Pre-existing oblique transfer zones and transfer/transform relationships in continental margins: New insights from the southeastern Gulf of Aden, Socotra Island, Yemen

N. Bellahsen a,b,*, S. Leroy a,b, J. Autin c, P. Razin d, E. d’Acremont a,b, H. Sloan e, R. Pik f, A. Ahmed g, K. Khanbari h

a UPMC Univ. Paris 06, UMR 7193, ISTeP, F-75005 Paris, France
b CNRS, UMR 7193, ISTeP, F-75005 Paris, France
c TVM, CNRS-UPS, Toulouse, France
d ENSIGE, INSA, Institut National des Sciences de l’Ingénieur et de l’Education, Génie Civil, Pôle Sciences et Technologies, France
e CNRS, UMR 7193, ISTeP, F-75005 Paris, France
f ENSCG-EOST, CNRS-Univ. Louis Pasteur, F-67084 Strasbourg, France
g CRPG-CNRS, BP20, 54501 Vandoeuvre-Lès-Nancy Cedex, France
h Yemen Remote Sensing and GIS Center, Sana’a University, Sana’a, Yemen

ARTICLE INFO

Article history:
Received 8 October 2012
Received in revised form 8 July 2013
Accepted 26 July 2013
Available online 9 August 2013

Keywords:
Transform fault
Transfer fault zone
Fracture zone
Segmentation
Structural inheritance
Oblique rifting

ABSTRACT

Transfer zones are ubiquitous features in continental rifts and margins, as are transform faults in oceanic lithosphere. Here, we present a structural study of the Hadibo Transfer Zone (HTZ), located in Socotra Island (Yemen) in the southeastern Gulf of Aden. There, we interpret this continental transfer fault zone to represent a reactivated pre-existing structure. Its trend is oblique to the direction of divergence and it has been active from the early up to the latest stages of rifting. One of the main oceanic fracture zones (FZ), the Hadibo-Sharbitath FZ, is aligned with and appears to be an extension of the HTZ and is probably genetically linked to it. Comparing this setting with observations from other Afro-Arabian rifts as well as with passive margins worldwide, it appears that many continental transfer zones are reactivated pre-existing structures, oblique to divergence. We therefore establish a classification system for oceanic FZ based upon their relationship with syn-rift structures. Type 1 FZ form at syn-rift structures and are late syn-rift to early syn-OCT. Type 2 FZ form during the OCT formation and Type 3 FZ form within the oceanic domain, after the oceanic spreading onset. The latter are controlled by far-field forces, magmatic processes, spreading rates, and oceanic crust rheology.

© 2013 Elsevier B.V. All rights reserved.

1. Introduction

Transfer zones and accommodation zones are ubiquitous features in continental rifts. These zones accommodate along strike structural changes within rifts (amount of extension, dip of tilted half grabens; e.g. Bellahsen and Daniel, 2005; Colletta et al., 1988; Corti et al., 2007; Jarrigue et al., 1990; Lezzar et al., 2002; Young et al., 2001) and have been extensively studied (e.g., for conceptual models, Bosworth, 1994; Rosendahl, 1987). In the literature, accommodation zones are defined as areas of diffuse strain transfer between two offset basins, while transfer zones are more localized fault systems. Acocella et al. (2005) suggested that accommodation zones are found in rifts and transfer zones in rifted continental margins. However, examples from the Gulf of Suez (e.g. Jarrigue et al., 1990; Moustafa, 1997) and the East African Rift System (e.g. Lezzar et al., 2002; Rosendahl, 1987) show that transfer zones were active during early stages of rifting. It is noteworthy that, in these two examples, the transfer zones are both pre-existing and oblique to the divergence. This may suggest that transfer zones do form during early rifting, only if they reactivated pre-existing structures.

Geometric consistencies between continental margin transfer zones and mid-ocean ridge transform faults suggest that transform faults originate at structures inherited from the rifting stage (Behn and Lin, 2000; Cochran and Martinez, 1988; Brehke, 2000; d’Acremont et al., 2005, 2006; McClay and Khalil, 1998; Watts and Stewart, 1998; Wilson, 1965). However, a significant number of studies suggest that transform faults are inherent to spreading processes and thus not related to continental margin segmentation (e.g., Taylor et al., 2009) but to thermal stress (Sandwell, 1986), and that they may evolve rapidly after their formation (e.g. d’Acremont et al., 2010). Their origin remains unclear as shown in recent reviews (Choi et al., 2008; Gerya, 2012). Analog models of transform-ridge systems (Freund and Merzer, 1976; O’Bryan et al., 1975; Oldenburg and Brune, 1972, 1975) showed that transform faults can nucleate spontaneously in accreting and cooling plates of wax. However, surprisingly, later experiments (e.g. Katz et al., 2005; Ragnarsson et al., 1996) were not as successful in reproducing transform faults (see Gerya, 2012 for more details and the complete
Numerical models confirmed that transform faults may indeed form spontaneously during oceanic spreading, even without initial ridge offset (Choi et al., 2008; Gerya, 2010, 2012 and references therein).

So far, few studies have attempted to determine, from onshore field data, whether or not transfer zones influence the location of oceanic transform faults. MacDonald et al. (1991) proposed a hierarchy for ridge discontinuities. From first order discontinuities (transform faults) to 2nd, 3rd, and 4th order (overlapping or connecting segments), this classification does not explicitly incorporate considerations about inheritance from syn-rift times. As noted above, the fact that transfer zones in rifts are often oblique to the divergence direction may be of importance. Indeed, in continental rifted margins, syn-rift faults that are parallel to the future transform faults, i.e. parallel to divergence, are almost never observed. These results led Taylor et al. (2009) to suggest that transform faults do not form at transfer zones. It may indeed be the case in some passive margins, as demonstrated by the numerous above-cited analog and numerical models. However, the observation that transform faults are rarely or never parallel to any transfer zones in the proximal margin domain does not necessarily imply that there is no genetic link between them. Two cases are possible. First, some transfer zones (parallel to divergence) can be imaged in distal margins and their relations with transform faults can be discussed. Of course, the imaging of such sub-vertical features is challenging and can be questioned, but several datasets show their existence especially in the distal margins (e.g. d’Acremont et al., 2005). Second, the fact that transfer zones are oblique does not imply that they are unlike precursors of transform faults as stated by Taylor et al. (2009). The question is, can we find any structures in the continental margin that determine the location and geometry of the transform faults? And, if so, can we base a classification for oceanic FZ on their relationships with continental transfer zones?

In this contribution, we present a study of the structural evolution of an onshore oblique transfer zone, based on a new geological map (Leroy et al., 2012; Razin et al., 2010). This feature is located in Socotra Island (in Yemen, Fig. 1), which constitutes a part of the southeastern margin of the Gulf of Aden. From structural and fault slip data, we detail the structural evolution of transfer zone, that continues offshore as an
oceanic FZ (Leroy et al., 2012) here named the Hadibo–Sharbithat FZ. We show that the transfer zone is composed of reactivated, pre-existing faults that have been activated as oblique-slip normal faults since early syn-rift times. We compare these results to Afro-Arabian rift systems (Fig. 1) to suggest that most of the transfer zones are actually pre-existing structures that are oblique to the divergence. Furthermore, we derive a classification for oceanic FZ based on their relationships with syn-rift structures.

2. Geological setting of Socotra Island, Yemen, Eastern Gulf of Aden

In the Gulf of Aden (Figs. 1 and 2), rifting commenced at around 34 Ma (Leroy et al., 2012; Pik et al., 2013-this volume; Robinet et al., 2013-this volume; Roger et al., 1989; Watchorn et al., 1998). At this time, the subduction of Tethyan slabs beneath the Eurasian plates induced extension in the Afro-Arabian plate, mainly because the northern Arabian plate entered the collision zone while the remainder of the subduction zone was still active (Bellahsen et al., 2003). The extension was located in the Afro-Arabian rifts (Red Sea, Gulf of Aden and East Africa), which were localized by the activity of the Afar hot spot. The combination of intraplate stresses with the Afar weak zone produced oblique rifts that did not involve significant reactivation of any major preexisting lithospheric weaknesses (with trend comparable with that of the future Gulf of Aden, Autin et al., 2010a, 2013-this volume; Bellahsen et al., 2003, 2006). The rifting pattern is characterized by Tertiary reactivation and formation of E–W to 140°E grabens (Fig. 2) inherited from a Cretaceous intraplate extensional event (see Birse et al., 1997; Brannan et al., 1997; Leroy et al., 2012). As a result of the rift obliquity, the reactivated grabens as well as newly formed ones are arranged en échelon along the continental margins. Many recent studies have shown that during rifting there were several successive directions of extension ranging from 020°E to 160°E (Bellahsen et al., 2006; Fournier et al., 2004; Huchon and Khanbari, 2003, 2006; Lepvrier et al., 2002). These extensional stresses most probably rotated counter-clockwise from 020°E to 160°E.

In the eastern Gulf of Aden, the Ocean–Continent Transition (OCT) ridge (Fig. 2) may represent exhumed serpentinitized mantle, which may or may not be intruded by post-rift mafic material that modified the OCT after its emplacement (Autin et al., 2010b; d’Acremont et al., 2006, 2010; Leroy et al., 2004, 2010a, 2010b; Watremez et al., 2011). The continental margin can be divided into several domains (Autin et al., 2010b; d’Acremont et al., 2005, 2006) with contrasting styles of rifting (Leroy et al., 2010a). Three major transform-generated FZ can be identified. They correspond to three major offsets of the coastline: from west to east, the Alula Fartak FZ, the Socotra–Hadbeen FZ, and the Hadibo–Sharbithat FZ. The latter is aligned with transverse structures cropping out on Socotra Island, the Hadibo Transfer Zone (HTZ, Figs. 2, 3, and 4). Between the two former FZ, transfer zones can be mapped out from seismic data (Fig. 2) in the distal continental margin.
and thus probably formed during the final stages of rifting, possibly during formation of the OCT (d’Acremont et al., 2005). These two FZ can be observed to be continuous with transfer zones (e.g. Salalah FZ, Fig. 2).

Oceanic spreading along the central part of the Aden–Sheba ridge commenced at 17.6 Ma (Fig. 2, d’Acremont et al., 2006, 2010; Leroy et al., 2012). In the westernmost part it started at 9 Ma (Audin et al., 2004; Leroy et al., 2012) and in the easternmost part, likely at 20 Ma (Fournier et al., 2010). From west to east, present-day spreading rates range from 13 mm/yr (azimuth 035°E) to 18 mm/yr (azimuth 025°E) (Jestin et al., 1994; Vigny et al., 2006). This opening direction demonstrates oblique divergence as the rift axis trend is about 075°E.

The following description of the stratigraphy is based on previous descriptions from South Arabia in Sultanate of Oman (Leroy et al., 2012; Razin et al., 2010; Robinet et al., 2013–this volume; Figs. 6 and 8 to 10). Pre-rift Formations (Fms.) are Cretaceous to Eocene and include Cenomanian limestones overlain by the Umm Er Radhuma platform carbonates (Thanetian to Ypresian) and the Rus peritidal platform carbonates (Ypresian). The middle to late Eocene is represented by the Dammam (Lutetian to Bartonian) and Ayardim (Priabonian) formations. Syn-rift series are composed of the Ashawq (platform, Rupelian) and Mughsayl (calcturbidites, Chattian) Fms. Locally, post-rift sedimentary rocks can be observed and consist of (late) Burdigalian to Langhian conglomerates (Ayaft Fm.).

3. Structural analysis of Socotra Island

Socotra Island is divided in two parts by the large HTZ (Figs. 3 and 4). The HTZ trends NE–SW and consists of several normal and oblique-slip faults. In the eastern part of the island, Mount Haggier and surrounding peaks comprise a large outcrop of basement rocks. The massif is bounded northward by two main north-dipping steep normal faults, one trending ENE–WSW and one trending NW–SE (Fig. 3). Other normal faults trending NW–SE, E–W, and 070°E have been mapped in the field. In the hanging-walls of these steep faults, low-angle normal faults (dip around 10–20°) affect the Mesozoic and Tertiary sedimentary layers, overlying the basement through tectonic contacts (Figs. 3, 5 and 6).

In the western part of the Island, the structural pattern is more complex. Three syn-rift basins (Allan–Kadarma, Sherubrub–Balan, and Central basins) formed during Oligo–Miocene times and few smaller ones formed along the transfer zone (Fig. 3). The two western basins are aligned along E–W to 110°E trending normal faults, while the central one is partly aligned along the southern part of the HTZ. The present study is focused on the northwestern and the central parts of the island that are described in detail in the following. We describe the structural setting along with fault slip data that provide an idea of the syn-rift stress states. The paleostress tensors have been calculated following the method of Angelier (1984). For the fault slip inversions, only small faults have been used (mm to dm scale displacements, Fig. 7). Balanced cross-sections are presented and based on the assumptions of constant area and length of sedimentary layers (Dalstrom, 1969).

3.1. The Allan–Kadarma basin

In its western part, the Allan–Kadarma basin is controlled by a low-angle normal fault segment that trends WNW–ESE, between the Qalansiya coast and the Jebel Kadarma (Figs. 5 and 8) and was active during deposition of the Mughsayl Fm. (Figs. 3 and 5, upper Oligocene–Miocene, Leroy et al., 2012). In the eastern part, south of the Jebel Allan, the basin is controlled by a steeper fault (dip around 60–70°) (Figs. 5 and 8). Between these two sub-basins, a NE–SW normal fault occurs, which is considered as a transfer fault (see below).

In the western part, the main normal fault is a low-angle segment trending WNW–ESE that was active under an extension around N–S to 020°E (sites 28, 25, 60, 271, Fig. 5). North of the main low-angle normal fault, in Jebel Kadarma, two fault trends can be mapped. A WNW–ESE south-dipping fault branches at depth from this low-angle segment (Fig. 8a). A segment trending 070°E was activated subsequently, as indicated by its abutment on the WNW–ESE fault segment. There, most of the fault slip data indicate a 020°E extension (sites 37, 141, 233, 234, Fig. 5). However, some fault-slip data suggest that there might have been a counterclockwise rotation of the extension, from N–S/020°E to around 160°E: the distribution of some fault slip data is not symmetrical (Fig. 4). Some stereonet analyses have a wide and asymmetrical distribution of fault slip (sites 37 and 234, but also 28, 265, 274, and 275 in other areas, Fig. 5). They differ from other sites such as sites 25, 326, 328, or 608 for example (Fig. 5), where the distribution is symmetrical around the perpendicular to the fault strike. When the distribution is
symmetrical, although wide, the local tectonic history can be considered as monophase (if no stria overprinting is observed on fault planes). Such wide distribution is indeed due to local fault interactions and/or fault plane irregularities. However, when the stria distribution is both wide and asymmetrical (sites 37 and 234, Fig. 5), it may be interpreted as indicative of stress rotations either counterclockwise (sites 37 and 234) or clockwise (site 274A, as well as site 275 in another area, Fig. 5, see caption for more details). The counterclockwise rotation (020°E then 160°E) of the least compressive stress is in accordance with the abutting of the 070°E fault against the WSW–ESE one (Jebel Kadarma, Fig. 5).

Finally, above the low-angle fault segment, an E–W extension is recorded (Fig. 5, sites 242B and 60B). Such extension is also witnessed by N–S faults that branch on the main low-angle fault segment, East of Qalansiya.

In the eastern part of the basin, the faults are steeper (Fig. 8b). Two main extension directions are observed, 160°E and 020°E, although intermediates are also found (sites 267, 273, 274, 267, 265, 263, 260, Fig. 5). At site 265 (Fig. 5), chronological relationships between striae (on the same fault plane), indicating a 020°E extension followed by a 160°E extension, have been observed in the field (see more examples in the next sub-section). Moreover, the analysis of some asymmetric fault slip distribution (see above) suggests that a counterclockwise rotation of the least compressive stress occurred (also at site 265, Fig. 5).

Between the two sub-basins, the NE–SW fault can be considered as a transfer fault as it separates two sub-domains with different geometries and amounts of extension (Fig. 8). We have little data from within the fault zone, but it might have been active during the entire period of rifting, particularly from early stages, to accommodate the along-strike variations in geometry and amount of extension. At site 274 (Fig. 5), two stress tensors are compatible with a N–S extension (one in an extensive regime, one in strike-slip regime). Such a stress permutation may be due to the proximity to the transfer zone (see similar interpretation in Fournier et al., 2007).

All the faults described above, as well as those described in the next sub-section, are considered to be Oligo–Miocene in age. All fault population types affect the Oligocene formations (see the

![Fig. 4](image-url) Hadibo Transfer Zone: (a) 3D Google Earth view (earth.google.com) and (b) photo. In the fault’s hanging wall, Mesozoic strata are sub-horizontal above the Paleozoic metasediments and basement. In the footwall, basement rocks crop out up to about 1500 m high (Mount Haggier). The Hadibo Transfer Zone presents both a right-lateral and a normal displacement (see Fig. 9).
location of the measurement site on Fig. 5). Faults observed in older formations (Cretaceous, Paleocene, and Eocene) display similar geometry and kinematics. We therefore consider that these are also Oligo-Miocene in age.

3.2. The Hadibo Transfer Zone

In the HTZ and surrounding area, normal faults trend NE-SW to 070°E and around 110°E (Figs. 3, 4 and 9). In the transfer zone sensu

![Geological map of the NW part of the Island with fault-slip data. Extension directions mainly range between 020°E and 160°E, with several exceptions around E-W. Cross sections of Fig. 8 are represented by the two thin lines. Few stereonets have a wide and asymmetrical distribution of fault slip: sites 28, 37, 265 and 274, 275. They differ from other sites such as 25, 326, 328 for example, where the distribution is symmetrical around the perpendicular to the fault strike. The asymmetry can be interpreted as witnessing stress rotations that were either clockwise (28, 37, 265) or counterclockwise (274, 275). We have drawn the perpendicular to the fault strike (dashed line). One may note that the fault striae are, in these cases, systematically lying on a line that was rotated either clockwise or counterclockwise. On these sites, the fault slip inversion has been performed, but manually drawn arrows for extension have been represented to suggest the polyphased history. Numbers 1 and 2 attest for probable first and second phases.]
stricto, the fault kinematics indicate that it was activated at some point of its history as a normal fault zone trending NE–SW to 070°E, with a sub-perpendicular extension, around 160°E (Fig. 9, sites 18A, 196, 259B, 367). There are also evidences of strike-slip events: ENE–WSW faults were active as right-lateral faults and record strike-slip regimes with a probable 020°E extension (Fig. 9, sites 18B, 258).

Indeed, faults oblique to the HTZ have been observed: there are faults trending E–W to 110°E with extension directions from N–S to 020° (Fig. 9, sites 15, 249, and 246) with syn-rift sedimentary rocks in their hanging wall. Relative chronologies between two sets of slickenlines on the same fault plane have been observed at few sites and provide us with a relative chronology between states of stress. At sites 4, 246A, 249B, and 259A (Fig. 9), striations on WNW–ESE normal faults display dip slip movements (extension around 020°E) followed by more strike-slip left lateral movement (extension around NNW–SSE), indicating a counterclockwise rotation of the extension. At sites 194 and 249B (and possibly at sites 246A and 259A, although more data is needed), the asymmetrical fault slip distribution (see above for details) can also be indicative of least compressive stress rotation. Moreover, at sites 138 (Fig. 9), we found chronologies between striations that suggest both clockwise and counterclockwise rotations. Finally, at sites 18 and 133 (Fig. 9), we found evidences of the opposite chronology on single fault planes.

From a structural point of view, at a larger scale, the overall geometry of the area suggests that the 110°E to NW–SE normal faults abut onto the NE–SW to ENE–WSW faults of the HTZ (Fig. 5 and 9). Moreover, the two domains east and west of the HTZ are very distinct and different. The western one presents many normal faults and associated syn-rift basins as described above. The eastern domain is almost undeformed, except in the northernmost part. It consists of a large, partly eroded monocline. In the northernmost part, NE–SW and NW–SE steep normal faults offset the monocline as well as low-angle faults now observed in the steep fault’s hanging wall (Figs 6, 9 and 10). The steep normal faults are thus considered to post-date the low-angle fault segment. Similar observation was made in Pik et al. (2013-this volume).

These observations (abutting relationships and different domains on each side of the transfer zone) suggest that the HTZ was active during early syn-rift times. Moreover, as it is strongly oblique to the global 020°E divergence, it may be a reactivated pre-existing structure.

4. Discussion
4.1. Structural evolution of Socotra Island
4.1.1. Pre-existing transfer zones
Based on our structural analysis, we propose that the HTZ is a reactivated pre-existing structure. The HTZ separates two very different domains and thus became active during the earliest stages of extension prior to other nearby faults that all abut on it. Moreover, the HTZ trends at angles of about 20°–50° to divergence, strongly suggesting that it was pre-existing and that it has been reactivated. Without tectonic inheritance, it is most likely that all extensional faults should trend sub-perpendicular to the divergence or the main directions of extension.

Another NE–SW-striking fault is also observed in the easternmost part of the island (Ras Momi, Fig. 3). The fault may be Mesozoic in age as it offsets the Permo–Triassic sedimentary rocks (600 m of throw). It is similar to faults observed in Huqf (north of Oman) and Sawqrah (southeast of Oman) regions (Leroy et al., 2012). These NE–SW faults thus have polyphased history and are prone to reactivation. Moreover, Proterozoic acidic and basaltic dykes in the basement at Socotra strike NE–SW (Denele et al., 2012). These dykes may very well have been reactivated as faults during the rifting.

Similar geometries and interpretations have been reported for the Suez Rift (Bosworth, 1994; Jarrigue et al., 1990; Moustafa, 1997) and

![Field photos of low-angle normal fault segments. a) East of Hadibo (Figs. 3 and 9). Eocene Umm Er Radhuma and Cretaceous limestone over low-angle normal faults, with basement rocks in the footwall. b) East of Qalansya (Figs. 3 and 5). Tertiary sedimentary rocks above low-angle normal fault with basement rocks in the footwall. The cliff is about 500 m high.](image-url)
East African Rift System (Corti et al., 2007; Rosendahl, 1987). In the Gulf of Suez, the Tertiary rift trends NW–SE to NNW–SSE with an extension direction around 060°E; transverse structures are described as transfer zones that separate asymmetrical rift segments (Bosworth and McClay, 2001; Bosworth et al., 2005; Jarrigue et al., 1990, and references therein). Three domains can be defined, the Darag, the October, and the Zeit domains (Younes and McClay, 2002), separated by the Zaafarana (ZAZ, NW–SE) and the Morgan (MAZ, N–S to NNE–SSW) accommodation zones, respectively (Fig. 12). These zones are oblique to the divergence direction and reactivated pre-existing basement lineaments (e.g. the Ribha shear zone for the ZAZ, Younes and McClay, 2002). At a smaller scale, numerous transfer zones trend around NNE–SSW. These faults are widespread within the rift and also oblique to the divergence.

The East African Rifts in Kenya and Tanzania (Fig. 13) formed during Miocene times (12–10 Ma, see Ebinger, 1989). Within that rift system, along-strike variability and alternating basin asymmetry are observed, and transfer zones are well documented (e.g. Corti et al., 2007; Morley et al., 1990; Rosendahl, 1987, and references therein) and trend NW–SE (Fig. 10). This pattern can be mapped at small and large scales. For example, the Rukwa rift (Morley et al., 1992) is a large transfer zone characterized by oblique-slip normal faults. At a smaller scale, within the Lake Tanganyika rift (Fig. 13), several (small) faults trending NW–SE (Fig. 10) provide example, the Rukwa rift (Morley et al., 1992) is a large transfer zone characterized by oblique-slip normal faults. At a smaller scale, within the Lake Tanganyika rift (Fig. 13), several (small) faults trending NW–SE (Fig. 10) are considered as transfer faults (Lezzar et al., 2002). Similarly, in the N–S Kenya rift (Fig. 13, Gregory rift), NW–SE to NNW–SSW oblique faults are pre-existing basement shear zones reactivated during rifting (Smith and Mosley, 1993). Similar observations were reported for the Malawi rift (Fig. 13) where pre-existing basement structures controlled the internal rift geometry (Ring, 1994).

The direction of divergence along the East African Rifts has long been debated (Fig. 13). Some authors conclude that the main extension is about NW–SE (Chorowicz, 2005; Sander and Rosendahl, 1989; Scott et al., 1992; Specht and Rosendahl, 1989; Versfelt and Rosendahl, 1989) while others conclude that the direction is E–W (Bosworth, 1992; Ebinger, 1989; Morley, 1989). Based upon analog and numerical models, Corti et al. (2007) suggested that it might be E–W. GPS data show that the present-day kinematics consists of E–W to WNW–ESE divergence between Nubia and Somalia (Calais et al., 2006; Déprez et al., 2013; Fernandez et al., 2004; Nocquet et al., 2006; Royer et al., 2006; Stamps et al., 2008). Thus, depending on the orientation of the divergence, the transfer zones might either be parallel or oblique to divergence. However, considering the present-day kinematics and also the probable Miocene kinematics, the transfer zones are oblique to divergence. Moreover, most of the transfer zones are strongly influenced by pre-existing older structures (see Fig. 13 and Rosendahl, 1987 for discussion).

To summarize, the Gulf of Aden, the Gulf of Suez and the East African Rifts have common characteristics among which is the geometry of the transfer zones: they trend oblique to both the divergence direction and the main normal faults (orthogonal to the divergence) and they are...
often basement structures reactivated as oblique-slip normal faults. In these rifts, it appears that there are no faults (or very few) strictly parallel to the divergence.

4.1.2. Early syn-rift depocentres

Early transfer zones usually control the location and geometry of early depocentres (e.g. Bellahsen and Daniel, 2005; Lezzar et al., 2002). However, looking at the basin and depocentre distribution on Socotra Island (Fig. 11), it appears that the main ones are located west of the HTZ. Moreover, the three main basins are located far away from this zone and the smaller ones are located at about 10 km westward, except the Central Basin (Figs. 3 and 11) that is probably partly controlled by the HTZ. Why is there no large syn-rift basin in the HTZ hanging wall as it seems active since early syn-rift times? Two answers may be envisaged: either the early syn-rift sediments were deposited and subsequently eroded or they were not deposited in this area that may have been relatively elevated since early rifting times.

There are late (syn-OCT, Burdigalian) detrital sedimentary rocks unconformably overlying the pre-rift, southwest of site 184 (Fig. 9). Thus, this area was already beginning at the end of OCT times. U/He thermochronology suggests that the main vertical offset along a 070°E steep normal faults (star symbol on Fig. 9, east of Hadibo) is rather late (around 18–20 Ma, syn-OCT, Pik et al., 2013-this volume). Finally, the syn-OCT to post-rift Miocene conglomerates (Burdigalian to Langhian) located north of the Mount Haggier, in the hanging wall of the normal faults east of Hadibo (star symbol, east of site 138, Fig. 9) contains many basement pebbles. No other detrital formation is observed. In any case, the amount of extension just west of the HTZ is very low (Fig. 10a), much lower than in the eastern part (Fig. 10b) and western part of the island (Fig. 8). Thus, the extension is accommodated elsewhere, most likely offshore, further north, where basins may also be found.

4.1.3. Polyphase tectonic history

The second main result is the structural evolution of the island and especially its early structural pattern (Fig. 11). We divide this evolution into three stages: (1) a first stage when the main WNW–ESE faults formed, along with the oblique reactivation of the NE–SW transfer zones during extension parallel to 020°E (Fig. 11). The faults with low-angle segments were active in both the western part (Kadarma and Sherurub basin, Figs. 5, 6, and 8) and the eastern part (east of Hadibo, Fig. 6, 9, and 10) of the island. The transfer zones separated blocks with different geometries and different amounts of extension. Thus, the island was already divided into three domains: the Kadarma–Sherubrub, the Central, and the Haggier (Fig. 11). The different amounts of extension (Figs. 5 and 8) accommodated in these three domains can be explained in two ways. Firstly, part of the extension in the central domain is not expressed at the surface, and some extension may, for example, be hidden beneath the coastal plain in the central-northern part of the island (Fig. 3). Small parts of syn-rift basins east of this plain can be observed, indicating that there may be normal faults hidden below. Secondly, and most likely, some of the extension is accommodated offshore, north of the island. In any case, we show here that tectonics evolve along strike, including the dip angle of the main normal faults and associated amounts of extension. During this stage, the transfer zones (especially the HTZ) are pre-existing structures reactivated as oblique-slip faults (right-lateral and normal), most probably with dominant strike-slip motion over dip-slip motion, given the angle between the extension (020°E) and the fault zone (~NE–SW).

(2) During a second stage, with increasing extension, 110°E-trending normal faults became more numerous, especially in the western part. During this stage, the extension may have rotated counterclockwise from 020°E to around 160°E resulting in the formation and reactivation of faults trending ENE–WSW to NE–SW. The transfer zone, striking NE–SW, was most probably also active as normal faults during this stage, as they become well-aligned at an angle of 65° to extension. Rift-parallel normal faults (trending 070°E) formed (for example in Jebel Kadarma, Fig. 5). The other normal faults (WSW–ESE) were still active and have an oblique-slip behavior. This stage probably did not last longer than few million years but is important as the onset of rift localization occurred during this time (Bellahsen et al., 2013-this volume). This stress rotation has been documented elsewhere in the Gulf (Bellahsen et al., 2006; Huchon and Khanbari, 2003; Lepvrier et al., 2002). Although, it cannot be unambiguously demonstrated by field data at Socotra Island, this is the most probable scenario at the scale of the entire Gulf as strongly suggested by field data and confirmed by analog models (Austin et al., 2010a, 2013-this volume) and numerical models (Austin et al., 2013; Bellahsen et al., 2013-this volume).

(3) During a third stage that led to the present-day fault network, the deformation migrated and localized, probably offshore to the north, followed by formation of the ocean-continent transition. The late syn-rift structure and the OCT emplacement may have been controlled by the far-field divergence (Bellahsen et al., 2013-this volume). Thermochronological data suggest that the HTZ was still active at this time (Pik et al., 2013-this volume).

4.1.4. Hadibo Transfer Zone/Hadibo–Sharbithat FZ continuity

The HTZ and the Hadibo–Sharbithat FZ continuity is supported by several observations and interpretations. (1) The FZ is aligned and continuous with the HTZ (Figs. 2 and 11). This alignment can be seen in the offset of oceanic magnetic anomalies offshore (Leroy et al., 2012). (2) In the distal margin between the HTZ and the Hadibo–Sharbithat FZ, a significant bathymetric escarpment defines two domains, suggesting the presence of a major structure (Fig. 3). This escarpment roughly links the transfer and the fracture zones, though more detailed bathymetry coverage and subsurface data are needed for a precise structural mapping. (3) U-Th/He thermochronology on apatites from basement rock in Socotra Island shows that the tectonic denudation, which started at the rifting onset, during late Eocene–early Oligocene time, continues during early Miocene (Aquitanian–Burdigalian) time (Pik et al., 2013-this volume). Thus, the HTZ was still active during the emplacement of the OCT, just before the formation of the oceanic transform fault. We therefore conclude that the HTZ and the Hadibo–Sharbithat FZ are structurally connected and that the Hadibo transform fault formed at the Hadibo Transfer Zone.

Fig. 9. Detailed map displaying the HTZ with fault-slip data. Extension directions mainly range between 020°E and 160°E, with several exceptions around E–W. The cross sections in Fig. 10 are represented by the two thin lines. Few stereonets have a wide and asymmetrical distribution of fault slip: sites 15, 246A, 249B. See caption of Fig. 5 for explanation and possible interpretation.
4.2. Axial segmentation in the Gulf of Aden and classification of oceanic fracture zone

In the Gulf of Aden, several transform faults and non-transform discontinuities offset the Aden–Sheba oceanic spreading ridge (Figs. 2 and 18). Here, we describe in detail the segmentation of the spreading ridge and present a new classification of the oceanic fracture zones. As they represent paleo-transform fault zones, the new classification will apply only to paleo-1st order transform fault zones (sensu MacDonald et al., 1991).

First, in the light of our results regarding the structural evolution on Socotra Island, we propose that a classification of oceanic fracture zones can be made based on the stage of rifting during which they or their precursors formed, and their relationships to the continental margin geometry and structures. In particular, we base our classification on whether the paleo-transform faults/FZ had offset and deformed either thinned continental crust and OCT, OCT only, or oceanic crust only. We also emphasize their relationship with pre-existing continental structures reactivated during rifting.

We name FZ's generated by transform faults that deformed continental crust Type 1 FZ. The Type 1 FZ offsets are usually (but not necessarily) large and result from a large offset between the two thinned continental domains that the proto-transform joined. Such geometry implies that part of the margin is a transform margin. Figs. 15 and 16 represent conceptual models of the transform fault zone and fracture zone evolution for orthogonal and oblique rifts, respectively. Three different Type 1 FZ are detailed below.

Fracture zones generated by transform faults that formed within the OCT during its formation (Figs. 15 and 16) are named Type 2 FZ. In the Gulf of Aden, Type 2 FZ are observed between Shukra El Sheik and Alula Fartak FZ (Figs. 15 and 16). These FZ offset the distal OCT boundary but do not offset its proximal boundary. Detailed studies of each FZ are required, but they clearly display geometries that differ from Type 1 FZ. They are more numerous in the central part of the Gulf because the orientation of the rift and OCT are oblique to the direction of divergence (Bellahsen et al., 2011-this volume). Some questions arise: What is their trend relative to the divergence? Are they reactivated pre-existing structures? Are they linked to any continental transfer zones? More data from distal margin is required to answer these questions, in particular high-resolution 3D seismic data.

We name FZ generated by transform faults that formed after the onset of oceanic spreading as Type 3 FZ or simply oceanic FZ. This type of FZ is not linked to any syn-rift structures and formed after rifting and OCT emplacement. Turcotte (1974) and Sandwell (1986) showed, on the basis of their spacing, that the oceanic transform faults are thermal contraction cracks, mainly due to the thermomechanical evolution of the oceanic lithosphere (Detrick et al., 1995; Gente et al., 1995; Hooft et al., 2000; and see Gerya, 2012, for a recent review). Type 3 FZ can be identified in the Gulf of Aden (Fig. 18a). They are indeed not correlated to any structure in the continental margin or the OCT and formed after the onset of oceanic spreading (see also d'Acremont et al., 2006, 2010).
Three cases of syn-rift Type 1 FZ can be distinguished, we describe them below.

4.2.1. Type 1-C (syn-rift) FZ

The first Type 1 FZ is named Type 1-C (C for continental, see below): such FZ formed as a transform fault that is linked to a blockage in along strike ridge propagation. In this case the transform fault is a "continental transform fault" that, at some point, separated an oceanic domain from a continental domain. Thus, it was located at the tip of a propagating spreading center (Figs. 15 and 16). Such a setting necessitates the presence of "continental" transform fault zone to accommodate the difference in the amount of divergence. This type of FZ can be found in several locations globally.

In the South Atlantic Ocean (Fig. 1), onset of seafloor spreading along the ridge between the Ascension and the Rio Grande FZ is younger than between the Rio Grande and the Falkland FZ (see Moulin et al., 2010). Thus, the Rio Grande FZ likely formed as a continental transform fault zone, which eventually corresponded with the northern tip of the oceanic spreading center. Such setting can last during a significant time only if there are continental transverse structures that accommodate the important differential opening (see reconstructions in Moulin et al., 2010; and references therein). The Rio Grande FZ is, for example, genetically related to continental transfer zones, namely the Pernambuco shear zone (Basile et al., 2005; Moulin et al., 2010) and Patos shear zone (Darros de Matos, 1999; Franqolin and Cobbold, 1994; Sénant and Popoff, 1991) in South America and the Sanaga and N’Gaoundere shear zones in West Africa (Basile et al., 2005; Moulin et al., 2010). For the Ascension and the Romanche FZ, the structural evolution may be similar although less straight forward (see a discussion in Basile et al., 2005 for Romanche FZ; compare Mohriak and Rosendahl (2003), Blaich et al. (2008), and Meyers et al. (1996) for Ascension FZ).

In the North Atlantic Ocean, similarly, the Senja–Greenland FZ (see Mosar et al., 2002 and references therein) separates two oceanic domains with different ages of spreading onset. In central Atlantic, the same is true for the Charlie-Gibbs, 15°20′N-Guinean, and Jacksonville-Cap Vert FZ (see Labails et al., 2010). The Owen FZ (easternmost Gulf of Aden, Fig. 1) is also most probably a Type 1-C FZ but with the notable difference that it likely formed within an oceanic domain (Mountain and Prell, 1990). These faults are parallel to divergence and their location and trend are probably (but not necessarily) determined by pre-existing structures. Indeed, as these faults are not particularly well oriented to be reactivated in an extensive stress field (because they trend parallel to the least compressive stress), it is more likely that they are reactivated pre-existing faults or shear zones. Finally, these transform
faults clearly form during rifting (at least for the continental domain situated on one side of the transform fault) and thus can be named syn-rift FZ.

In the Gulf of Aden, the Shukra El Sheik FZ is a Type 1-C FZ from the onset of spreading (17.6 Ma) and until 10 Ma, there was no spreading west of this fault (Leroy et al., 2012). Thus, motion on this fault activated because the westward ridge propagation was blocked there. This major fault extended southwestward into Afar where it was probably connected to the other active faults (Audin et al., 2004) (Fig. 18).

4.2.2. Type 1-T (syn-rift to syn-OCT) fracture zones

A second Type 1 FZ is named Type 1-T (T for transfer) when they formed at continental transfer zone, as in the Hadibo–Sharbithat FZ case (Figs. 2 and 11). In this case, a transform fault formed where a continental transfer zone was active during rifting (this study) and probably during OCT formation (Pik et al., 2013-this volume), leaving a fracture zone behind. The FZ and transfer zone have different trends (Figs. 11 and 17). In Figs. 15 and 16, the Type 1-T FZ are located in a zone that was acting as a transfer zone during rifting. The extension is accommodated by faults oblique to both the extension and the main rift-related faults. As shown above, such transfer zones are very often reactivated pre-existing faults. In the East African Rift System, the Rukwa rift zone (Figs. 13 and 15a) is an oblique rift zone where slip is occurring along old reactivated lineaments. If this zone eventually becomes an ocean basin, it is most likely that a Type 1-T FZ would form. In such case, one could not find any faults parallel to the divergence within the continental margin.

These Type 1-T FZ may form either during the latest rifting stages or during earliest OCT times. If the OCT proximal boundary is clearly offset, it probably means that transform fault formation is slightly older than the onset of OCT formation, thus late syn-rift (Fig. 17a, upper panel). If the continental transfer zone and the FZ join at the proximal boundary of the OCT, the formation of the transform fault zone may be dated to the age of the transition between syn-rift and syn-OCT (Fig. 17a, middle panel). If the offset of the OCT proximal boundary is controlled by the continental transfer zone (Fig. 17a, lower panel), it is most likely that the transform fault formed during OCT times. In the Hadibo–Sharbithat FZ, the OCT proximal boundary is clearly offset (Figs. 2 and 13, Leroy et al., 2012) and this offset suggests that the paleo-transform fault formed during syn-rift to syn-OCT times (compare Figs. 11 and 17a).

This provisional conclusion requires more data in the distal continental domains in order to be confirmed.
4.2.3 Type 1-P (syn-rift) fracture zones

On the Mid-Norwegian continental margin, a final Cretaceous–Tertiary rifting stage led to continental breakup and opening of the northeastern Atlantic Ocean during Early Eocene times (for recent studies see for example Brekke, 2000; Skogseid et al., 2000; Mosar et al., 2002). The long Jan Mayen FZ separates the margin into two basins: the Vøring basin in the north (Brekke, 2000) and the Møre basin in the south (Gabrielsen et al., 1999) (Fig. 14). This FZ is thought to continue onshore (Doré et al., 1997; Olesen et al., 2007) in the form of a Proterozoic lineament. The Norwegian continental margin is affected by numerous normal faults sub-perpendicular to the future transform fault, thus perpendicular to the Cretaceous–Tertiary divergence (see Mosar et al., 2002 for the various directions of divergence). However, in the accommodation zone, i.e. around the supposed trace of the Jan Mayen lineament, normal faults are strongly oblique to divergence (NW–SE) and strike N–S to NNE–SSW (Fig. 14). During rifting, the deformation and the depocentre locations across the Jan Mayen transfer zone clearly evolved from a diffuse accommodation zone (with some orientations that are not perpendicular to the divergence) to more localized transfer zone (with divergence-perpendicular faults, compare Figs. 7 and 13 in Brekke, 2000, for Cretaceous and Figs. 15 and 16 for early Tertiary times). This clearly indicates that an accommodation zone existed during rifting and that this zone evolved into a transfer/transform (see below) fault zone during the late rifting phase; i.e. when the amount of extension became large. Note that this is consistent with the review of Acocella et al. (2005) on transfer zones, where it is suggested that transfer zones form during late syn-rift phases, while accommodation zones are active during early syn-rift phases. However, here the late syn-rift structures that segment the margin are named

![Fig. 14. The Norwegian continental margin. Structural map simplified after Brekke (2000) and Olesen et al. (2007). The margin as well as a number of important normal faults trends NE–SW, except within the Jan Mayen accommodation zone, that are aligned along the Jan Mayen fracture zone. In this zone, the normal faults are oblique to both the divergence direction (Mosar et al., 2002, and references therein) and the future FZ.](image1)

![Fig. 15. Conceptual model of oceanic FZ in relation with rifting geometry, for orthogonal rifts. Three kinds of type 1 FZ are differentiated: Type 1-C when they derive from a “continental transform zone” that has been separating an oceanic basin from a continental one (as the Levant fault zone for example), Type 1-T when their location is influenced by an (often pre-existing) oblique continental transfer zone (re)activated at the onset of (or during) the rifting, and Type 1-P when their location is due to a pre-existing continental structure reactivated at the very end of the rifting. Type 2 FZ are defined as deforming only the OCT lithosphere: it means that the proximal boundary of the OCT is not offset, while the distal one is offset. Type 3 FZ initiate after the oceanic spreading onset and are related to evolving stresses within the oceanic crust.](image2)
transform fault (Fig. 17b), as it divergence parallel and represent a proto-plate boundary.

We name such faults syn-rift or Type 1-P FZ (P for pre-existing) when the transform fault location is influenced by a pre-existing fault parallel to the divergence (Figs. 15 and 16). This does not necessarily mean that such a transform fault was active in the proximal margin. Indeed, such a pre-existing fault may be reactivated only in the distal margin, where the continental crust is extremely thin within a narrow zone. This probably occurs during or slightly before OCT formation and/or mantle exhumation. The timing of paleo-transform fault zone formation depends on the structural relationships between the FZ and the proximal boundary of the OCT (Fig. 17b), similarly to the discussion about Type 1-T FZ (see 4.2.2). In Norway, the Jan Mayen FZ clearly strongly offset the proximal boundary of the OCT (Fig. 14) and is most likely related to a paleo-transform fault zone that formed during the latest stages of rifting.

In the Gulf of Aden, Alula Fartak FZ is a large-offset FZ. There is a main coastal offset and early syn-rift normal faults in Qamar basin are also offset by the FZ. There is no clear continuity of the FZ in the proximal domain (Fig. 2), near the Yemen/Oman border; this led Taylor et al. (2009) to propose that it nucleated during the onset of spreading or soon after. However, Autin et al. (2013-this volume) suggest that this FZ formed because of the presence of the pre-existing Jiza/Qamar-Gardafui Mesozoic basin that implied a stronger crust. There, the strain localization occurred on each side of the pre-existing basin (d’Acremont et al., 2005) as a consequence of rift obliquity; this setting necessitated a transform fault when the stretching localized during the late syn-rift/OCT. In this case, the FZ does not correspond to any continental transfer zone but its location is “inherited” from the rifted basin pattern. The FZ may also be a Type 1-P FZ, in our classification, as NNE-SSW pre-existing structures were and are present in the Arabian shield (Al-Husseini, 2000; Denele et al., 2012; Stern and Johnson, 2010).

Fig. 16. Conceptual model of oceanic FZ in relation with rifting geometry, in the specific case of oblique rifts. Same legend that is in Fig. 15. Three Type 1 FZ are differentiated: Type 1-C, Type 1-T, and Type 1-P. Type 2 and Type 3 FZ are also represented.
5. Conclusions

In this geological study of Socotra Island (Yemen), we have described in detail the structural evolution of an oblique transfer zone in a continental margin. This zone was active as an oblique-slip normal fault zone and is not parallel to the plate divergence, even if during its history, some pure strike-slip movements have occurred. We also show that the faults in this zone are pre-existing structures that were reactivated at the onset of rifting. Similar features can be observed in many extensional settings such as in the Gulf of Suez and the East African Rifts. On the basis of our findings, we propose a new classification of oceanic FZ based on their spatial-temporal relationships with syn-rift structures. We distinguish three distinct Type 1 FZ, i.e. oceanic FZ that are related to paleo-transform fault zones that deformed continental margins: Type 1-C that form at “continental transform fault zones” that are syn-rift in age, Type 1-T that form at continental transfer zones during late syn-rift to early syn-OCT times, and Type 1-P that form during late syn-rift to early syn-OCT at pre-existing structures that were not reactivated during early rifting. Type 2 FZ initiated during the OCT emplacement; Type 3 FZ form after the onset of seafloor spreading. Our classification requires the acquisition of more high-resolution data particularly in distal oceanic domains where transfer zones and fracture zones connect, to better constrain the geometry, the structural evolution, and the timing of transform fault zone formation.

Acknowledgments

The comments by C. Ebinger and an anonymous reviewer and the significant work of G. Peron-Pinvidic and P.T. Osmundsen as invited editors greatly improved the initial version of this contribution. This work benefited from financial support from GDR Marges, Action Marges (CNRS-INSU, TOTAL, BRGM, IFREMER), ANR YOCMAL and Rift2Ridge.
Fig. 18. Segmentation of the Gulf of Aden through time. a) The Gulf of Aden rift at around 20 Ma, at the time of OCT onset (purple line). Light orange color is for the pre-existing Mesozoic basins, bright orange represents the OCT. Type 1 FZ (in red) that initiated at that time are represented by: a Type 1-C FZ in the west (S. El S., Shukra El Sheik), a Type 1-T FZ in the east (S.H., Socotra Hadbeen and H.S. Hadibo–Sharbithat), and possibly a Type 1-P FZ in the center (A.F., Alula Fartak). b) The Gulf of Aden rift at the end of the OCT (yellow areas) formation at ~18 Ma. Type 2 FZ (syn-OCT, in blue). c) Present-day Gulf of Aden. Type 1 FZ are still active (except Shukra El Sheik). Some Type 2 FZ died out (in the eastern part). Type 3 FZ were initiated after the onset of oceanic spreading.
The authors thank Y. Denele for fruitful discussions. CSMRB is thanked for technical support.

References


