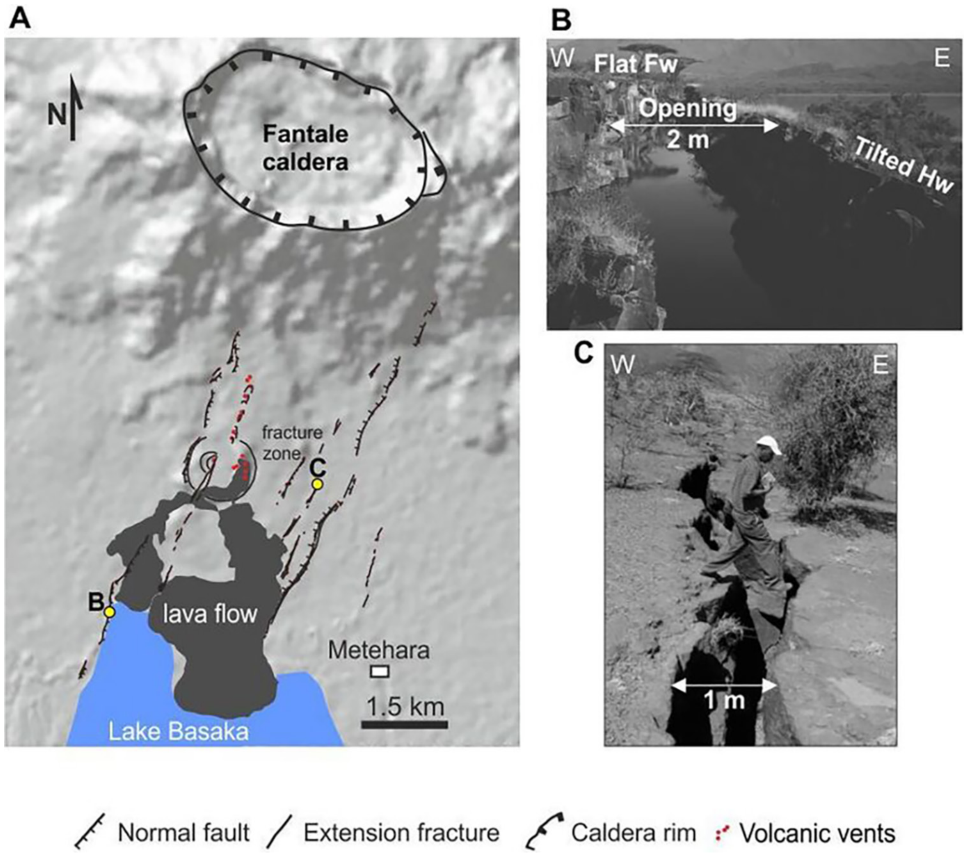

Figure 5. (A) Structural map of the southern portion of Fantale magmatic system; Main Ethiopian Rift. (B) Open normal fault with tilted hanging wall (Hw) forming the western border of the graben (location in Fig. 4A). Fw—footwall. (C) Meter-scale open fracture located within the major graben (photo by J. Ruch; location in Fig. 4A).

The structure of the central portion of the Krafla magmatic system thus appears irregular, with a faulted area with nested grabens within, and asymmetric. Rift segments with similar structure include the southern Manda Hararo Rift (Afar; Acocella et al., 2008), Gedemsa (Main Ethiopian Rift), and Vogar (Iceland). Gedemsa consists of an area of distributed but asymmetric deformation as much as 15 km wide, mainly shaped by the activity of west-facing fault scarps, with displacement of a few tens of meters, and by minor graben-like structures, <2 km wide; the general westward dip of the fault systems may be related to the proximity to the nearby eastern border of the rift (Corti, 2008, 2009). Vogar consists of a 3–4.5-km-wide asymmetric deformation zone (or half-graben) delimited on the western edge by an east-facing normal fault and on the eastern edge by a series of open fissures (Villemin and Bergerat, 2013).

The presence of regular and symmetric (i.e., Fantale case) and irregular and asymmetric (i.e., Krafla case) structures on continental, transitional, and oceanic crust highlights their independence from the nature of the rifted crust and the considered spreading rate range (from a few millimeters to a few centimeters per year).

The deep portion of a magmatic system may be appreciated along the eroded portions of extinct rift zones; Iceland provides such a rare opportunity. Here the Neogene and Pleistocene bedrock (12–8 Ma) crops out to the eastern and western sides of the active neovolcanic zone (Saemundsson, 1979). Several studies have been made on the structure of the deeper rift zone, and more in particular about dike and fault geometry and related crustal dilation, in southwest, northwest (Gudmundsson, 1990; Forslund and Gudmundsson, 1992), and east Iceland (Walker, 1958, 1960, 1963, 1974; Gudmundsson, 1983; Helgason and Zentilli, 1985; Forslund and Gudmundsson, 1992; Paquet et al., 2007; Urbani et al., 2015). Here we focus on eastern Iceland, where the original top of the lava pile was estimated to be ~1.3 km above the present top of the eroded lavas (Walker, 1960), making this an ideal site to study the deeper portion of an extinct axial rift portion. Here the bedrock is mainly constituted by

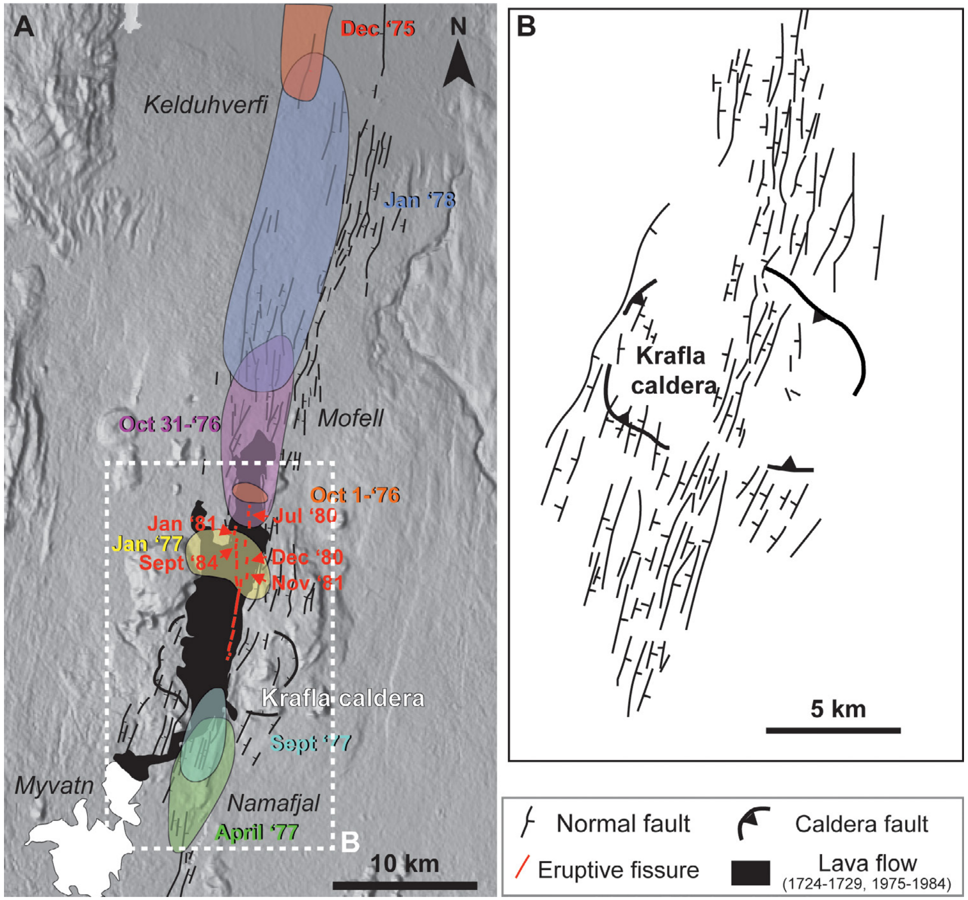

Figure 6. (A) Structural map of Krafla magmatic system, Iceland; crater rows produced during the last rifting event (1975–1984) are indicated in red (Saemundsson, 1991, *in* Bjornsson et al., 2007). Activated areas of the magmatic system in the period 1975–1978, identified locating earthquake epicenters, are highlighted with different colors (Einarsson and Brandsdóttir, 1980, *in* Bjornsson et al., 2007). (B) Detail of the structure of the central part of the Krafla magmatic system, showing the distributed and asymmetric deformation due to normal faulting (Sigurdsson, 1979).

6°–10° west-southwest–dipping lavas originating from the activity of at least 4 volcanic centers (Fig. 7A) and several fissure eruptions (Walker, 1974, and references therein). The lava pile is frequently interrupted by subvertical and parallel dikes, with mean thickness of ~3 m (Walker, 1958, 1960, 1974; Gudmundsson, 1983; Helgason and Zentilli, 1985; Paquet et al., 2007). Dikes are not uniformly distributed but cluster in 4 approximately north-south–oriented swarms generally 5–10 km wide and tens of kilometers long. Each swarm is also associated with a remnant volcanic center, therefore identifying a fossil magmatic system (Walker, 1974; Gudmundsson, 1995a). Most of the dikes focus along the central part, or axis, of the swarms and close to the volcanic centers, where the dike-induced host-rock dilation reaches 8%, gradually decreasing to 2% on the sides of the swarms (Fig. 7A; Walker, 1974). This percentage of dilation and the dike frequency also gradually decrease upward within the lava pile. This is shown in Figure 7B, where the number of dikes per 1.6 km measured in eight swaths along the Alftafjordur and Breiddalur volcanic systems diminishes with altitude (Walker, 1960). In general, only ~20% of the dikes observed at depths of ~1.5 km reach the surface (Trippanera et al., 2015a). A progressive thinning of the dikes (maximum thickness of 1–2 m) toward shallower crustal levels has been also observed (i.e., Helgason and Zentilli, 1985). No direct geometric connection between the upper dike tips and the bottom of the normal faults has been observed (Gudmundsson, 1983). At these crustal levels faults are very rare compared to dikes. Normal faults become gradually more abundant toward shallower levels (<1 km), where they have an average throw of 2.7 m, accounting only for 0.5%–1% of crustal dilation (Bodvarsson and Walker, 1964; Lomize, 1976; Gudmundsson, 1995a; Helgason and Zentilli, 1985; Paquet et al., 2007). These observations suggest

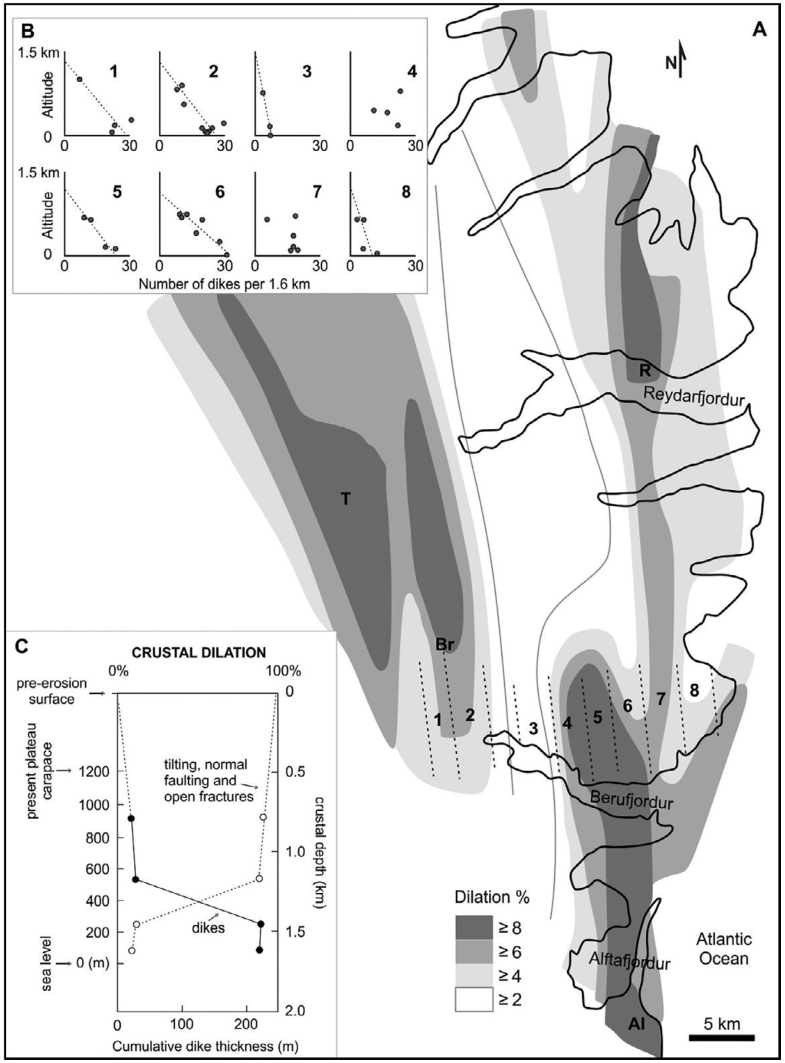

Figure 7. (A) Percentage of bedrock dilation due to dike intrusions in eastern Iceland (redrawn from Eriksson et al., 2014, and references therein). The areas with highest dilation values (dark gray) are approximately north-south elongated, highlighting the axis of at least 4 extinct magmatic systems composing the Alftafjordur (Al), Breiddalur (Br), Reydarfjordur (R) and Thingmùli (T) remnant central volcanoes. (B) Diagrams showing the number of dikes per 1.6 km at different altitudes measured across 8 swaths north to Berufjordur (from Walker, 1960). Number of each diagram refers to A. Each swath is delimited by two dashed lines. (C) Diagram of crustal dilation of faults versus dikes for western Reydarfjordur. Crustal extension is mainly due to dikes at depths >1.5 km, whereas at depth <1.2 km the effects of tilting, normal faulting, and the opening of fissures gradually increase (Helgason and Zentilli, 1985).

an inverse relationship between faults and dikes (Fig. 7C; Helgason and Zentilli, 1985; Forslund and Gudmundsson, 1991, 1992); dikes are abundant at deeper crustal levels (>1 km), where they accommodate almost all extension and faults are rare, whereas dikes (also feeding eruptive fissures) are less frequent at the surface, where the crustal dilation is mostly accommodated by extension fractures and normal faults (i.e., Helgason and Zentilli, 1985; Urbani et al., 2015).

These data on the deeper (1–2 km depth) structure of extinct magmatic systems in east Iceland show that diking is by far the predominant mechanism of extension at depth, where normal faulting plays a negligible role. However, the frequency of the dikes decreases toward the surface, where normal faults become more frequent. Overall, an ideal section view of a magmatic segment consists of an upper thin layer of normal faults and extension fractures, with few feeder dikes, overlying a deeper dike complex becoming more closely spaced at depth, likely fed by sill-like reservoirs.

DISCUSSION

Merging Available Data

The eroded magmatic systems in Iceland show that at a depth of ≤1 km dikes take up most of the extension. At shallower depths their frequency significantly decreases, so the importance of dikes in determining crustal extension decreases toward the surface, where normal faults become more frequent and important in controlling crustal extension; a minor portion of dikes reaches the surface as feeders, systematically developing a depression at the surface. The dike-induced depression may show distinct features, depending on the thickness and depth of the top of the dike (Fig. 8A): deeper (>2–3 km) or thinner (a few meters) dikes may induce mainly elastic deformation at the surface, with a broad depression and minor faulting to the sides (as at Dallol and, partly, Lake Natron); shallower (~1 km) and thicker (several meters) dikes, including feeder dikes, will induce anelastic deformation at the surface, forming a graben-like structure bordered by opposite-dipping normal faults (e.g., at Laki, Dabbahu, and Bardarbunga; Trippanera et al., 2015a). Dike-induced field and experimental models (Forslund and Gudmundsson, 1992; Gudmundsson, 1992; Acocella et al., 2003; Trippanera et al., 2015a, 2015b) suggest that these dike-induced normal faults, promoted by the effect of a free boundary, usually form at the surface and propagate downward. This important feature may explain the general decrease in the frequency of the faults with depth, as well as the observed lack of physical connection between the bottom of the faults and the dike tips (Gudmundsson, 1983). Several analytical and experimental models support the structure (geometry and kinematics) obtained during dike-induced transient rifting episodes, showing how diking systematically forms a depression that evolves into a graben-like structure at the surface (Pollard et al., 1983; Mastin and Pollard, 1988; Rubin and Pollard, 1988; Rubin, 1992; Gudmundsson, 2003; Trippanera et al., 2015b).

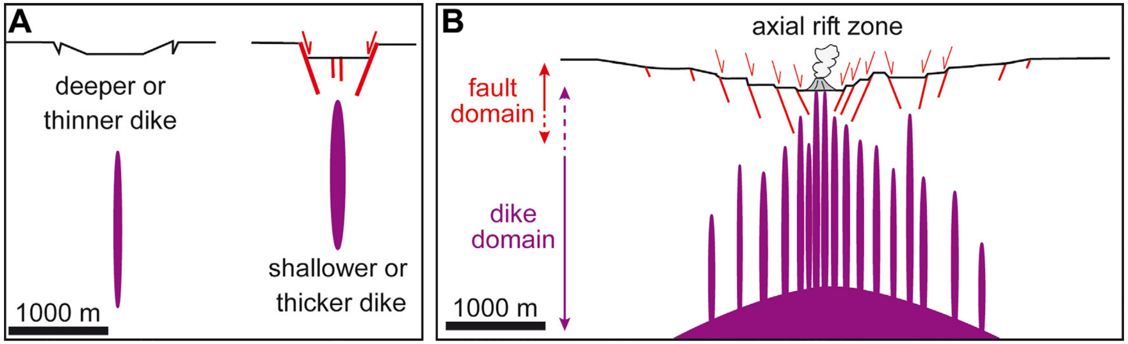

Figure 8. (A) Surface deformation induced by deeper or thinner dikes and shallower or thicker dikes. (B) Layered upper crustal structure of the axial portion of a magmatic divergent plate boundary. Extension due to normal faults (propagating downward from the surface) becomes progressively more significant close to the surface (<1 km depth), whereas at greater depths (>1 km) extension occurs through dike emplacement.

There is a close similarity between the surface deformation of these transient diking episodes and the cumulative one observed along fissure swarms, highlighted by the grabens in the fissure swarms (see discussion of longer term cumulative deformation; Fantale, Thingvellir, and Asal). These grabens, a few kilometers wide and bordered by normal faults with displacement of tens of meters, resemble a more strained version (with higher displacement) of the grabens formed by a diking episode. This suggests that these larger rift portions may be also dike induced, albeit resulting from repeated and focused intrusive episodes. The association between the deformation induced by the transient diking episodes and the group of fissure swarms with irregular and asymmetric structure surface (as at Vogar, Krafla, Gedemsa, and the southern Manda Hararo Rift) is less straightforward. However, this may be explained by the emplacement of distributed (not focused) dikes below a magmatic system, at times (Gedemsa) controlled by the topography of the nearby rift border; distributed parallel intrusions may be thus responsible for the development of irregular and asymmetric fissure swarms. The fact that the longer term or cumulative structure of a fissure swarm may be largely or entirely dike induced (provided that magma is available) is also supported by experimental models investigating the surface deformation due to repeated dike injections (Trippanera et al., 2015b).

All this evidence shows how transient diking episodes systematically affect the topography of a rift and suggests that most, if not all, of the deformation observed along the boundaries may in principle result from repeating diking episodes. To try to assess this latter possibility, we compare the structure of the eroded and active magmatic systems of Iceland. In the eroded rift portions, the frequency of the dikes decreases upward; only a fraction (<20%) of the dikes found at a paleodepth of ~1.5 km reach a paleodepth of ~0.5 km. For 100 ± 20 dikes emplaced at ~1.5 km (Walker, 1960; Trippanera et al., 2015a), <20 dikes are thus expected to reach ~0.5 km depth below a magmatic system. The thicker and shallower (<1 km) dikes will induce a slip on the normal faults at the surface of 1–3 m, consistent with geodetic and geological observations in Afar, the Red Sea, and Iceland (e.g., Rubin and Pollard, 1988; Rowland et al., 2007; Pallister et al., 2010). Assuming that 10–20 dikes may reach the uppermost 0.5 km in the active portion of a rift zone, one may expect a cumulative amount of vertical deformation at the surface between 10 m (assuming the lower bound of the mean slip value, 1 m, for 10 dikes) and 60 m (assuming the upper bound of the mean slip value, 3 m, for 20 dikes). These values are similar to the common vertical displacements carried either along a few major faults or several minor faults currently observed at the surface along the active fissure swarms, as at Krafla, Thingvellir, Vogar, and Fantale; this suggests that the surface structure of these fissure swarms may be induced by repeated diking events. Moreover, since single diking episodes, as recently observed at Bardarbunga, may induce surface faulting with vertical slip of 6–7 m or more, fewer shallow dikes may also explain the large surface deformation observed along active rift segments. These results suggests that shallow dike emplacement (<1 km) may justify all the observed deformation along the considered rift segments, so that the overall shape, structure, and development of these boundaries may be essentially magma induced (Fig. 8B). A magmatic system may be thus envisaged as predominantly, if not entirely, resulting from dike emplacement; while deeper and thinner dikes may have a limited impact on the mainly elastic surface deformation, shallower and thicker dikes may induce faulting at the surface, with vertical displacement of several meters (Fig. 8B). Available data suggest that the depth of the dike top at which surface faulting begins is ~1 km; deeper dikes, even if thick, should mainly induce elastic deformation, whereas shallower dikes, if at least a few meters thick, will fault the host rock at the surface.

Model

To summarize the possible general tectonomagmatic relationships along the magmatic slow spreading plate boundaries, we first focus on the best-constrained case, Iceland, the main rifting events of which are listed in Table 1 (Sigmundsson, 2006, and references therein; Gudmundsson et al., 2014; Sigmundsson et al., 2015). Here the magmatically active East and North Volcanic Zones (EVZ and NVZ, respectively) have been characterized in the last centuries by repeated rifting episodes due to the emplacement of dikes feeding fissure eruptions; the West Volcanic Zone and most of the Reykjanes Peninsula

TABLE 1. RECENT RIFTING EVENTS OR EPISODES IN ICELAND, AFAR, SAUDI ARABIA, AND KENYA-TANZANIA RIFT

Location	Years (A.D.)
Iceland	
Bardarbunga	2014–2015
Krafla: Krafla Fires	1975–1984
Askja - Sveinagjá	1874–1875
Trollagigar	1862–1864
Thingvellir	1789
Laki	1783–1784
Krafla: Myvatn Fires	1724–1729
Veldivotn	1480
Eldjà	934
Vatnadlur	871± 2
Afar	
Dabbhau	2005–2010
Dallol	2004
Asal-Ghoubbet	1978–1979
Kenya Rift	
Lake Natron	2007
Saudi Arabia	
Harrat Lunayyir	2009

Note: References: Bjornsson et al. (1977); Ruegg et al. (1979); Tarantola et al. (1980); Tryggvason (1984); Rubin and Pollard (1988); Wright et al. (2006); Sigmundsson (2006), and references therein; Biggs et al. (2009); Pallister et al. (2010); Ebinger et al. (2010); Nobile et al. (2012); Gudmundsson et al. (2014); Sigmundsson et al. (2015).

lack a similar frequency and are less active; therefore, here we focus on the EVZ and the NVZ. The recurrence period of the rifting episodes in a same rift portion is ~200 yr for the EVZ and, with more limited information, ~250 yr in the NVZ; however, in the last few thousand years the frequency of the rifting episodes in the NVZ may have been significantly lower (F. Sigmundsson, 2015, personal commun.). In a first approximation it may be considered that, within the better known last centuries, each portion of both the EVZ and the NVZ usually undergoes a rifting episode every ~200 yr; this may be the duration of a rifting cycle in any given portion of the rift.

The known or inferred mean thickness of the dike or dikes responsible for these rifting episodes ranges from ~4 m (as in 2014 in Bardarbunga; Sigmundsson et al., 2015) to more common values of 6–13 m for the 8 ka Sveinar dike (Gudmundsson et al., 2008), the 1783 Laki dike (T. Thordarson, 2012, personal commun.), and the 1975–1984 Krafla dikes (Sigmundsson, 2006, and references therein). Therefore, the mean dike-induced strain rates along the axis of the EVZ and NVZ, given by ~4–9-km-thick dikes emplaced every ~200 yr, range from 2 to 4.5 cm/yr. The mean spreading rate along the plate boundary of EVZ and NVZ is ~2 cm/yr, thus on the lower bound of the magmatically induced strain rates. Therefore, thinner dikes will release the tectonic spreading rates in just a mean rifting cycle, whereas thicker dikes will produce a strain rate in excess of the mean spreading rate, possibly inducing local and transient contraction to the sides of the dike (Fig. 9A; Gudmundsson et al., 2008); this contraction may eventually be relieved, reversing into extension, by plate spreading after several decades. However, the general point here is that the strain rate promoted by diking is expected to be similar or in excess of as much as ~2.5 cm/yr (4.5–2 mm/yr) with regard to the regional one. This implies that, before reaching the critical amount of regional extension required to trigger, seismically or not, any amagmatic faulting or creep, a dike-induced rifting episode will likely reactivate the cycle after ~200 yr; this reactivation will promote temporary contraction along the rift axis and further prevent any amagmatic faulting for the successive ~200 yr (Fig. 9A). In addition, minor dike injections, not necessarily reaching the surface or feeding any eruption, may also occur between the main dike-induced rifting events, further increasing the contraction and thus preventing any amagmatic faulting along the plate boundary. Therefore, diking appears more effective and faster in extending the plate boundary, always anticipating any amagmatic faulting or creep. Provided that magma is available at depth, this leads to a general and repeated activation of the plate boundary through diking episodes, with negligible role of seismic or aseismic amagmatic faulting or creep, not matching the magmatic rates; in this context, any amagmatic faulting becomes subordinate. For Iceland, this model is also in agreement with the fact that any major seismicity not related to dike injection, with earthquakes reaching M7, focuses in the amagmatic (Iceland South and North Seismic Zones) or poorly magmatic (Reykjanes Peninsula) portions, and is generally lacking along the volcanic portions during interdiking episodes; this confirms that large tectonic stress is not usually accumulated along the volcanic portions, where plate spreading is instead episodically achieved through diking.

The model proposed in Figure 9 is based on the spreading rate and recurrence time of rifting episodes from the EVZ and NVZ of Iceland. Nevertheless, with minor modifications it may be applied to other transitional and continental magmatic boundaries with slow or possibly ultraslow spreading. As far as transitional boundaries are concerned, as at Afar, there is extremely limited information of the frequency of the rifting events (e.g., Medynski et al., 2013). Nevertheless, the overall spreading rates along the Red Sea and Aden Rifts of Afar are similar or only slightly smaller than those detected in Iceland (McClusky et al., 2010, and references therein); moreover, Afar has had at least four rifting episodes (Asal-Ghoubbet, Dallol, Dalafilla, and Dabbahu) in the past four decades, double those occurring in Iceland in the same period. These rifting episodes show how the short-term (years or less; corifting) and even the slightly longer term (years or decades; postrifting, as at Dabbahu or Asal; Hofton and Foulger, 1996; Vigny et al., 2007; Pagli et al., 2014) strain in Afar results from the shallow emplacement of magma. Limited seismicity occurred in the magmatic systems outside the rifting episodes, as in the Asal Rift between 1997 and 2005, but still due to the overpressurization of the magmatic system beneath Fieale (Doubre and Peltzer, 2007); moreover, this occurred during the postdiking crustal relaxation following the 1978 Asal-Ghoubbet rifting episode,

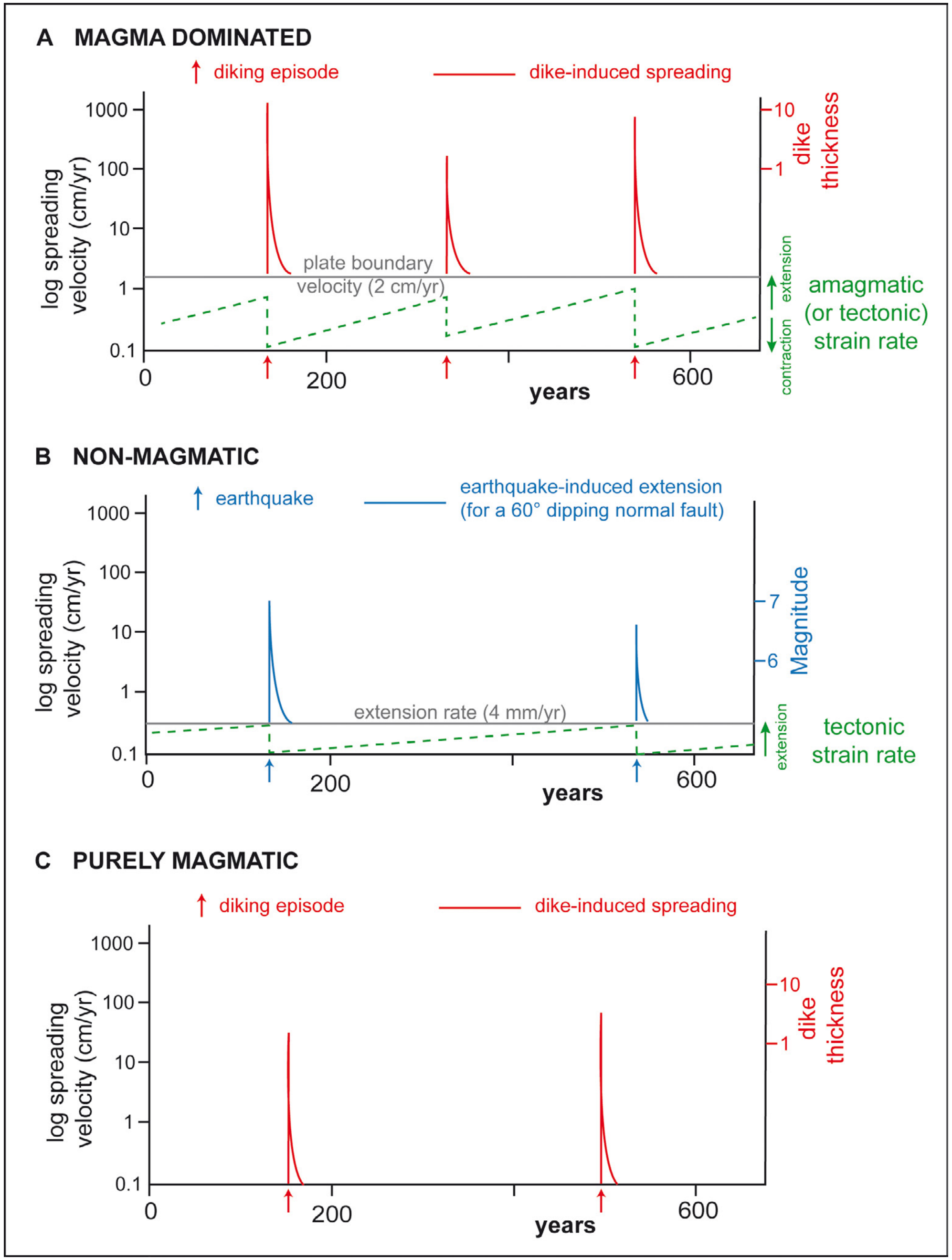

Figure 9. (A) General model summarizing the mode of magma dominated (or dike induced) extension along slow spreading plate boundaries, with particular application to the East and North Volcanic Zones of Iceland; here extension is achieved through diking episodes (red lines) with recurrence of ~200 yr, responsible for a temporary increase of several orders of magnitude of the plate boundary spreading rate. Soon after dike emplacement, the amagmatic (or tectonic) strain rate (green lines, in cm/yr) becomes negative, indicating local contraction; this becomes again positive (indicating local extension) after several decades, but does not usually reach values high enough to induce amagmatic faulting before the next diking event. (B) An end-member model of crust undergoing amagmatic ultraslow extension, represented by the Apennines of Italy; here extension is largely achieved through M < 7 earthquakes (same magnitude as recorded in Iceland) with a recurrence time of a few hundred years. Here the tectonic strain rates (green lines) promote faulting. The example of the Apennines helps in understanding the behavior of amagmatic portions of ultraslow divergent plate boundaries, also highlighting the discrepancy between the more effective extension induced by diking (spreading rates of 10^2–10^3 cm/yr) and the much more limited one induced by amagmatic faulting (spreading rates of 10 cm/yr). (C) An end-member model of crust undergoing purely magmatic extension, without any regional tectonic contribution. Here extension is exclusively achieved through the repeated injection of dikes, possibly with lower frequency and relaxation time with regard to the tectonically assisted case (A). While case C remains largely ideal, it may find application in ultraslow spreading oceanic ridges and/or in volcanic rift zones, even though in the latter case any gravitational instability of the volcano flanks may simulate a far-field tectonic behavior assisting magmatic extension.

in a context where the geometry of faulting and topography is related to longer term diking intrusions at depth (Vigny et al., 2007; Pinzuti et al., 2010). These features confirm that the transitional crust of Afar is magmatically active and its behavior may be not very different from that proposed for Iceland (Ebinger et al., 2010; Buck, 2013; Desissa et al., 2013).

On magmatic continental boundaries, as on the Ethiopia and Kenya domes, the spreading rate is a few millimeters per year, typical of ultraslow rifts. Here no detailed information on recent and historical rifting activity (except for the 2007 Lake Natron event) is available, challenging our understanding of the tectonomagmatic relationships. In general, it may be expected that the decrease of the magmatic activity within a rift increases the importance of amagmatic processes (faulting, creep) in shaping the rift structure. To better understand the possible evolution of ultraslow boundaries and also relate these to the magmatically induced extension of the Icelandic case in Figure 9A, we first consider an end-member example of amagmatic extension (Fig. 9B). This relates to the Apennines of Italy; the 50–100-km-wide inner part parallel to the chain undergoes extension (Jolivet et al., 1998; Acocella and Funiciello, 2006). Here the ~4 mm/yr of extension is partly achieved seismically, being the largest earthquakes of $6 < M < 7$, with average recurrence times, for any given portion, between 200 and 600 yr (Devoti et al., 2011; D'Agostino, 2014). These earthquakes usually induce a fault slip of 0.1 (for M6) to 1 m (for M7) (Wells and Coppersmith, 1994). For a 60° dipping normal fault only ~0.58% of the total slip is transformed into horizontal extension; this implies that the earthquake-associated extension is a few tens of centimeters or less (e.g., Pantosti and Valensise, 1990; Anzidei et al., 2009). While the Apennines cannot be definitely ascribed as a slow spreading plate boundary, their slow extension is achieved amagmatically, providing an ideal example of crust extending only through regional tectonic processes. The Apennines provide an example of how tectonic extension may be achieved along the amagmatic (or poorly magmatic) portions of ultraslow divergent plate boundaries, for example, as for a significant part of the East African Rift System and, more important, show how any $M < 7$ earthquake (the largest commonly observed along any magmatic divergent plate boundary) may induce no more than a few tens of centimeters of horizontal extension, with recurrence intervals of some hundreds of years along the same fault. As seen in comparing Figures 8A and 8B, this extension is at least one order of magnitude smaller than that created by an ordinary diking episode in Iceland, that is, several meters. This difference of ≥1 order of magnitude in the extension highlights the general difficulty of amagmatic faulting in keeping up with diking in extending a crust, even in an ultraslow spreading boundary. It may be argued that the slow spreading of Iceland may promote diking more easily than the ultraslow spreading of the magmatic portions of the East African Rift System. However, available information suggests that this is not necessarily the case. The 2007 Lake Natron rifting event, associated with both faulting and diking, provides a compelling example for this imbalance between magmatic and amagmatic extension. Despite the controversial nature of faulting, probably also magma induced (Baer et al., 2008), the detected extension due to diking (2.4 m) was much larger than that created by the slip of the fault (≤0.2 m) (Calais et al., 2008); this underlines the effectiveness of diking in extending the crust with regard to any amagmatic faulting, supporting theoretical considerations (Buck, 2006). Limited information is available on the seismic activity of magmatic ultraslow spreading boundaries in the absence of clearly established diking events. For example, the seismicity along the Main Ethiopian Rift between 2001 and 2003 focuses on the zones of Quaternary magmatism and faulting, above the 20-km-wide mafic intrusions that rise to 8–10 km subsurface; the seismogenic zone produces largely extensional earthquakes above these intrusion zones, suggesting that intrusion may trigger faulting in the upper crust (Keir et al., 2006). All this independent evidence suggests that in the magmatic portions of the ultraslow spreading East African Rift System, diking is still the easier and thus predominant means of extending a crust, inhibiting amagmatic faulting and thus largely supporting the proposed model of locked amagmatic faulting promoted by diking. While the model of Figure 9A can be best calibrated on the oceanic slow spreading boundary of Iceland, there is widespread evidence that its general lines may hold for the ultraslow spreading of transitional and continental magmatic boundaries. However, some ultraslow spreading oceanic plate boundaries may reach permanent or temporary conditions with negligible or no regional tectonic extension (i.e., <1 mm/yr; Fig. 9C). In this end-member case, extension is exclusively achieved through the repeated injection of dikes, possibly with lower frequency and relaxation time with regard to the tectonically assisted case of Figure 9A.

Our proposed model underlying the importance of magmatic activity (diking) in rifting slow and ultraslow spreading boundaries supports previous studies, mostly referring to the Afar case (Wright et al., 2006; Sigmundsson, 2006; Buck, 2006; Rowland et al., 2007; Ebinger et al., 2010; Wright et al., 2012). However, our study aims to take into account the more general spectrum of continental, transitional and oceanic slow and ultraslow spreading magmatic boundaries (for an overview see van Wyk de Vries, 2015), thus suggesting a general mechanism. This mechanism appears independent of the stage of rifting, at least in magmatic boundaries, characterized by crustal dike complexes. Where crustal dike complexes are lacking, as in the amagmatic portions of continental divergent plate boundaries, amagmatic faulting or creep still appears the dominant mode of extension. In simpler terms, when there is no magma, the rift is fault controlled, and when there is magma available, it is magma controlled. More important, our proposed mechanism provides an original justification for the self-consistency of diking in shaping these boundaries; this justification is in the fact that any amagmatic faulting becomes hindered by the emplacement of dikes, a process that requires hundreds of years to relax the crust and does not usually allow the buildup tectonic stresses in the crust.

CONCLUSIONS

This overview suggests that most, if not all, of the shorter and longer term surface structure of the magmatic slow and ultraslow spreading plate boundaries can be related to magmatic activity, or diking with associated faulting.

This regards single diking episodes and events forming symmetric minor grabens (as Lakagigar, Dallol, and Bardarbunga), as well as portions of magmatic systems characterized by the repeated focused or distributed emplacement of dikes below symmetric or asymmetric larger graben structures (including Krafla and Fantale). Dike-induced graben-like structures thus appear as the basic and recurrent structural feature, formed by normal faults usually propagating downward from the surface. We propose a general mechanism, independent of the stage of rifting, that highlights the self-consistency of diking in shaping any magmatic slow and ultraslow spreading plate boundary. This is due to the fact that any amagmatic faulting becomes essentially hindered by the emplacement of dikes, a process that requires hundreds of years to relax the crust and usually does not allow the buildup tectonic stresses in the crust.

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