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The effect of thermal weakening and buoyancy forces on rift localization: Field evidences from the Gulf of Aden oblique rifting

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ABSTRACT

On the basis of field and geophysical data, analog and numerical models, we here discuss the role of buoyancy forces arising from thickness variations in the lithosphere during rifting. In the Gulf of Aden, an oceanized Tertiary oblique rift, several successive directions of extension and associated normal faults suggest that transient stress rotations occurred during rifting. Especially, rift-parallel faults (070°E) overprinted the early divergence-perpendicular normal faults (110°E). Moreover, some first-order differences are noticeable between the western part of the Gulf, which deformed under the Afar hot spot influence, and the eastern part. In the western Gulf of Aden, the ocean–continent transition (OCT) and the oceanic ridge have cut obliquely through the inherited and reactivated Mesozoic basins (100°E to 140°E). The OCT trend is parallel to the overall Gulf trend (070°E). In the eastern part, the oceanization occurred within few syn-rift 110°E-trending basins and the OCT trends mostly perpendicular to the divergence direction. Here, we propose that this contrast is strongly controlled by the Afar hot spot: during rifting times, the hot spot likely induced a hot thermal anomaly in the western asthenosphere. This may have triggered both thermal buoyancy forces and thermal weakening of the lithosphere that helped localizing the rift obliquely. In such localized rift, rift-perpendicular trending crustal buoyancy forces (i.e. around 160°E) have enhanced rift-parallel normal faults (070°E) during final rift localization into a narrow zone strongly oblique to the early syn-rift basins. As a consequence of the Afar hot spot, in the west, the ridge is long and straight; in the east, the ridge segments are rather long too (although less than in the west) as the ridge initiated parallel to the OCT; in between, the ridge is more segmented as both the hot spot influence gradually decreases eastward and the ridge initiated obliquely to the OCT.

1. Introduction

Rift localization occurs in the lithosphere when extensional strain focuses in a relatively narrow zone, possibly leading to continental break-up and oceanization. Such localization is obviously witnessed by rifts that evolved into oceanic basins wherein strain localization occurred in the distal passive margin, leading to mantle exhumation at some margins (Beslier et al., 1996, 2004; Boillot et al., 1987; Bonatti et al., 1990) and subsequent oceanic spreading. Strain localization is also documented in active rifts where old faults are located on the rift shoulders while new and active ones are observed in the rift center, along with seismicity and volcanism, as in the East African Rift System (EARS; Corti, 2009; Ebinger and Casey, 2001; Ebinger et al., 1993; and references therein). On the contrary, the localization did not occur in some wide rifts (sensu Buck, 1991; Brun, 1999), like in the basin and range (e.g. Dickinson, 2002; and references therein): in this context, extension is distributed over wide areas, mainly because of the thickness and the thermal state of the lithosphere, without any further oceanization.

Less clear however is why and when such localization occurs, and what processes control it. Strain localization may be due to several processes. (1) It may occur because of significant strain softening (e.g. Huismans and Beaumont, 2007; Lavier and Manatschal, 2006), due to the feldspar–mica reaction (Bos and Spiers, 2002), pore pressure (Sibson, 1990), grain-size reduction, and/or mantle serpentinization (Pérez-Gussinyé and Reston, 2001).

(2) Specific rheological stratification of the lithosphere may also have a localizing effect: analog as well as numerical models have shown that 4 layer brittle–ductile models promote a strongly localized rifting while 2-layer models promote a rather distributed one (e.g. Brun, 1999; Gueydan et al., 2008; and references therein).

(3) In volcanic margins, the break up is accompanied by high amount of magmatism, underplated, intruded in the crust, and/or extruded at surface in Seaward Dipping Reflectors (SDR, e.g. Berndt et al., 2001; Planke et al., 2000), suggesting temperature anomaly in the asthenosphere

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(Holbrook et al., 2001). Such hot asthenosphere may weaken the lithosphere by injections of melts (Saunders et al., 1992; and references therein) and by dyking that allows stretching at low stress (Buck, 2004). It may also decrease the strength by heat conduction and thermal erosion at the base of the lithosphere (Saunders et al., 1992; White, 1992; and references therein).

(4) Finally, buoyancy forces arising from density variations in the lithosphere (Artyushkov, 1973; Fleitout and Froidevaux, 1982) may also provide driving forces for rift localization (Burov, 2007; Davis and Kusznir, 2002; Huismans et al., 2001). Crustal thickness variations are primordial, but the mantle lithosphere may also have slightly negative buoyancy (Griffin et al., 2009; Watremez et al., in press). In such case, a mantle lithosphere denser than the asthenosphere may help the localization. Such local stresses arising from density and thickness variations have been suggested to be active in the EARS (Zoback, 1992) and in the Gulf of Aden (Autin et al., 2010b) and may have produced stress rotations, because of oblique rifting context, that are actually observed in the field (Bellahsen et al., 2006). Buck (1991) and Buck et al. (1999) proposed that the mode of continental rifting was partly controlled by the ratio between crustal buoyancy and lithospheric necking effect and that the temperature controls this ratio. In the mantle lithosphere, it has been suggested that Rayleigh–Taylor instabilities can significantly thin the lithosphere (Conrad and Molnar, 1997; Gemmer and Houseman, 2007; Molnar and Houseman, 2004; Molnar et al., 1998). Similarly, Pascal and Cloetingh (2009) showed that gravitational potential stresses deriving from lateral variations in lithosphere structure at continental margins (South-Norway shelf) may explain features such as the seismicity pattern and the present-day stress orientations.

In orthogonal rifts, such interpretations may be supported by numerical models (Burov, 2007; Huismans et al., 2001). However, the force that causes faulting, whether originating from far-field tectonics or local buoyancy forces, cannot unambiguously be disentangled from the geological record as all faults have the same trend and their ages are uncertain due to the lack of precise absolute and relative dating.

Conversely, oblique rifts provide a unique opportunity to decipher the mechanisms at play and to establish their timing during rift localization as many fault populations develop, leading to complex fault patterns (see Autin et al., 2010b; Corti, 2009 and references therein) that includes both rift-parallel and rift-oblique faults. As suggested by Bellahsen et al. (2006), rift-parallel faults may witness rift localization processes. The mechanisms that promote oblique rifting are multiple and do not univocally include lithospheric structural inheritance or intrinsic mechanical behavior of the lithosphere (Autin et al., 2010b).

Among the youngest and best-documented oblique continental rifted margins are the eastern margins of the Gulf of Aden, whose ridge is the Arabia/Somalia plate boundary (Fig. 1). Based on the study of the Gulf opening and its structural evolution from rifting to active spreading, the aim of this contribution is to show that buoyancy forces have a feedback effect on rift dynamics. Moreover, we discuss the differences between the eastern and the western parts of the Gulf and the role of the Afar hot spot on the rifting geometry and evolution. We synthesize published and new onshore and offshore structural data and compare them to a synthesis of analog models. This analysis advocates for a structural evolution that is discussed in the light of numerical considerations of stresses arising from buoyancy forces.

2. Geological setting

2.1. The Afar hot spot

Seismic tomography reveals the presence of a body of low seismic velocities, in the west, underneath the Afar (e.g. Bastow et al., 2005; Benoit et al., 2006; Chang and Van der Lee, 2011; Hansen et al., 2012; Montelli et al., 2006). It ramifies to the Gulf of Aden, but does not go further east than, approximately, the Alula Fartak fault zone (Fig. 1) and, at shallow depths underneath the lithosphere, remains confined to the rift itself (Chang and Van der Lee, 2011; Sicilia et al., 2008). The low velocity body is not documented further east because of a very poor resolution due to the lack of receiver in eastern Yemen and southern Oman. This low seismic velocity anomaly certainly mirrors the source of the Afar hot spot, but imaged by Basuyau et al. (2010) and potentially laterally feeds the Aden spreading ridge up to the eastern Gulf of Aden (Leroy et al., 2010b). The hot spot has been active since 45 Ma (George et al, 1998), with pulses ca. 30 Ma (Hofmann et al, 1997) and is still active as attested by volcanic activity in the Afar area, as well as Quaternary volcanics along the Yemeni and Somali margins (Leroy et al., 2010b). This corresponds to a hotter asthenosphere that is readily inferred from the shallower bathymetry and higher heat flow in the western Gulf of Aden (Fig. 1) (e.g. Lucazeau et al., 2010; Rolandone et al., 2013 and references therein). Moreover, the sedimentary record shows that the western Gulf has been topographically high compared to the eastern part since at least Eocene times: siliciclastic deposits indicate that the western part was emerged and submitted to erosion while the eastern part was mainly the location of a carbonate-rich sedimentation (Leroy et al., 2012).

2.2. Continental rifting

Rifting started at about 34 Ma (Leroy et al., 2012; Roger et al., 1989; Watchorn et al., 1998) all along the Gulf. At this time, the subduction of Tethyan slabs underneath the Eurasian plates induced extension in the Afro-Arabian plate (Bellahsen et al., 2003). Such extensional strain field is due to the collision in the northern Arabian plate while the northeastern part was still subducting, this along-strike variation of the convergence mode being most probably the relevant boundary condition for intraplate extension. The extension was located in the Afro-Arabian rifts (Red Sea, Gulf of Aden and East African rifts) as the Afar hot spot activity localized the rifts. The combination of intraplate stresses with such a weakness produced oblique rifts, without any oblique preexisting lithospheric weakness (Autin et al., 2010b; Bellahsen et al., 2003, 2006).

The inherited Precambrian structures likely did not have a strong impact on the rift development (Fig. 1): none of the N–S, NE–SW, or NW–SE structures were reactivated in the Arabian plate. On the contrary, the structural pattern is strongly controlled by basins inherited from Cretaceous intraplate extensive event (see Birse et al., 1997; Brannan et al., 1997; Fantozzi, 1996; Leroy et al., 2012 for a synthesis; and Fig. 1). Those basins were partly reactivated during Tertiary rifting and trend from 90°E to 140°E (Fig. 1, Balhaf, Masila, Jiza-Qamar, Berbera, Nogal, Darror, and Gardafui basins). As a result of rift obliquity, the reactivated basins (and newly formed ones) are arranged en échelon along the continental margins (Fig. 2). Finally, the Indian Ocean and its ridge (Carlsberg, Fig. 1) most likely influenced the location of extensive deformation in the eastern most part of the Gulf of Aden (Manighetti et al., 1997).

At the Gulf scale, many recent studies showed that there were successive directions of extension during rifting ranging from 020°E to 160°E (Bellahsen et al., 2006; Fournier et al., 2004; Huchon and Khanbari, 2003; Lepvrier et al., 2002) (Fig. 2). It appears that extensional stresses most probably rotated counter-clockwise from 020° to 160°E, although it might be more complicated especially in distal parts of the margins (Autin et al., 2010b).

2.3. Ocean–continent transition

After the development of syn-rift grabens and horsts, the deformation localized where the crust was the thinnest, i.e. in distal margin grabens close to the future ocean–continent transition (OCT, Fig. 2, yellow).

In the eastern Gulf of Aden (i.e. east of Alula Fartak F.Z.), the OCT ridge may represent exhumed serpentinized mantle locally intruded by post-rift magmatic material, which modified the OCT after its emplacement (Autin et al., 2010a; d'Acremont et al., 2006, 2010; Leroy et al., 2010a; Watremez et al., 2011). The OCT segments are about 50 km long and oriented 110°E (Fig. 2). It is noteworthy that