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ORIGINAL PAPER

From rifting to oceanic spreading in the Gulf of Aden: a synthesis

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Abstract We present here a synthesis of the evolution of rifted continental margin systems in the Gulf of Aden. These margins are volcanic to the west of the Gulf of Aden, where they are influenced by the Afar hotspot, and non-volcanic east of longitude 46° E. The combined use of magnetics, gravity, seismic reflection, field observations (tectonic, stratigraphic and sedimentological) and oil well data allowed us to obtain better constraints on the timing of continental rifting and seafloor spreading. From the Permo–Triassic to the Oligocene, the Arabian–African plate was subject to distributed extension, probably due, at least from the Cretaceous, to tensile stresses related to the subduction

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of the Tethysian slab in the north. In Late Eocene–Early Oligocene, 34–33 Ma ago, rifting started to localise along the future area of continental breakup. Initially guided by the inherited basins, continental rifting then occurred synchronously over the entire gulf before becoming localised on the northern and southern borders of the inherited grabens, in the direction of the Afar hotspot. In the areas with non-volcanic margins (in the east), the faults marking the end of rifting trend parallel to the inherited grabens. Only the transfer faults cross-cut the inherited grabens, and some of these faults later developed into transform faults. The most important of these transform faults follow a Precambrian trend. Volcanic margins were formed in the west of the Gulf, up to the Guban graben in the southeast and as far as the southern boundary of the Bahlaf graben in the northeast. Seaward dipping reflectors can be observed on many oil industry seismic profiles. The influence of the hotspot during rifting was concentrated on the western part of the gulf. Therefore, it seems that the western domain was uplifted and eroded at the onset of rifting, while the eastern domain was characterised by more continuous sedimentation. The phase of distributed deformation was followed by a phase of strain localisation during the final rifting stage, just before formation of the Ocean–Continent Transition (OCT), in the most distal graben (DIM graben). About 20 Ma ago, at the time of the continental break-up, the emplacement of the OCT started in the east with exhumation of the subcontinental mantle. Farther west, the system was heated up by the strong influence of the Afar hotspot, which led to breakup with much less extension. In the Gulf of Aden (s.str), up to the Shukra El Sheik fracture zone, oceanic spreading started 17.6 Ma ago. West of this fracture zone, oceanic accretion started 10 Ma ago, and 2 Ma ago in the Gulf of Tadjoura. Post-rift deformation of the eastern margins of the Gulf of Aden can be seen in the distal and proximal domains. Indeed, the substantial post-rift uplift of these margins could be associated with either the continental break-up, or activity of the Afar hotspot and related volcanic/magmatic activity. Uplift of the northern proximal margin was still active (e.g. stepped beach rocks exposed at 60 m of 2 Ma; 30 m of 35,200 years; 10 and 2 m) and active volcanoes can be inferred at depths of between 70 and 200 km beneath the margin (at 5–10 km distance from the coast). On the distal margin, heat flow measurements show a high value that is associated with post-rift volcanic activity and the development of a volcano (with flows and sills) shortly after the formation of the OCT. The Afar hotspot is therefore important for several reasons. It allows the localisation of deformation along the Red Sea/Aden system and the rapid opening of the Gulf after the continental break-up; its influence also seems to persist during the post-rift period.

Keywords Gulf of Aden · Continental margins · Oblique rifting · Continental break-up · Ocean Continent transition · Oceanic spreading · Segmentation · Afar plume · Inheritance

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Introduction

The study of continental margins in varied geological settings has revealed considerable lithospheric and morphological differences. On some margins, volcanic activity during rifting is very limited and the mantle can sometimes crop out at the surface (e.g. Lavier and Manatschal 2006). On so-called "volcanic margins", a great variety of volcanic products can be found that are emitted during extension (e.g. White and McKenzie 1989). The existence of persistent volcanic and magmatic domains far away from any ridge and at different stages of margin evolution also raise the issue of mantle behaviour from the initial rifting to the late stages of margin history. Hotspots could play a role in thermal weakening of the lithosphere, as well as by initiating future rifts, and this must be taken into account for localising the break-up (Courtillot et al. 1999). Similarly, it is possible that lithospheric-scale structures, such as suture zones inherited from earlier tectonic episodes, guided the emergence of rifts and future oceans. Thus, the superficial and external expression of continental margins is resulting from complex geodynamic processes originating in the mantle.

Ongoing studies are focused on the mechanisms of extreme lithospheric stretching and thinning, on the reactivation of pre-existing tectonic features, on the evolution of marginal deformation and thermal regimes during and after rifting, on the part played by magmatic processes, and by the mantle and lower crust (Hopper and Buck 1996; Muntener and Hermann 2001). In addition, interactions between climate, erosion, sedimentation and the dynamics of vertical movement on these margins are also being investigated at various time scales, using seismic imaging to carry out a detailed analysis of the observed sedimentary architectures.

The largest gas hydrate traps (Borowski et al. 1999) are identified in this type of margin setting, in areas of fluid seepage associated with living organisms (Aharon 1994). Rifted margins appear key areas for understanding carbon sinks. The sources of fluids, sedimentary fluxes, structures controlling fluid migration and seepage, all remain to be explored and parameterised before they can be integrated into global mass balances.

Three key issues concerning lithospheric extension are currently subject of debate: (1) the mechanisms involved in extreme distension and the way in which the crust thins down to just a few kilometre in the absence of major normal faults, (2) the thermal structure of the stretched lithosphere, (3) the initiation and degree of melting and its mechanical and rheological impact combined with lithospheric extension, which lead to oceanic opening. These questions extend far beyond the current state of knowledge on continental margins. One of the ways to answer these questions is by combining several methods and approaches to improve the acquired dataset, the geological interpretations and their modelling.

Continental rifting in the Red Sea and the Gulf of Aden area was a long-term process that started in the Permo-Trias and ended with the separation of the African and Arabian plates (Fig. 1, e.g. Bosworth et al. 2005). With its extensional system affecting the intracratonic basement, the Gulf of Aden (Fig. 1) is undoubtedly one of the best places to improve our understanding of continental rifting and ocean crust formation. The relatively young rift (<34 Ma) was followed by the emplacement of an active spreading ridge (17.6 Ma). The conjugate margins are easy to correlate and the pre-rift sediments, the basement and syn-rift sediments as well as the geometry of the associated faults can be visualised and sampled locally. The crustal structure of the entire system and its ocean–continent transition (OCT), as well as the oceanic accretion, can be visualised on a kilometric scale; the kinematics can be finely resolved and,

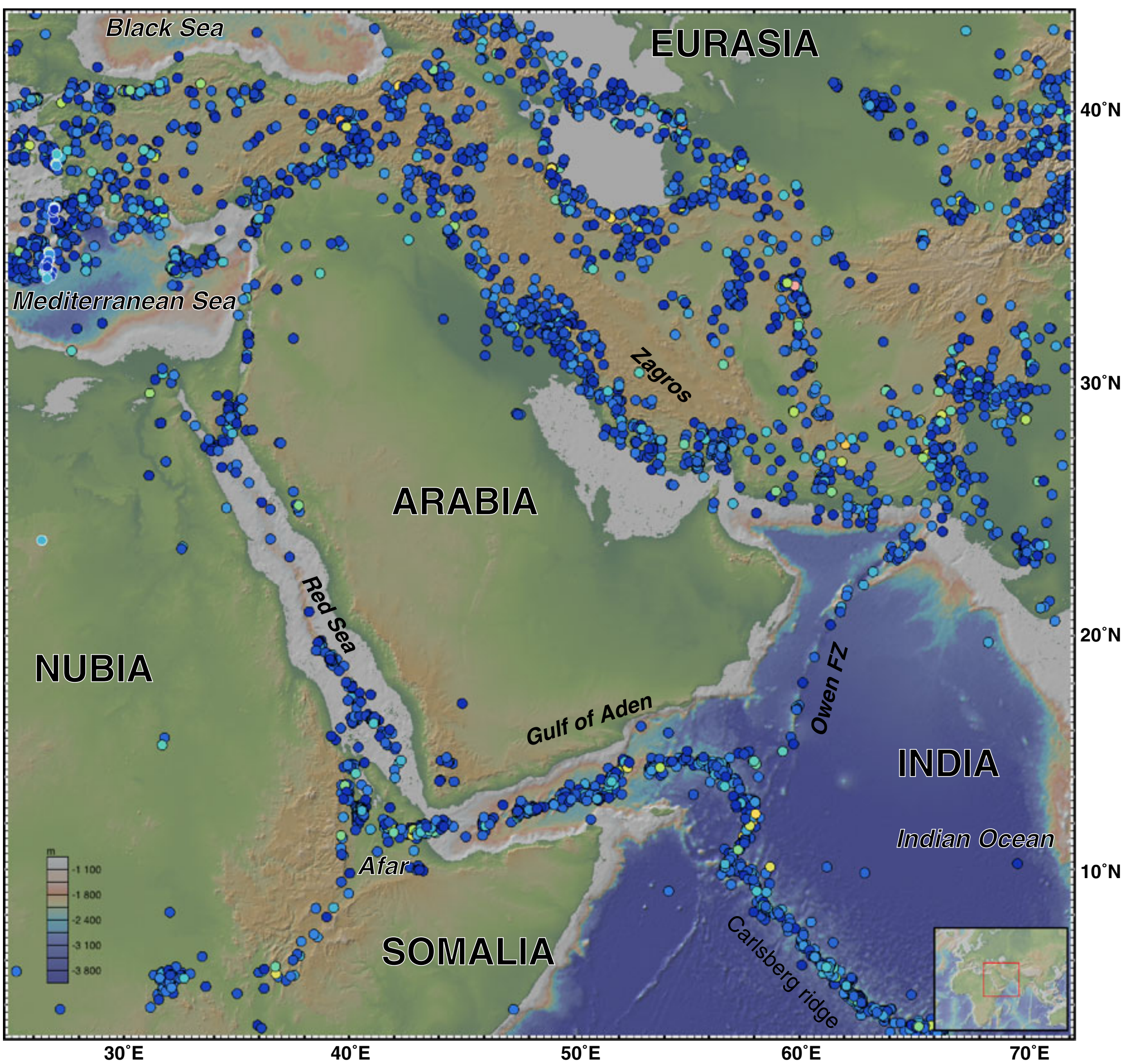


Fig. 1 Seismicity of the Arabian plate and surrounding regions. © GeoMapApp

finally, the extensional processes are not linked (directly or indirectly) with subduction systems like most other examples of young rift systems that are related to continental breakup (e.g. Gulf of California, Woodlark basin, Western Mediterranean, etc.).

The findings concerning lithospheric extension, the formation and evolution of the margins of the eastern Gulf of Aden result from several approaches: firstly, we investigate the inherited tectonic, stratigraphic, sedimentary and magmatic framework; and secondly, we explore the spatial distribution of faults and crustal/lithospheric structures of the margins. Finally, we make use of analogue modelling to study oblique rifting. In this paper, we present the sedimentary cycles. Then, we identify the location of the OCT and the segmentation, on both crustal and lithospheric scale, along the entire margin and Sheba ridge. We describe initiation of the spreading ridge and the development of volcanic activity, ridge jumps and oceanic spreading. All these features provide us with information on oceanic accretion processes in relation to plate kinematics. New magnetic and seismic data have allowed us to establish the history of rifting and oceanic spreading in the Gulf of Aden. These data indicate a synchronous opening of the Gulf of Aden, which started over 35 Ma ago, between the fracture zones of Shukra el Sheik to the west and Momi-Madrakah to the east (Fig. 2).

Geodynamic evolution of the Arabian plate

Large-scale geodynamic history of the study area: Late Permian to Maastrichtian

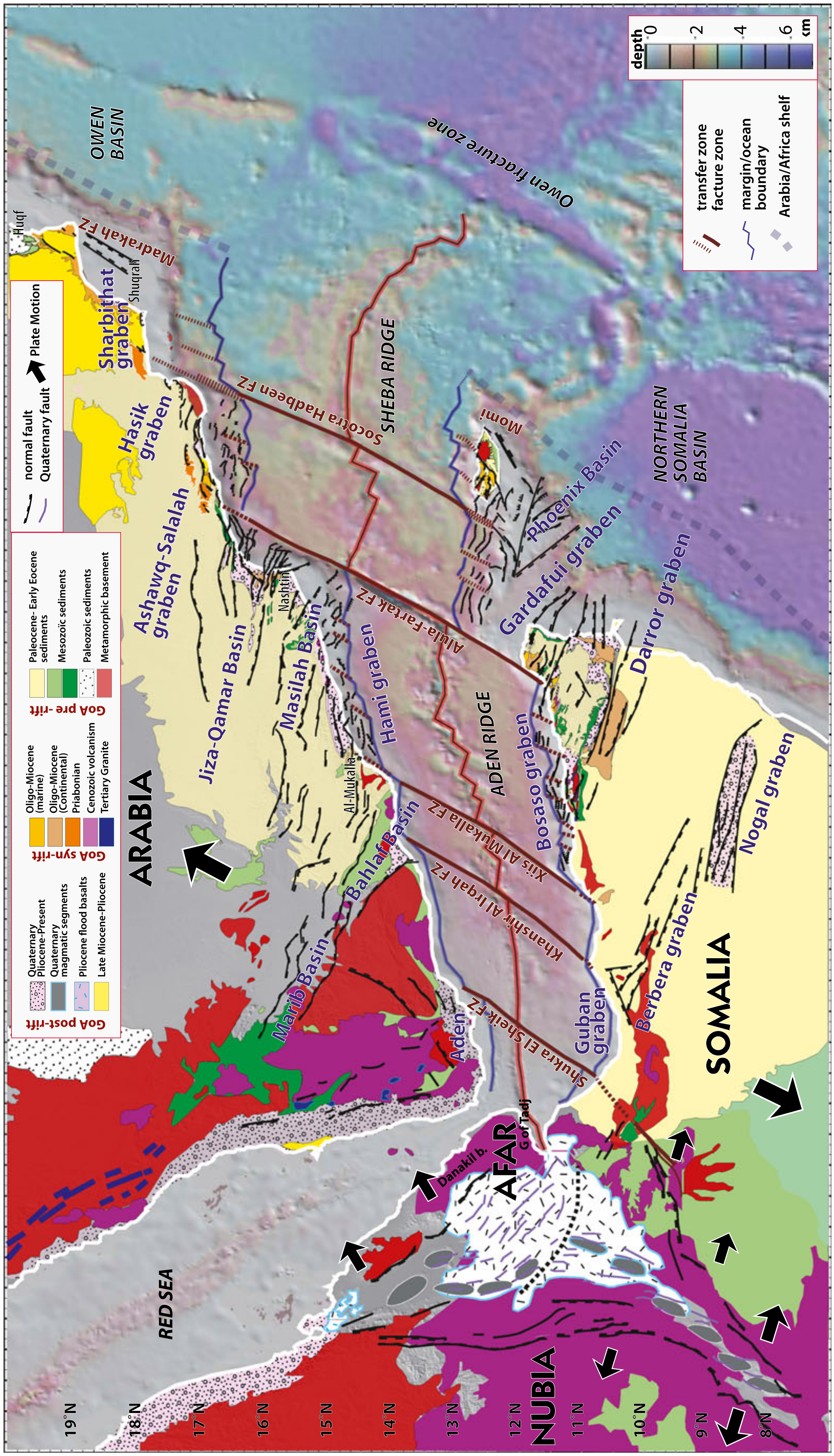
The local geodynamic history extends from the Late Permian to the Maastrichtian (Fig. 3) and includes the following stages: (1) opening of the North Somalian basin started with the M22 magnetic anomaly (150 Ma ago) inducing the split of the Madagascar–Seychelles–India block from the African–Arabian block (e.g. Marquer et al. 1998); (2) formation, 118 Ma ago, of a sinistral transform fault between Madagascar and the Seychelles. This structure possibly extends along the eastern Oman margin, thus accommodating the rapid motion of India to the north (Patriat and Achache 1984); (3) oceanic accretion between Madagascar and the Seychelles–India block starting 86 Ma ago (e.g. Dyment 1998); (4) initiation of oceanic accretion between the Seychelles and India, 65.5 Ma ago, and the possible collision between India and the Karman block led to the obduction of the Masirah (m in Fig. 3) ophiolites in the Maastrichtian–Palaeocene (Chaubey et al. 2002). Radiolaria overlying the oceanic crust on the Masirah Island has been dated late Jurassic/Early Cretaceous (Le Métour et al. 1995). (5) India–Eurasia collision started 53 Ma ago, implying a significant slowed down of the northward drift. At this time, the Owen transform fault was most probably located along the coast of Oman (Royer 2002).

Tectonic activity before Gulf of Aden opening: Permo–Triassic to Burdigalian

The history of rift/graben activity in the area has been determined on the basis of unpublished oil industry data and published maps (Wolde and Anonymous 1987; Scotese et al. 1988; Ellis et al. 1996; Beydoun 1998; Bunter et al. 1998; Ziegler 2001). The data synthesis presented here demonstrates a diachronous activity of these rifts and grabens as well as space variation.

- In the area of Socotra, Rukh, Binna, Nogal and the northwest of Madagascar Karoo rifting was active during the Permo–Triassic (Karoo), and NE–SW faults are responsible for the segmentation of the Marib graben in Yemen (Fig. 4).
- The Marib and Masilah rifts in Yemen and the Berbera rift in Somalia were active during the Late Jurassic/basal Cretaceous in a marine setting. At this time, the area corresponding to the future Gulf of Aden was topographically elevated (Fig. 4).
- The tectonic activity was distributed during the Cretaceous throughout the entire region (Fig. 4). In eastern Yemen, the Jiza-Qamar rift was strongly active from the Barremian up to the Late Cretaceous, both inland and offshore. To the south, the Bahlaf (Yemen), Darror, Berbera and Nogal (Somalia) grabens outline the tectonic and sedimentary activity, as in the grabens of Aban, in Central Arabia (Hancock et al. 1984), in Sirhan and Euphrates in the north (Litak et al. 1997; Brew et al. 2001). There is a major discontinuity at the base of Barremian, with truncation of previous successions (Fig. 5).
- The inception of the Afar hotspot has been recorded 45 Ma ago (age of the first lava flows; George et al. (1998)), and developed coeval with the formation of the Carlsberg Ridge (Fig. 3; Royer 2002).
- In the Gulf of Aden, an oblique rifting started during the Late Eocene/Early Oligocene (34–33 Ma; Roger et al. 1989; Watchorn et al. 1998; Razin et al. 2010). The rifting process developed synchronously in the entire gulf. The variability of the sedimentary facies expresses

Fig. 2 General sketch map of the Gulf of Aden, drawn up using oil company data, data from the ENCENS-Sheba cruise (Leroy et al. 2004; d'Acremont et al. 2005; 2006; Bellahsen et al. 2006), the Encens cruise (Leroy et al. 2010b; Autin et al. 2010b; d'Acremont et al. 2010), and from studies carried out in Afar (Ebinger et al. 2008); onland and offshore geology and structural sketch map on worldwide bathymetry base. *G of Tadj* Gulf of Tadjoura ▶



ARABIA
SOMALIA
NUBIA
AFAR
RED SEA
OWEN BASIN
NORTHERN SOMALIA BASIN
SHEBA RIDGE
ADEN RIDGE
Owen fracture zone
Madrakah FZ
Socotra Hadbeen FZ
Alula-Fartak FZ
Al Mukalla FZ
Khanshiir Al Irqah FZ
Shukra El Sheik FZ
Sharbithat graben
Hasik graben
Ashawq-Salalah graben
Jiza-Qamar Basin
Masilah Basin
Hami graben
Bahlaf Basin
Marib Basin
Phoenix Basin
Gardafui graben
Darror graben
Bosaso graben
Nogal graben
Berbera graben
Guban graben
Momi
Huqf
Shuqrah
Nashtin
Al-Mukalla
Aden
Danakil b.
G. of Tadj
GoA post-rift
Quaternary Pliocene-Present
Quaternary magmatic segments
Pliocene flood basalts
Late Miocene-Pliocene
GoA syn-rift
Oligo-Miocene (marine)
Oligo-Miocene (Continental)
Priabonian
Cenozoic volcanism
Tertiary Granite
GoA pre-rift
Paleocene- Early Eocene sediments
Mesozoic sediments
Paleozoic sediments
Metamorphic basement
normal fault
Quaternary fault
Plate Motion
transfer zone facture zone
margin/ocean boundary
Arabia/Africa shelf
depth
0
2
4
6
km
38°E
39°E
40°E
41°E
42°E
43°E
44°E
45°E
46°E
47°E
48°E
49°E
50°E
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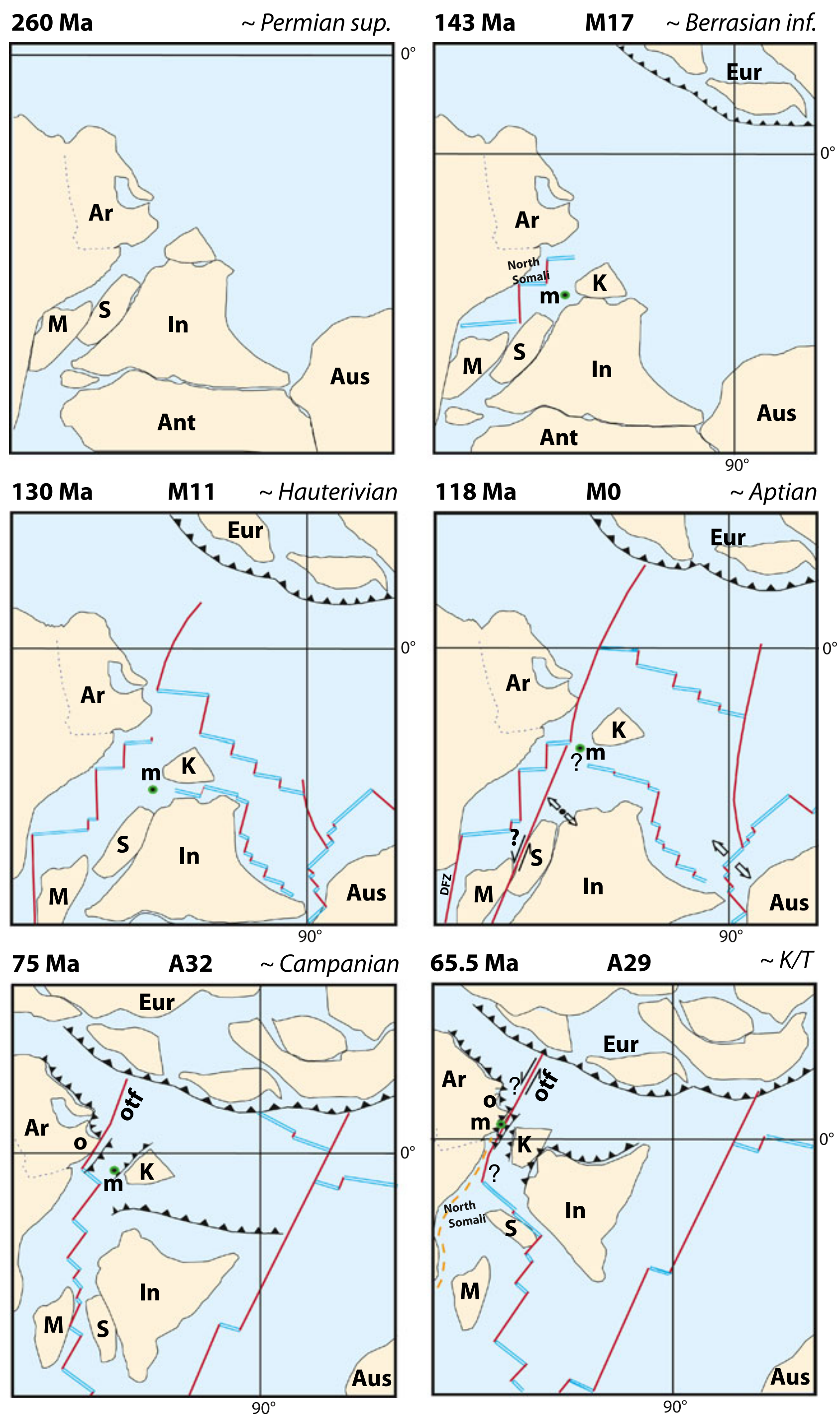


Fig. 3 Plate reconstruction and evolution of the western part of the Indian Ocean between 260 and 66 Ma. Modified after Le Métour et al. (1995), Gnos et al. (1997), Marquer et al. (1998), Edwards et al. (2000), Immenhauser et al. (2000). *Black dot circled in green* Masirah ophiolites (m), *Eur* Eurasia, *K* Karman block, *S* Seychelles, *M* Madagascar, *In* India, *Aus* Australia, *Ar* future Arabia, *o* Oman, *otf* Owen transform fault

a complex pattern of structural evolution along the Gulf (Fig. 2).

- In the Red Sea, the orthogonal rift started in its central (Jeddah; Hughes and Filatoff 1995) and southern parts (Margin of Eritrea and Sudan; Bunter and Magid 1989; Bunter et al. 1998) during the Early Oligocene.
- The major activity of the Afar hotspot is dated 30±1 Ma ago (Early Oligocene; Hofmann et al. 1997), as well as the initiation of rifting in the part of the Red sea located north of Saudi Arabia (Purser and Hotzl 1988) and in the southern Gulf of Suez (El-Shinnawi 1975). Volcanic activity of Afar hot spot induced the development of volcanic rifted margins in the western part of the Gulf of Aden (Tard et al. 1991). In the eastern part the margins are non-volcanic (Leroy et al. 2004, 2010b).
- Rifting was extremely active in the Gulf of Aden during the Early Miocene (Platel and Roger 1989; Fig. 4) and developed during the inception of the collision between

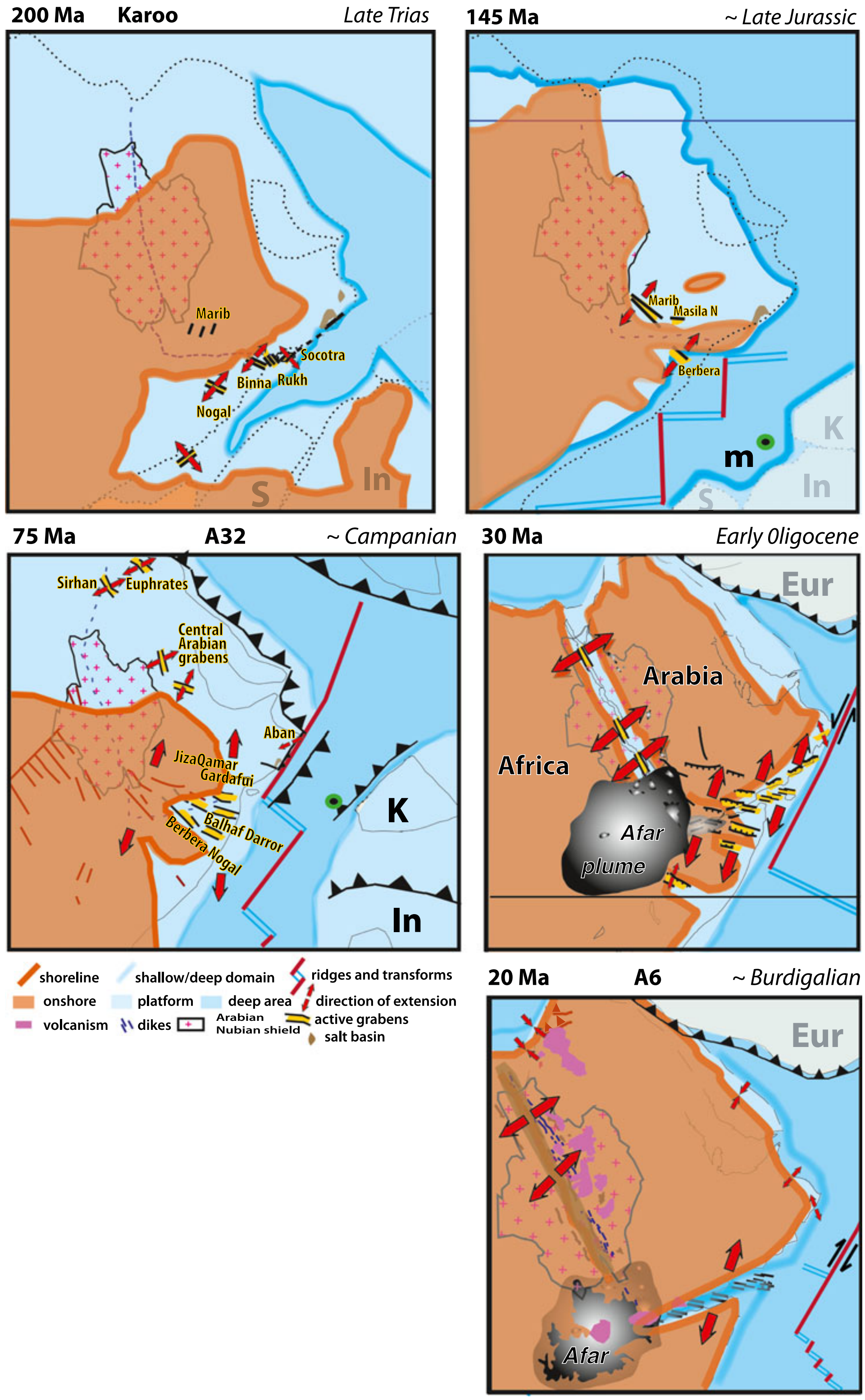


Fig. 4 Schematic paleogeographic maps of the Gulf of Aden region, from the Karoo (Permo–Triassic) to the Burdigalian. Drawn up after the literature cited in this article and using unpublished oil data as well as data from studies undertaken in the Gulf of Aden (e.g. Leroy et al. 2010a)

Arabia and Eurasia (25 Ma ago) in Zagros (e.g. Agard et al. 2005).

- In the easternmost part of the Gulf of Aden, the onset of oceanic opening is dated at 20 Ma (Fournier et al. 2010), occurred within an old oceanic lithosphere probably of Early Cretaceous age. This oceanic lithosphere fragment correlates in age with the oceanic crust of the Ras Madrakah and Masirah ophiolites (Platel et

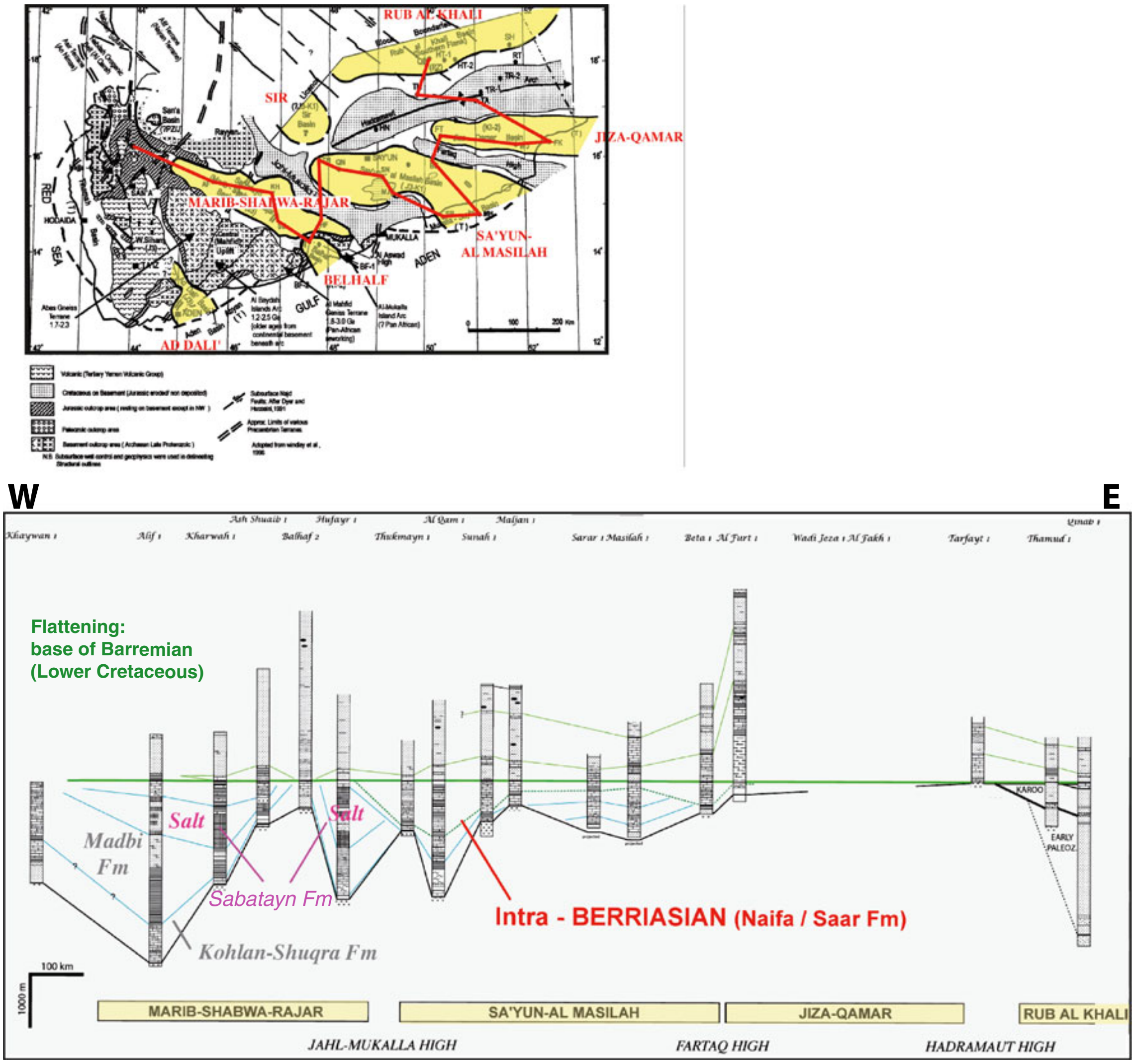


Fig. 5 Correlations between well data from the Yemen basins (data extracted from Bott et al. 1992)

al. 1992a). At the same period, in the eastern Gulf of Aden the oceanic spreading is not yet active but exhumation of the subcontinental mantle occurs (east of the Alula–Fartak fracture zone (d'Acremont et al. 2006; Leroy 2010a, b)).

Continental rifting in the Gulf of Aden

Since Late Eocene, the Gulf of Aden represents a good example of oblique rifting (Figs. 1, 2 and 4), relative to N30° E-trending divergence (e.g. Bellahsen et al. 2006; Autin et al. 2010b). As a result of rifting obliquity, the gulf is segmented, so that it can be divided into three main study areas (Figs. 2 and 6). The western area extends from the active rift of the Gulf of Tadjoura to the Shukra-el-Sheik transform fault (Fig. 6). The central area is comprised between this transform fault and the Alula–Fartak transform fault. The eastern part extends from the Alula–Fartak transform to the eastern border of the African craton (Fig. 6). These three zones show very different structural features that are described briefly below.

Western part of the Gulf (area I)

The extension is always directed ~N37° E (Dauteuil et al. 2001) whereas the trend of the Gulf is changing from west to east. The Gulf of Tadjoura is trending N70° E and the gulf has a E–W direction between longitudes 45°30 E and 43°30 E in the west of the Shukra El Sheik transform fault (Figs. 2 and 6). The obliquity of the rift (relative to the direction of extension) thus ranges from 39° to 59°, which explains the presence of different fault patterns along the axis of this zone. When the obliquity is low (39°), normal faults are sigmoidal with their axis striking N80° E to N140° E, which is purely extensional (i.e. perpendicular to the extension direction), while the extremities have a dextral strike slip component parallel to the walls of the axial valley. In cases where the obliquity is greater (59°), the deformed area is narrower and more subsided and bounded by faults with strike slip and normal components (Dauteuil et al. 2001).

In the Ba'tays area (southwest of Yemen; Fig. 6), normal faults strike parallel to the gulf (N60–80° E), with the tilted blocks on each side being crosscut by dextral N140° E strike-slip faults (Thoué et al. 1997; Khanbari and Huchon 2010). Huchon et al. (1991) suggested the existence of an E–W extension during the Oligocene and up to 22 Ma ago. Between 22 and 18 Ma ago, extension took place in two successive phases: firstly trending N20° E to N40° E and then N160° E to N–S (Fig. 6).

The central part of the Gulf (area II)

In Somalia, the Berbera, Darror, Gardafui (Fantozzi 1996; Fantozzi and Sgavetti 1998) and the Nogal (Granath 2001) basins are active. The axis of all these segments follow a N110° E–N120° E trend, with the exception of few faults striking N45°E (Fig. 6).

In Yemen, the axial trends of the Balhaf, Masilah and Jiza Qamar basins range from N130° E, N110° E to N90° E. The N20° E-trending extension phase appears to have preceded the N160° E phase (Khanbari 2000; Huchon and Khanbari 2003). The N20° E trend is that associated with the extension direction of the Gulf opening. The N75° striking normal faults are parallel to the Gulf axis. The N160° E trend, being orientated perpendicular to the axis of the Gulf, is suggested corresponding to the displacement between Africa and Arabia (Fig. 6).

The eastern part (area III): Oman and Socotra Island

In the region of Dhofar in Oman (Figs. 6 and 7a), extension follows the same N20° E and N160° E directions (Lepvrier et al. 2002; Fournier et al. 2004; Bellahsen et al. 2006). Although Fournier et al. (2004) were unable to establish the chronology between these, they suggested that the N75° E-striking faults developed along the zone of weakness responsible for the overall orientation of the Gulf, after which the N110° E-striking faults would have formed in response to the continuous extension occurring between the African and Arabian plates. However, analogue modelling

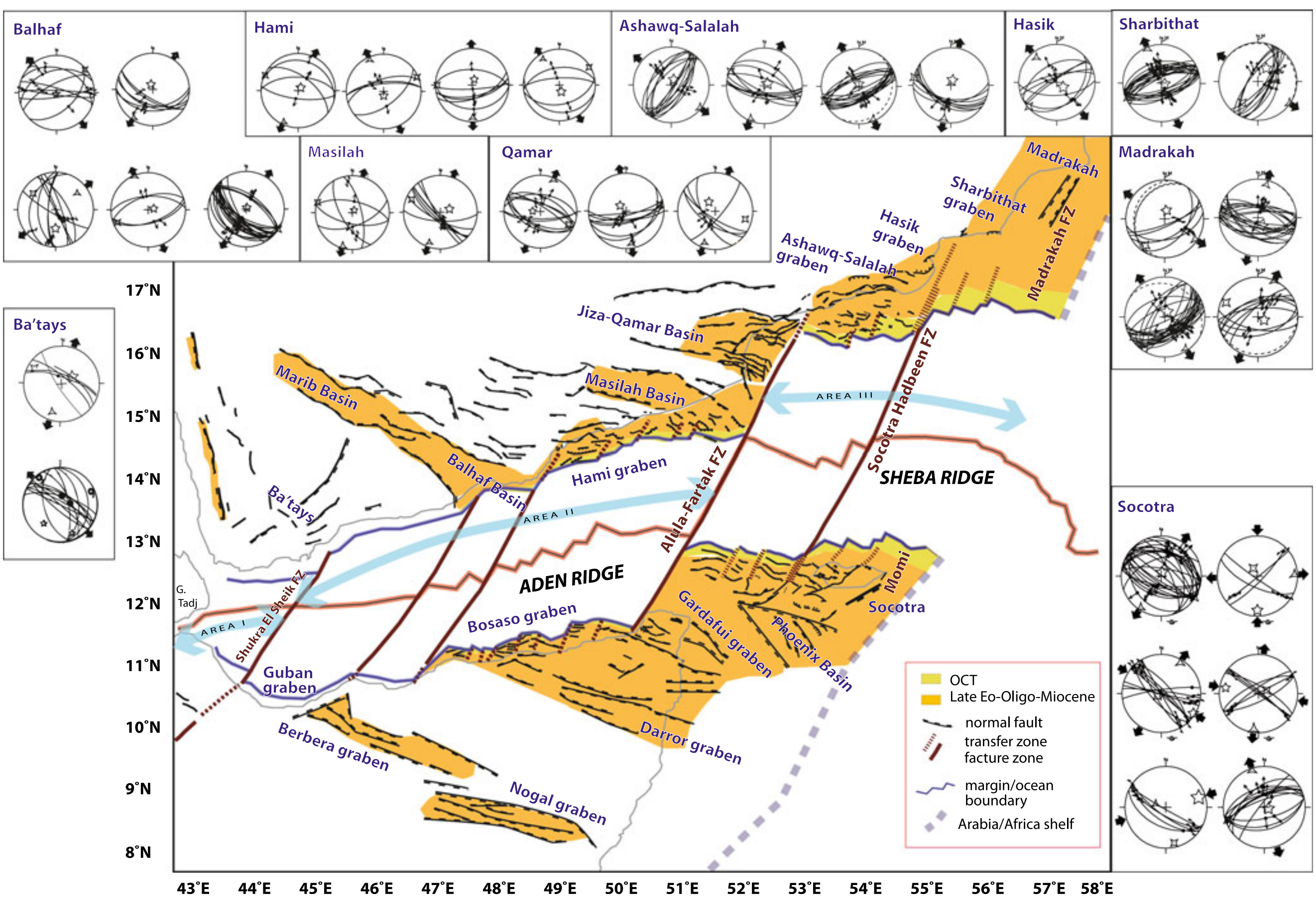


Fig. 6 Synthetic structural map drawn up using published data (Thoué et al. 1997; Bellahsen et al. 2006; Autin 2008) and unpublished oil data. Stereogram synthesis from Thoué et al. (1997), Huchon and Khanbari (2003), Fournier et al. (2004), Bellahsen et al. (2006), and Autin (2008)

carried out by Bellahsen et al. (2003) showed that the orientation of the Gulf was not necessarily controlled by a weakness zone. Both Afar plume impingement and the Neo-Tethys subduction/collision are only necessary to reproduce the trends observed in the Red Sea and the Gulf of Aden, and their associated directions of extension.

Crustal thickness and stresses

Variations of crustal thickness of the region under rifting could generate local stresses. We have deployed a seismological network of 29 broadband seismic stations in northern margin in order to determine crustal thickness. The receiver function method allowed us to map the variations in thickness beneath each station and to discuss the nature and structure of the crust (Fig. 8; Tiberi et al. 2007). Crustal thinning can be observed near the coast where the future ocean will be formed. The crust is 35-km thick in the north and thins down rather abruptly to 26 km under the Salalah Plain (the main extensional basin of the margin outcropping on land). This thinning occurs over a distance of less than 30 km, thus placing the crustal thinning under the first known tilted block of this margin. The second margin transect, which is 50 km farther east, shows no thinning between the inner continent and the coast over a distance of at least 50 km. The maximum crustal thinning only occurs offshore. The thickness variations observed parallel to the crustal structure can be used to determine the scale (wavelength) and persistence of sedimentation parallel to the axis. North of the rifted zone, the thinned crust indicates a large amount of extension, producing a wider rift than suggested by the geological data. This may reflect the existence of a structure inherited from previous rifting, or from local crust thickness variations. In the Dhofar area, the intermediate crust, between the continental and oceanic crust, seems to have a similar felsic to normal composition (Fig. 8). The basement/sediment interface is found at a depth of 3 km, which is consistent with the available geological data (Roger et al. 1989). The sedimentary layer is believed to have a major impact on the values of the Vp/Vs ratio (ranging from 1.67 to 1.91; Tiberi et al. 2007). The upper/lower crust interface is imaged underneath one of the stations (S03; Fig. 8).

A preliminary structural study conducted on the southern margin of the Socotra Island did not lead to the determination of a chronology (Fournier et al. 2007). Our work on Socotra Island (Autin 2008; Bellahsen et al. 2009) identified three fault systems (N110° E, N70° E and N90° E; Figs. 6 and 7b). We attempt to establish two chronologies for these systems (Figs. 6, 7b, c), which could be explained by three major phases (Bellahsen et al. 2006, Autin et al. 2010a). (1) The development of N110° E faults, influenced by a N20° E extension at the boundaries of the plates. The *en echelon* pattern of these faults resulted from lithospheric heterogeneity, or, as suggested by Bellahsen et al. (2003), from an interaction between the Afar hotspot and the boundary conditions. (2) Lithospheric thinning induced major variations in thickness, and consequently local stresses (extension perpendicular to the rift), which led to the development of N70° E faults (Fig. 7c). (3) Extension and thinning continued. Differences in thickness remained only on the rift edges as shown by receiver functions (Fig. 8; Tiberi et al. 2007; Al-Riyami 2007) and new N110° E faults developed in the most thinned sectors in the centre of the rift, sometimes accompanied by the reactivation of N70° E faults.

Transfer fault zones

Socotra Island provides the opportunity to study very carefully a transfer fault zone outcropping in the middle of the island near the Hadibo city (Fig. 7b). The Hadibo transfer zone (HTZ) is composed of several dextral N50–70° E trending transform faults. On each side of the HTZ, the eastern and western parts of Socotra Island display very different structures in terms of deformation distribution and vertical movements. The ages of the terrains (mainly Tertiary limestone in the west and Proterozoic basement in the east) show that movements in the area mainly consisted of major normal-fault displacement (predominantly westward dipping). This fault displacement is even greater in places where the eastern basement compartment has the highest elevations (Figs. 7 and 9). Micro-structural data reveal a pure normal fault displacement for the minor faults showing the same trend as the HTZ (Fig. 6). This interpretation is supported by the fact that these faults have a non-vertical dip, roughly 60–70°, which is typical of normal faults. These faults were then reactivated in a strike-slip regime. Therefore, prior to becoming a transfer zone, the HTZ was initially a major zone of normal faulting before the Oligo-Miocene rifting. Strike-slip faults are numerous in this area, notably due to the permutation of stresses with the extensional regime developed more distally. We consider that the HTZ itself first behaved as a normal fault during the initial stages of extension. HTZ is thus an inherited structure. No E-W to N110°E faults cross-cut this transfer zone, which suggests that the HTZ belongs to the very early history of the margin.

Finally, a NE–SW-striking fault is observed at the eastern extremity of the island, with an fault throw of more than 600 m displacing the Permo–Triassic deposits (Fig. 9). This fault could be Jurassic/Cretaceous in age (Figs. 4 and 7b). By closing the Gulf of Aden, we can correlate the orientation of the faults with those observed in the areas of Huqf (north of Oman) and Sawqrah (southeast of Oman;

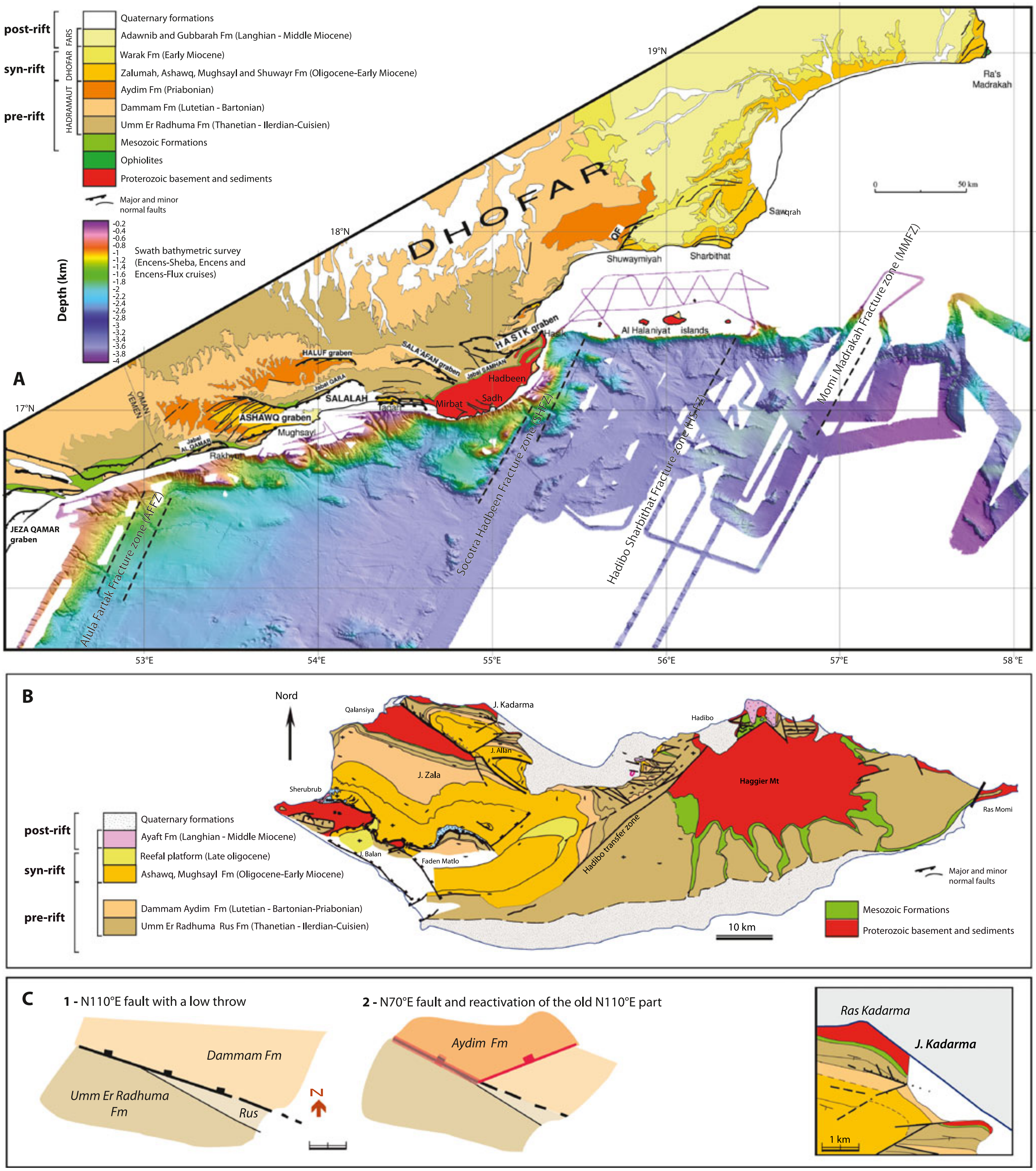


Fig. 7 **a** Geological sketch map of the northern margin of the eastern Gulf of Aden (eastern Yemen and southern Oman) and bathymetric map from the Encens-Sheba cruise (Leroy et al. 2004), the Encens cruise (Leroy et al. 2006; 2010) and the Encens-Flux cruise (Lucazeau et al. 2008); *QF* Qarabyian fault. **b** Geological sketch map of Socotra Island showing complex structure in the western part, with several tilted blocks, and, in the eastern part, a Precambrian granite megapluton controlling the emplacement of the single tilted block. The transfer zone is made up of several fault segments. This map is simplified to point out correlations with the northern conjugate margin. **c** Local chronology in the northwest of Socotra Island, showing a N110° E striking fault system and a second striking N70°E which joins up with the first system. The part in dotted lines has been abandoned. *Inset* Location map

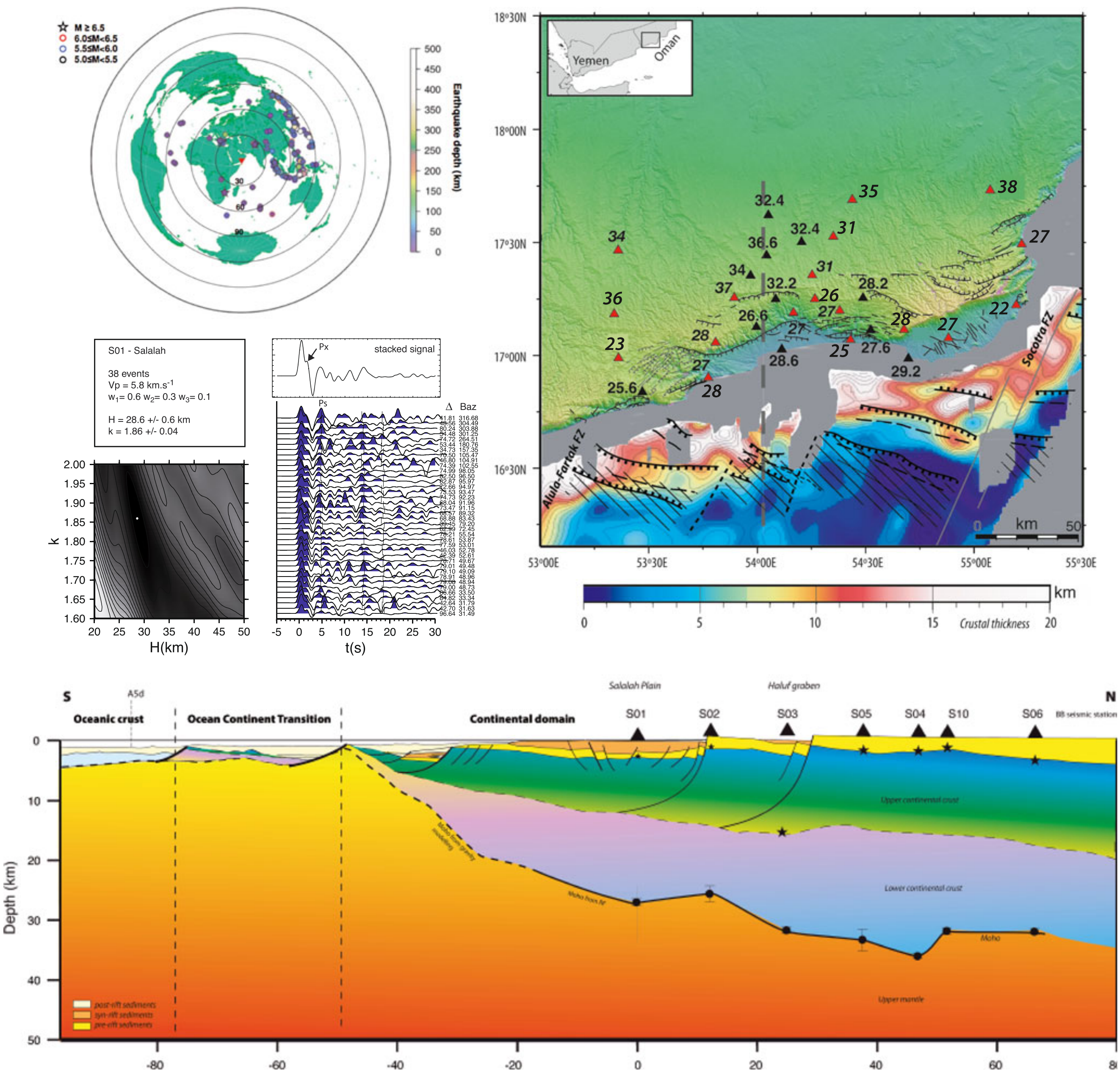


Fig. 8 *Upper left* Azimuthal distribution of the 78 teleseismic events selected for the receiver function analysis. Projection is focused on the Dhofar region network. Crustal thicknesses are determined by analysis of the receiver functions. The stations are represented by *black triangles*. *Upper right* Crustal thickness and the Vp/Vs ratio are indicated next to the considered station. The results from two networks are reported (Dhofar Seismic Experiment; Tiberi et al. 2007; Encens-UK; Leroy et al. 2006; Al-Riyami 2007). The shallowest Moho is observed for stations that are nearest to the coast. Onland faults are noted on the map. Offshore faults and crustal thicknesses are taken from d'Acremont et al. (2006). *Middle left* Example of crustal thickness estimation at the S1 station (Salalah). The *left panel* corresponds to the grid search method in the (H,k) domain (Zhu and Kanamori 2000). Greyscale coding indicates the amplitude of the objective function S (H,k). High amplitudes are reported in *black*. The maximum is indicated by a white dot and corresponds to a given value. The panel on the *right* represents the receiver function obtained using the 38 events (Tiberi et al. 2007). *Bottom* Crustal section on the N-S profile, with analysis of crustal receiver functions (Basuyau 2006)

Figs. 2 and 4). These faults could be related to the formation of the Afro-Arabian margin during the Jurassic–Cretaceous (Fig. 4) and reactivated during the Oligo-Miocene opening of the Gulf of Aden.

Analogue modelling

Analogue modelling of lithospheric oblique rifting allows a comprehensive approach to the problem of fault distribu-

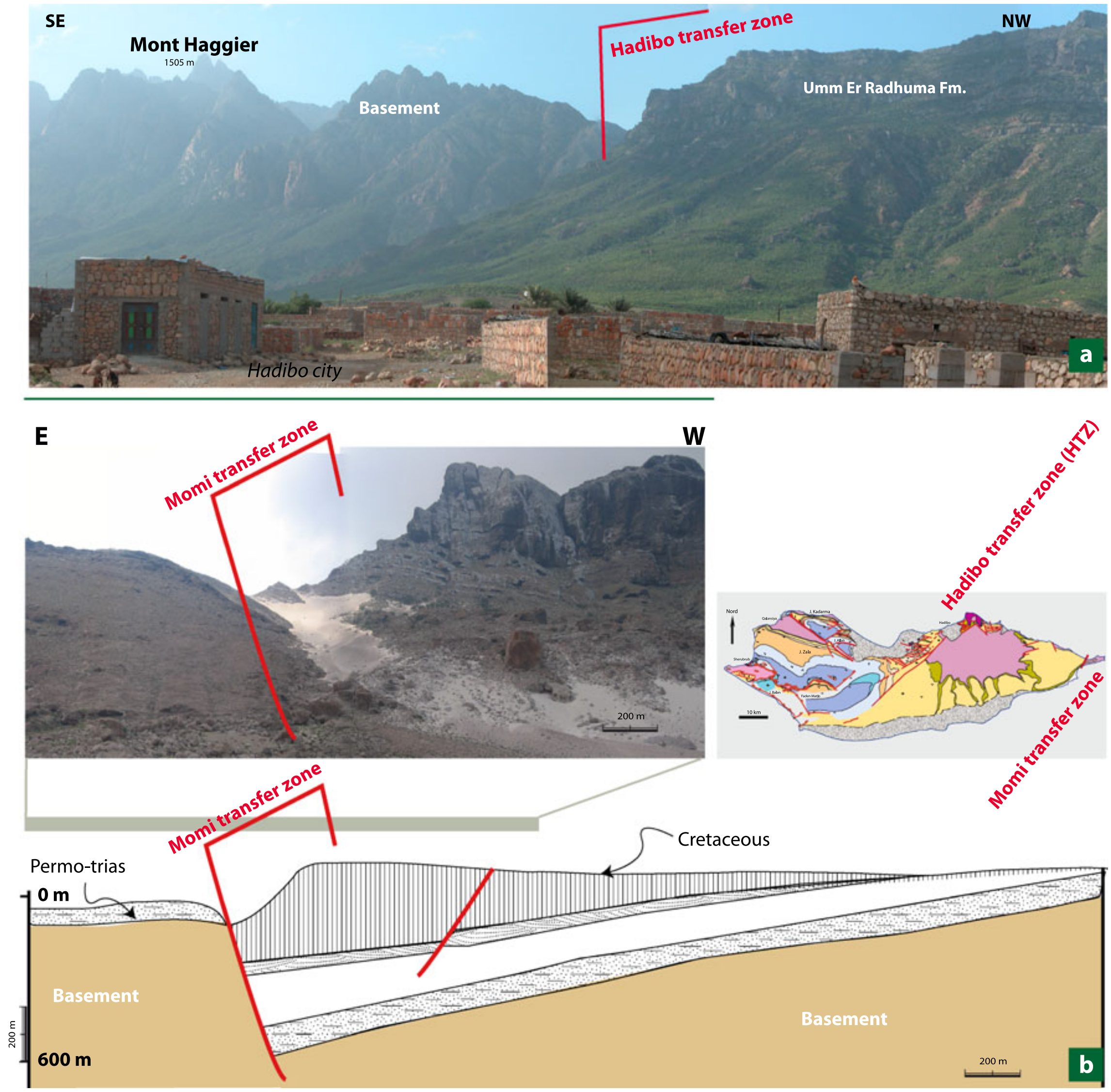


Fig. 9 **a** Hadibo transfer zone (Socotra Island). **b** Fault in the east of Socotra Island, with the same strike as the Hadibo transfer zone

tion, continental rift spreading and chronology. It also takes into account most of the characteristics of natural oblique rifts and offers an overview of the development of faults and segmentation of basins during the formation of a rift (Corti 2008; Autin et al. 2010a; Agostini et al. 2011). Two oblique rift models were constructed using an obliquity of 40° (obliquity being the angle between the direction of displacement in the model and the perpendicular to the orientation of the rift). One of the models does not integrate any pre-existing weakness of the lithosphere (model A; Fig. 10a), while the other takes this factor into account (model B; see Autin et al. (2010a) for details on the modelling). Both models show the following features:

1. The strike of the faults is intermediate between that of the rift and the perpendicular to the displacement direction, particularly during the initial stages of deformation. Such faults, which are typical of oblique rifts, have also been reproduced in previous models (e.g. Tron and Brun 1991).
2. At later stages, the rift becomes localised along the oblique zone, which results in a greater number of rift-parallel faults. The local stresses due to variations in crustal and lithospheric thicknesses during rifting are evidently very important and control the formation of these faults.
3. When thinning is sufficient, deformation occurs in the centre of the rift, where crustal thickness no longer varies significantly. During the final stage of extension in model B, the crust is deformed by rift-parallel faults; in the basins, small-scale deformation is produced by faults perpendicular to the extension direction (Autin et al. 2010a). In model A (Fig. 10d). Faults striking perpendicularly to the displacement direction tend to

accommodate most of the extension. These faults control the final stages of extension, probably conditioning the formation of the ocean–continent transition and the development of mantle exhumation, when this occurs, as well as the geometry of future oceanic spreading centres.

Sedimentary architecture of the proximal margins of the Gulf of Aden

The syn-rift and post-rift Tertiary successions, which have recorded the various opening stages of the Gulf of Aden between the Late Eocene and the Miocene, can be

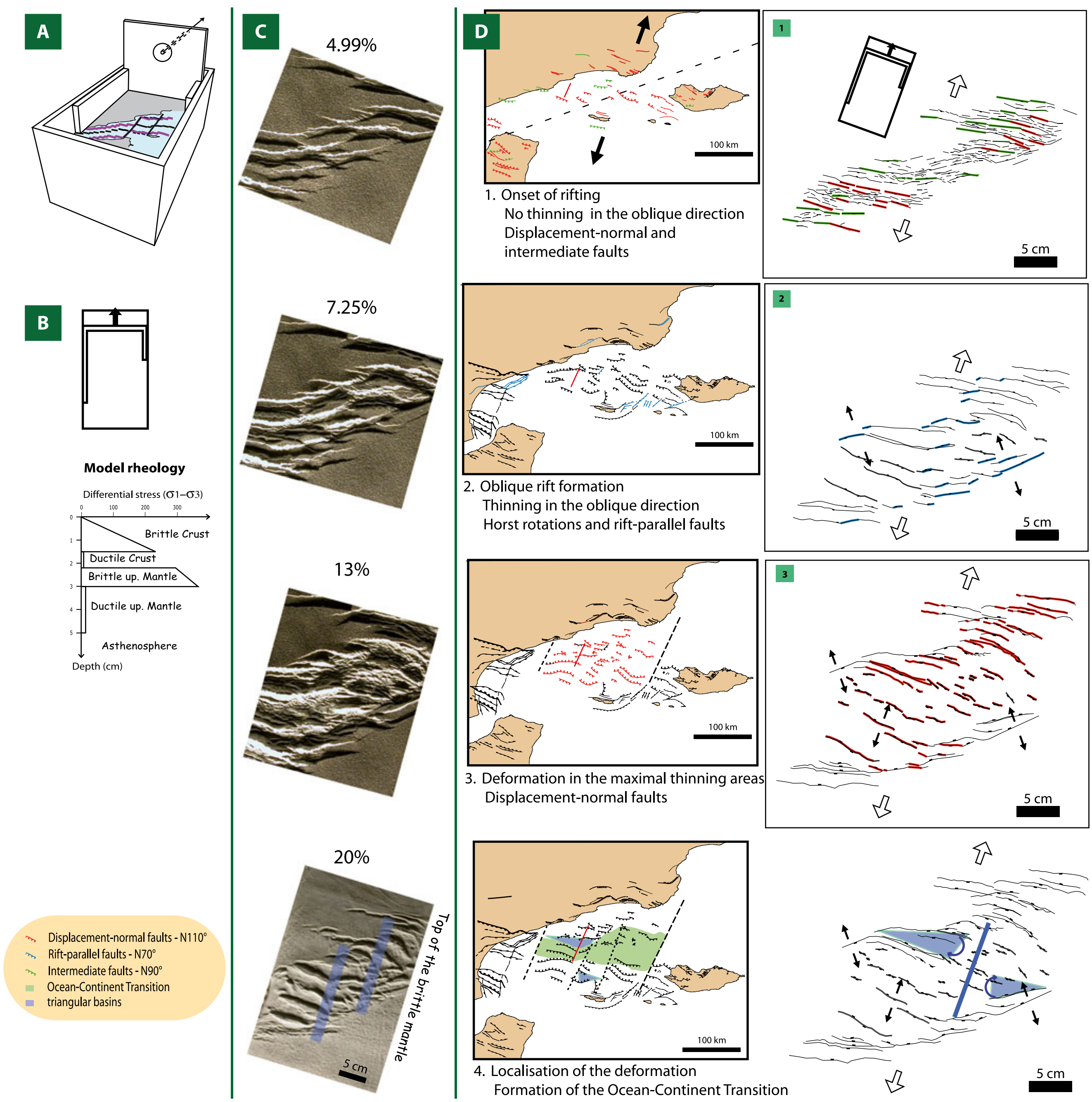


Fig. 10 **a** Box used for analogue modelling. The arms of the sliding box vary in length, thus creating two lateral discontinuities in velocity that impose an oblique orientation to the deformation, similar to that found in the Gulf of Aden. **b** Rheological profile of the model. **c** Main stage of evolution of the model. The model is rotated to match the Gulf of Aden. **d** Analogies between the model and the eastern Gulf of Aden. Reconstruction of the opening of the eastern Gulf of Aden, based on the onland and offshore structural pattern. Line-drawing of the model, tilted so as to match the direction of the gulf extension (see Autin et al. 2010a)

investigated on the northern margin of the Gulf, in Yemen and Oman (Province of Dhofar), as well as on its southern margin, on the Yemeni island of Socotra (Figs. 2 and 7). The outcrops of the Somalian margin have been studied by several authors (Fantozzi 1996; Fantozzi and Sgavetti 1998; Fantozzi and Ali-Kassim 2002).

The eastern part of the Gulf of Aden

The northern margin: Dhofar

The Tertiary successions of Dhofar were studied extensively during the BRGM mapping expeditions, which established the broad outlines of the stratigraphy and tectono-sedimentary evolution of this region in relation to the opening of the Gulf of Aden (Platel and Roger 1989; Roger et al. 1989). The present study provides additional stratigraphic details and suggests some new elements of interpretation. Geometric features of the stratigraphic architecture are examined closely here to establish a link with the offshore part of the margin and refine our knowledge of seismic horizons and margin history. This study includes an overview of the different successions, which are described in more detail in the literature for this area, and discusses the various time-markers in the history of the margin, as revealed by the onland seismic profiles established during our comprehensive field study.

The pre-rift successions in Dhofar The pre-rift sequence of the study area includes a Precambrian crystalline basement, which crops out in the region of Mirbat and on Hallanyat Island (Fig. 7a; Mercolli et al. 2006). This basement extends over the entire margin, beneath the syn-rift basins (Fig. 7a). However, the history of exhumation of this basement is different because of its particular evolution (Petit et al. 2007).

A transgression during the Barremian–Aptian is recorded in the Mesozoic sediments (Platel and Roger 1989). The pre-rift Tertiary succession, known as the Hadramaut group (Fig. 11), includes two thick units showing shallow shelf facies, the Umm er Radhuma Formation (Palaeocene/Early Eocene) and the Dammam formation (Early Lutetian/Bartonian). The deposition of these two shelf sequences is interrupted by a period of emergence, characterised by the Rus Formation (60 m of evaporites and dolomites).

During the Eocene, all the margins of the Arabian shelf record a regression indicating a major uplift. This is followed by an accumulation of Priabonian sediments (ca 100 m), making up the Aydim Formation, which required the development of areas of subsidence along the future emplacement of the rift (Roger et al. 1989).

The syn-rift series in Dhofar The syn-rift succession is located within strongly subsiding grabens that reflect a regional transgression related to rifting (Fig. 7a). Our study first investigated the area situated between the Alula–Fartak fracture zone (AFFZ) and the Socotra–Hadbeen fracture zone (SHFZ), and then the area between the SHFZ and the Momi–Madrakah fracture zone (MMFZ; Fig. 7a).

1. Between the AFFZ and the SHFZ

   The Zalumah Formation, which is dated as Priabonian/Early Oligocene, is made up of sabkra and dolomitic paleosols located in confined areas (Roger et al. 1989). The Ashawq Formation (Rupelian/Early Chattian) is made of shallow water deposits that can be divided into two members: the Shizar Member, which is sandy and clayey, and the Nakhlit Member, which forms a thick band of reefal limestone (Fig. 12). At the end of the Oligocene, the rifting intensified and sedimentation took place, associated with major faults, of deposits typical of the basal Mughsayl Formation, with turbidites giving rise to the rapid accumulation of prograding sedimentary prisms during a transgressive period (Figs. 11 and 12).

   There are varied types of contact between the Ashawq and Mughsayl Formations. In places, the Ashawq Formation is composed of reefs draped by sediments of Mughsayl-type facies (Fig. 13a, b); the contact is sometimes faulted (Fig.12) and sometimes stratigraphic. These reefs can be up to 400 m thick and are observed locally on land (Fig. 13b) as well as on seismic profiles crossing the Salalah plain (Fig. 13c), which facilitated their mapping. These reefs were also encountered in the SLP1 well (Fig. 13a, c).

   The top of the Mughsayl Formation is dated as Mid Miocene (Middle Burdigalian) (Roger et al. 1989). A regressive facies sequence can be identified here, capped by a Gilbert-delta type prograding system (Figs. 13c and 14a). These prograding clinoforms result from the filling of the basin by gravity-driven sedimentation, which continued from the Burdigalian until emergence during the Langhian. The build-up of this system provides evidence of the existence of active relief, whose erosion supplied the Gilbert-type deltas. The discovery of this prograding system, accompanied by the emplacement of "Gilbert-type deltas" (Leroy et al. 2007) shows that basin filling took place in the proximal part of the margin, during a phase of increasing deformation and the development of marked relief, at the onset of the break-up of the continental crust.
2. Between the SHFZ and the MMFZ

   The two lithologies making up the Ashawq Formation, i.e. the detrital facies of the Shizar Member and

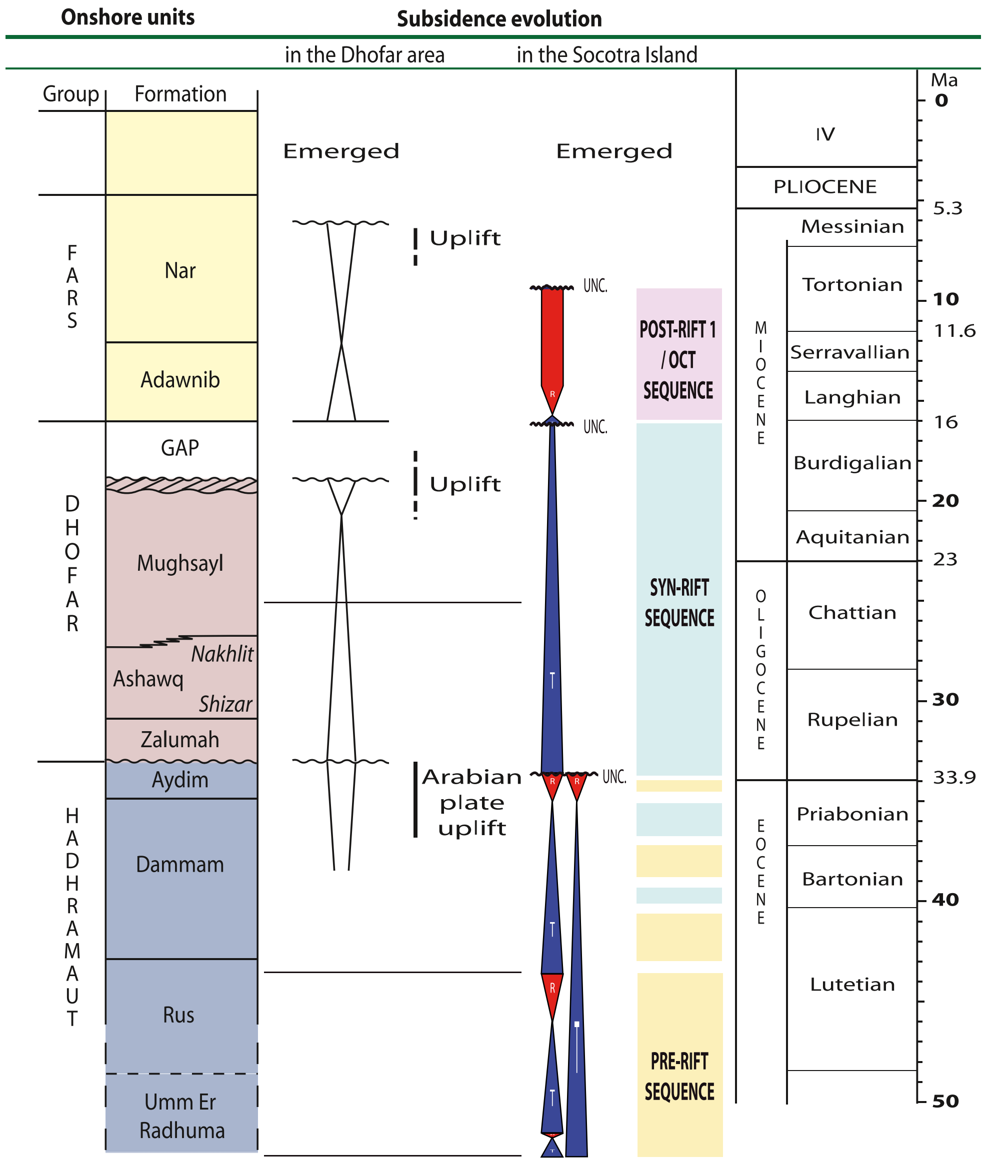


Fig. 11 Chart summarising the stratigraphic formations of the various sedimentary formations and sequences on the conjugate margins (Dhofar and on Socotra Island)

the shallow carbonate shelf facies of the Nakhlit Member, are also found in this area, but to a much lesser extent. Indeed, the entire formation barely reaches a thickness of 100 m in this area, compared with more than 700 m in Western Dhofar (Platel et al. 1992a).

The Shuwayr and Warak Formations have been defined east of the Qarabiyan fault. They account for most of the limestones cropping out on the coastal plain of Sharbithat and extend westwards to the town of Shuwaymiyah and onto the slopes of wadi Warak (Fig. 7a). Their thickness ranges from 150 to 250 m. These well-bedded coral limestones of reefal bioclastic type are intercalated with debris flows rich in coral fragments, echinoid spiculae and reworked foraminifera (Fig. 14b; Platel et al. 1992a).

Dating from the Early Oligocene to the Early Langhian, according to the fauna found in the debris flows (Bivalves), these deposits are considered coeval with the Mughsayl Formation found further west in the area of Salalah (Figs. 11 and 14b; Platel et al. 1992b).

In comparison with the succession found throughout the “Ashawq graben/Salalah plain” area, the reduced thickness of syn-rift deposits in the Shuwaymiyah area (Ashawq and Shuwayr/Warak Formations) is indicative of a lesser degree of subsidence in this zone during the Oligo-Miocene.

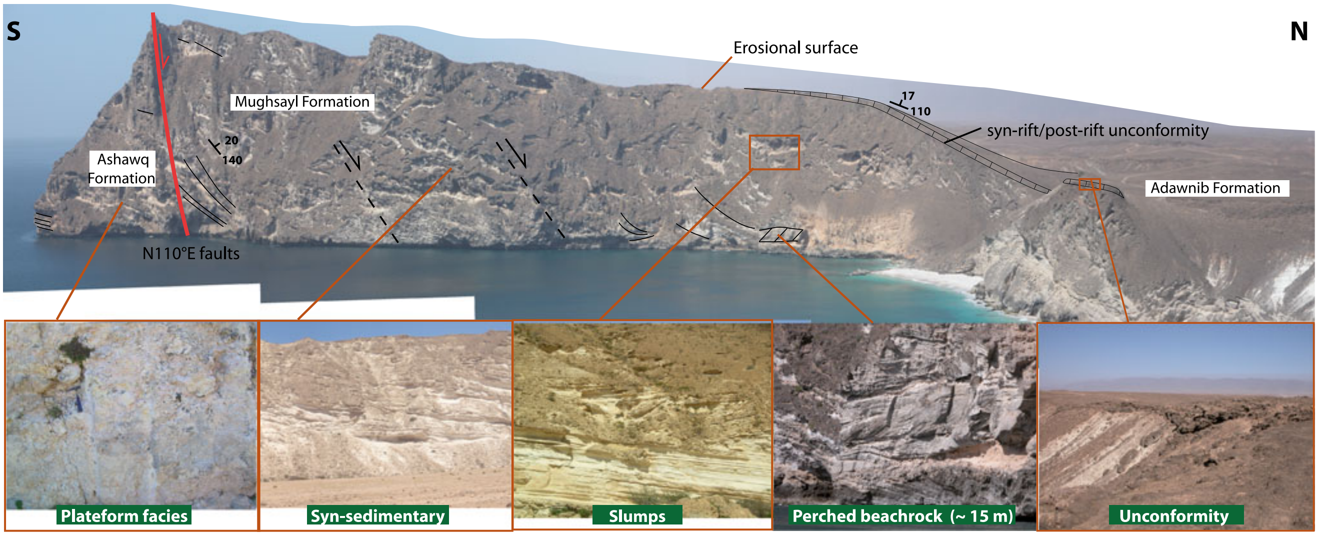


Fig. 12 Exposure of the syn-rift successions and syn-rift/post-rift unconformities in the south of the Salalah plain. Close-up view of the various outcrops and sedimentary systems observed (Ras Hamar). See Fig. 13a for location

In the Shuwaymiyah area, the Warak Formation is also cross-cut by major normal faults. Along these faults, the throw can reach 200 m (Platel et al. 1992a). By comparing the pre-reef facies of the Warak/Shuwayr Formations with the reefal facies of the Nakhlit member, we can identify an episode of retrogradation in this area, but in a context with less marked subsidence. A phase of progradation is also visible in the Sharbithat plain. However, the sedimentary dynamics are entirely different, showing neither active relief nor any emergence (Pointu 2007).

The geometry of deposition and the lateral facies variations, from west to east, indicate a segmentation of the margin, parallel to the paleo-rift axis (Fig. 7a).

The post-rift successions in Dhofar The post-rift sequence can be identified by the continental deposits lying unconformably on top of the syn-rift sediments (Fig. 11). A phase of emergence, during which there is no sedimentary record, is marked by major erosion surfaces visible on most of the uppermost syn-rift deposits (Fig. 12).

The Adawnib Formation is characterised by a very shallow-water depositional environment during a transgressive phase (Langhian to Serravallian; Figs. 11 and 14c). Regression started here at the beginning of the Late Miocene (Fig. 10; Platel et al. 1992b). The Nar Formation, which is dated as Miocene/Pliocene, contains continental detrital material (Fig. 14c). These conglomeratic deposits of the Nar formation reflect the final emergence of Dhofar. However, the preservation of this 30-m thick section of conglomerates indicates a subsidence of this currently emerged part of the margin.

In the Sharbithat area and as far as Ras Madrakah, the post-rift Gubbara Formation was deposited unconformably on the pre-rift (Dammam Formation) and syn-rift successions (Fig. 7a). Analysis of the molluscan and foraminiferal fossil record indicates that these deposits are of end-Langhian to Serravalian age (Platel et al. 1992a). The lateral stratigraphic correlation established with the Adawnib Formation is yet another argument in favour of attributing a Langhian–Serravallian age to the Gubbara Formation.

The Quaternary deposits in Dhofar The Dhofar Quaternary deposits include a succession of coastal and lacustrine deposits (travertines), alluvial terraces and piedmont deposits (Platel and Roger 1992). Moreover, several paleo-beach terraces (stepped beachrock) can be seen all along the Oman coast (Fig. 7a).

Four levels of stepped beach rock have been identified (Fig. 15). Two have been dated and the oldest (~2 Ma) is located at an elevation of 60 m. The second terrace is 20–26 m above the present-day beach. Carbon-14 dating yields an age of 35,200 years BP. The third terrace, which lies 10–13 m above sea level has not been dated. The fourth terrace borders the coastal plain at an elevation of 2 m (Platel et al. 1992a; Leroy et al. 2007; Fig. 15).

Summary of the sedimentary record of vertical movements on the Dhofar margin The syn-rift period started in the Late Priabonian/Early Rupelian (33–34 Ma) and ended in the Early Miocene (18 Ma), after the Late-Eocene emergence of the Arabian platform (Fig. 16).

The margin has recorded an uplift, followed by an increase of the available space due to overall subsidence. The lacustrine limestones of the Zalumah Formation were

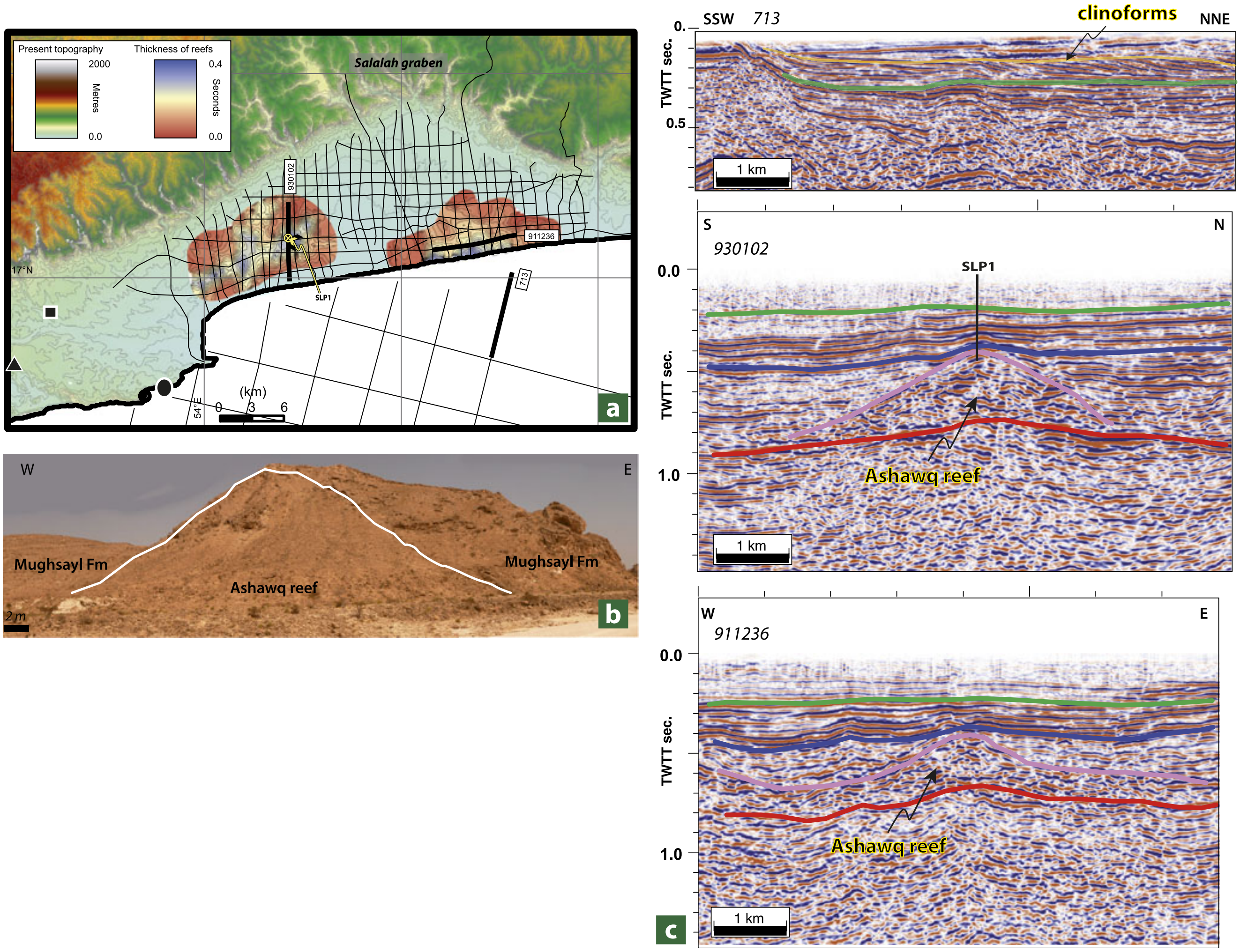


Fig. 13 **a** Topography of the Salalah graben and isopach map of the reefs investigated by the seismic survey. Location of Fig. 12 (*black dot*), Fig. 14a and c (*black square*), **b** (*black triangle*) and profiles on **c** (*black lines*). **b** Outcrops of reefs (Ashawq Formation) draped by deeper-water deposits of the same type as in the Mughsayl Formation. **c** Seismic profiles traversing the reefs (Ashawq Formation) draped by deeper water deposits of the same type as in the Mughsayl Formation (*bottom and middle*). Seismic profile near the coastline showing a Gilbert-type delta filling. Colour coding is as follows: *red* top pre-rift (Shizar Mb), *pink* top Oligocene reef, *blue* intra Mughsayl, *green* top Mughsayl Fm

deposited at that time. Subsidence intensified on the margin, notably towards the east, which created a general south-easterly slope. The marly deposits of the Shizar member belonging to the Ashawq Formation are followed by reefal platform deposits (Nakhlit Member). This formation prograded vertically and laterally with respect to the deep-water carbonate systems of the Mughsayl Formation. The latter corresponds to calcareous gravity-driven slump deposits, which indicate the development of steep slopes and deep basins controlled by faults along the continental margin (Fig. 16).

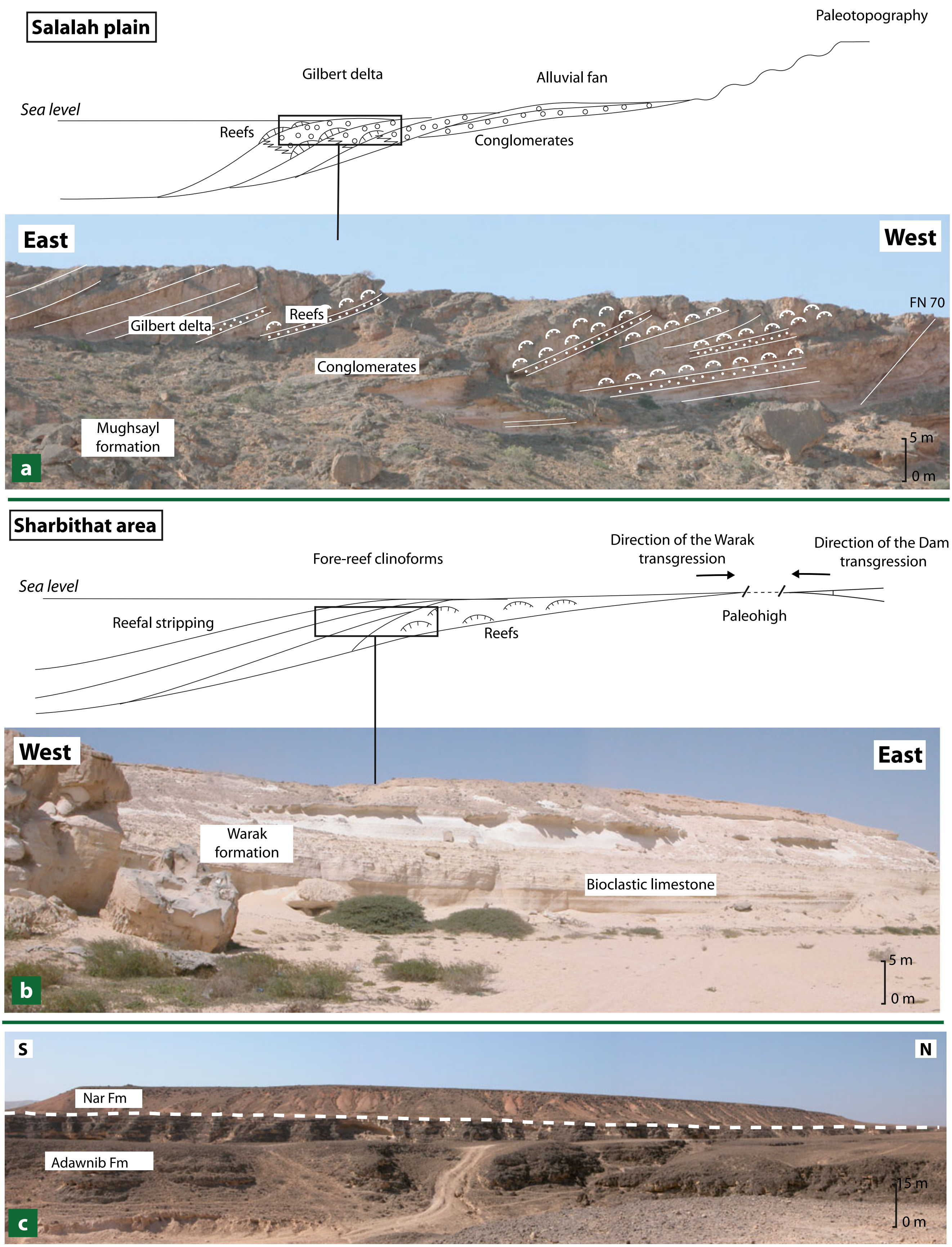
During the Burdigalian, the proximal part of these deep syn-rift basins was filled by the conglomerates of prograding alluvial fans. Such sedimentary systems indicate a major uplift of the continental margin during the continental break-up (Autin et al. 2010b).

Since the Langhian, a thick post-rift succession accumulated on the distal part of the margin, extending all the way to the present-day coastline.

The southern margin: Socotra Island

The conjugate margin can be analysed in detail on land on the Island of Socotra (Figs. 2 and 17a). Two structural domains are separated by a transfer zone, with a tilted block

Fig. 14 **a** Top of the Mughsayl Formation showing a Gilbert-type delta filling. **b** Warak formation and interpretation of the depositional sequence. **c** Post-rift successions (Adawnib/Nar formations contact). See Fig. 13a for location



Salalah plain
Paleotopography
Gilbert delta
Alluvial fan
Sea level
Reefs
Conglomerates
East
West
FN 70
Gilbert delta
Reefs
Conglomerates
Mughsayl formation
5 m
0 m
a
Sharbithat area
Direction of the Warak transgression
Direction of the Dam transgression
Fore-reef clinoforms
Sea level
Paleohigh
Reefal stripping
Reefs
West
East
Warak formation
Bioclastic limestone
5 m
0 m
b
S
N
Nar Fm
Adawnib Fm
5 m
0 m
c

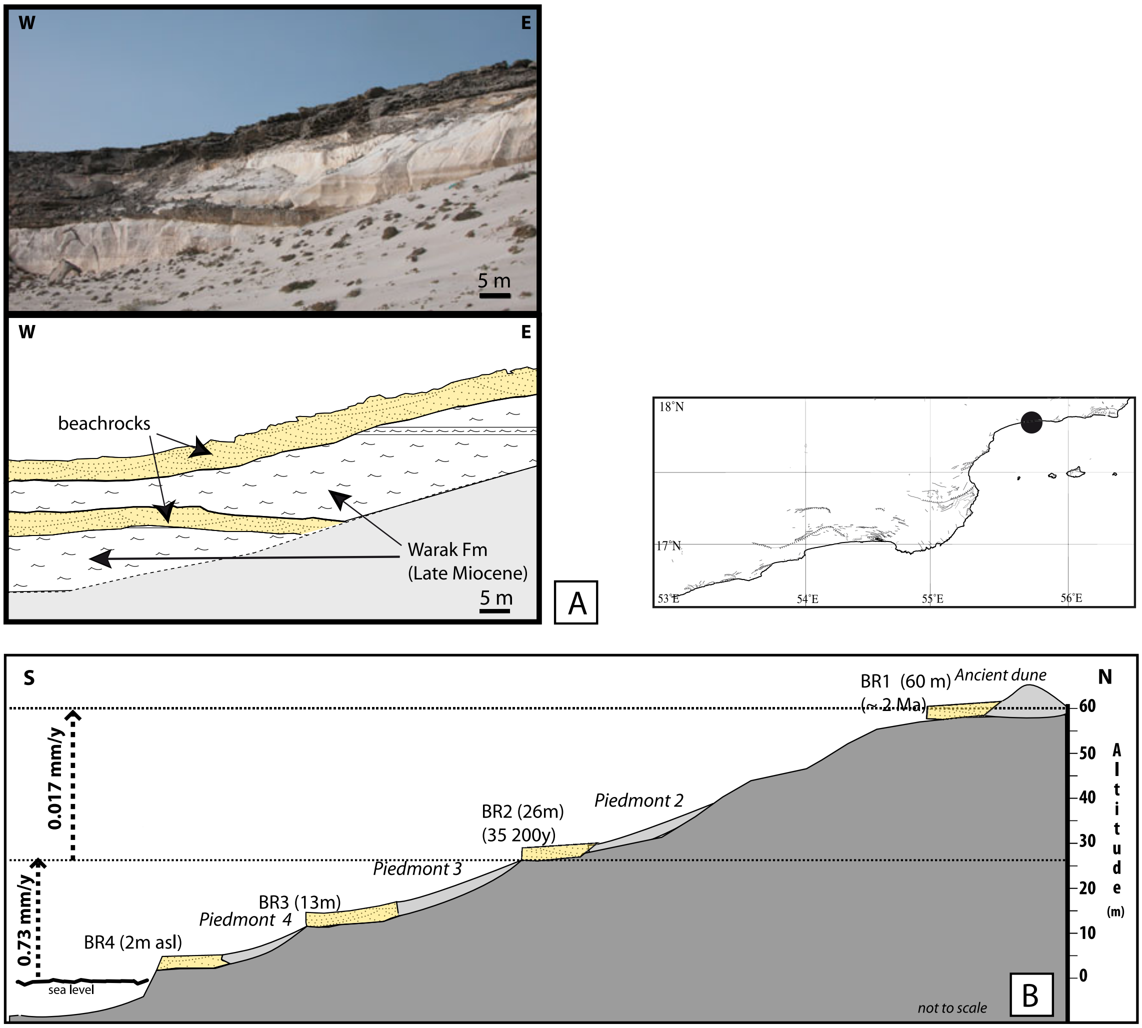


Fig. 15 Photos of perched and stepped beachrocks. Location map and N-S section showing heights above sea level and ages of these uplifted paleo-beaches

in the east and two tilted blocks in the west, which enabled us to conduct an intensive study of the tectonic/sedimentary relationships (Fig. 17a).

A stratigraphic review and complete mapping of the Tertiary successions on Socotra were performed to characterise the geometry and mode of functioning of this sector of the southern margin of the Gulf of Aden, and to compare these structures with the northern conjugate margin, both inland (Dhofar, Sultanate of Oman) and offshore.

Some of these major structures had been identified in previous studies (Beydoun and Bichan 1969; Samuel et al. 1997; Fournier et al. 2007). Unfortunately, their geological interpretation was distorted due to an erroneous geological map. A meticulous stratigraphic and tectonic study reveals a complex structural pattern, which considerably improves our understanding of the formation of the margin system as a whole (Fig. 7b).

The normal faults, striking N50° E to N70° E and N90° E to N110° E, display fairly complex geometrical features, with non-cylindrical flats and ramps (Figs. 7b and 17b). The Paleogene and Neogene successions, with a thickness of 1,500–2,000 m show considerable stratigraphic and sedimentological similarities with those found in Dhofar (Oman) on the north conjugate margin. They can be divided into three major groups (Fig. 11).

Paleocene to Mid Eocene pre-rift deposits The pre-rift successions of the Paleocene/Middle Eocene correspond to carbonate shelf successions comparable to those found on the Arabian plate and in Somalia. The two major transgressive phases of the Thanetian/Early Eocene (Umm Er Rhaduma formation) and the Middle Eocene (Dammam formation) are recorded here and are separated by a regressive episode during the Late Cruisian–earlymost Lutetian (Rus formation; Figs. 11 and 18a).

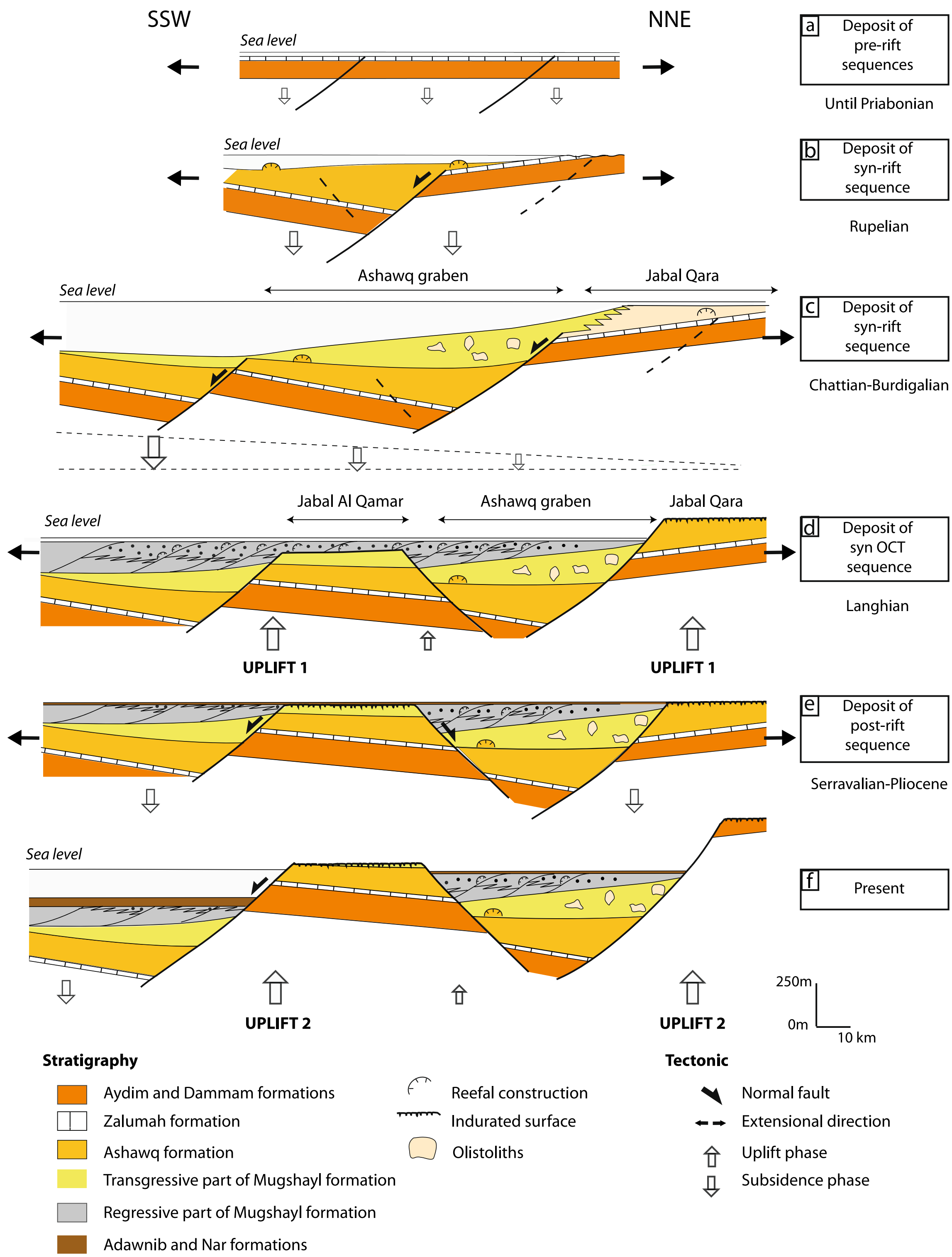


Fig. 16 Diagram of the tectonic and sedimentary evolution of the northern margin of the Gulf of Aden cropping out in Dhofar. See text for explanations

Late Eocene to Early Miocene syn-rift deposits Towards the end of the mid Eocene, the carbonate shelf recorded a change in depositional profiles due to the initiation of tectonic activity, and hence the beginning of the rifting phase (Fig. 11). The Early Oligocene syn-rift deposits correspond to sub-reefal carbonate shelf facies (Ashawq Formation). During an initial stage, syn-sedimentary fault displacement and differential subsidence were broadly

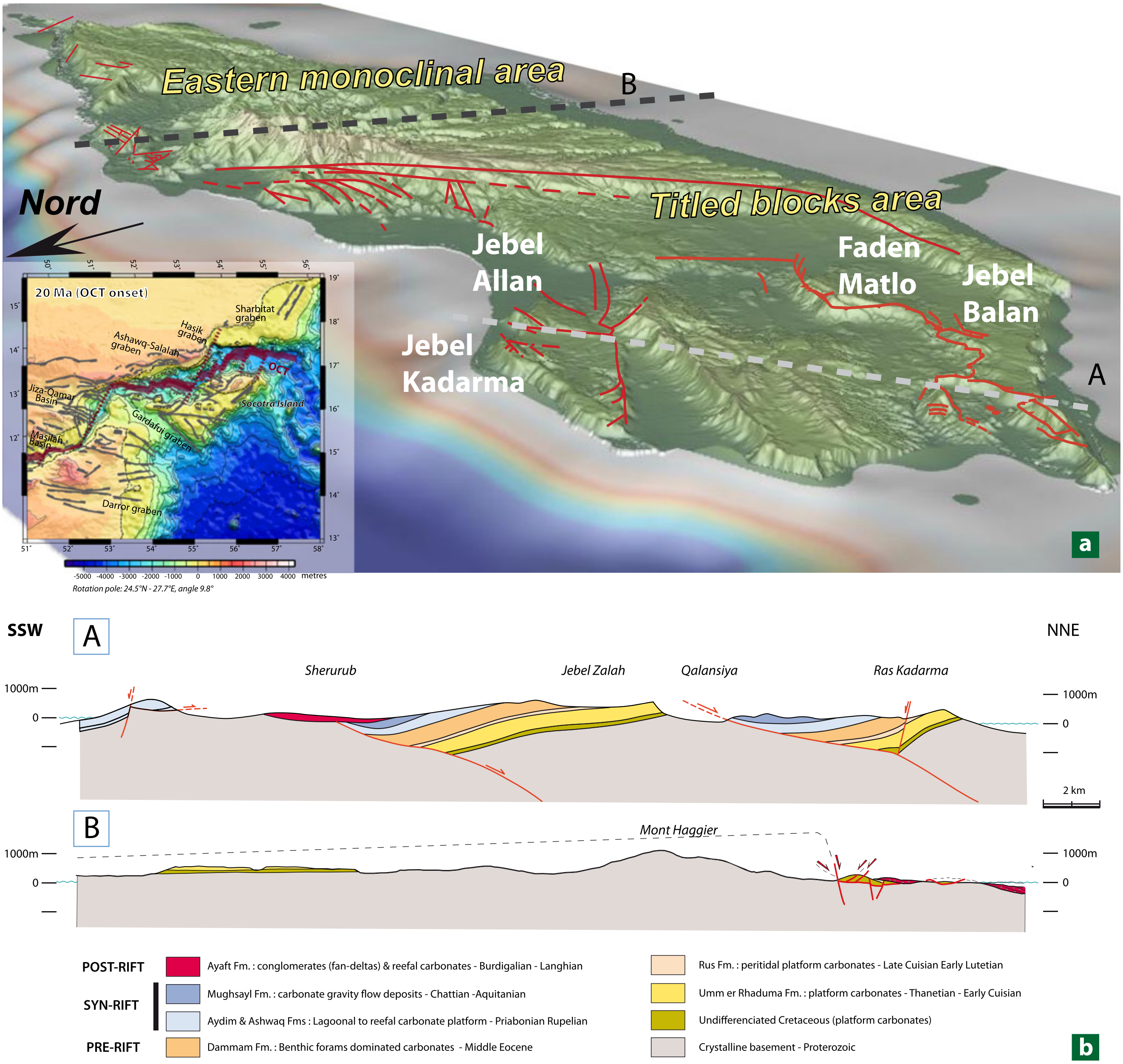


Fig. 17 **a** Three-dimensional view of the topography and bathymetry of Socotra Island and diagram showing structure of the two main domains. *Inset* Kinematic reconstruction of the eastern Gulf of Aden at the onset of formation of the ocean–continent transition (OCT), about 20 Ma ago. **b** Synthetic geological sections of the two parts of Socotra Island. Section A crosses the two tilted blocks that crop out on land in the western area. Section B traverses the eastern part of Mount Haggier, showing its monoclinal structure

compensated by carbonate accumulation, which managed to maintain the shelf profile (Fig. 18b, c).

The accentuation of extensional processes triggered a subsidence of the shelf during the Late Oligocene and created deep sub-basins subject to gravity-driven carbonate sedimentation. These marginal reef shelves persisted on the structural highs, supplying sediment to the gravity-driven systems. The rate of sedimentation remained relatively low in the basin, thus determining a complex topography of the margin, marked by its segmentation into numerous more or less connected sub-basins, separated by submarine escarpments (Fig. 18a). These basins are highlighted by the deposition of breccia prisms along the active normal faults (Fig. 18a–c).

Miocene to Quaternary syn-OCT (coeval with OCT emplacement) and post-rift deposits Towards the end of the Early Miocene, the progradation of the conglomerate fan-delta deposits filled in the local basins (Fig. 18d), which is indicative of a major uplift phase, similar to that occurring on the northern margin of the Gulf (Dhofar; Fig. 14a). This was very rapidly followed by a new phase of subsidence, which contributed to the preservation of major fan-delta deposits (and of the equivalent peri-reefal shelf; Fig. 18d) lying discordantly on different members of the successions. These include syn-rift and pre-rift successions and the exhumed Proterozoic basement (Fig. 18e).

Some thin horizontal formations post-dating any deformation are exposed on the coastal plain in the north of Socotra Island. The elevation of the oldest unit is between 20 and 30 m. It constitutes metre-thick limestone beds, totally recrystallised and containing abundant monospecific bivalves. This unit is locally karstified. On lower ground, a coastal carbonate succession, made up of a combination of reefal and bioclastic facies (red algae and corals) forms the substratum of the coastal plain. Layers of polygenic conglomerate, corresponding to detrital inputs during flooding, are interspersed between these coastal reef flat-lying deposits. Superficial fossil/residual continental formations locally overlie these Plio-Quaternary marine deposits. These either represent conglomerate outwash from the alluvial fans, or eolian deposits composed of sandy carbonate material.

Contrary to recent sediments in Dhofar, which are often located at relatively high elevations on the northern margin, the Socotra Plio-Quaternary formations remain very close to sea level.

Summary of the sedimentary record of vertical movements on the Socotra margin The pre-rift carbonate shelf successions between the Paleocene and the Middle Eocene are identical to those of Dhofar and, overall, to those of the Arabian platform (e.g. Bott 1982). The syn-rift formations are also comparable. As on the northern margin, they have recorded distensive tectonic events during the Late Eocene/Oligocene, as well as the development of syn-sedimentary faults and the creation of a deep basin subject to gravity-driven sedimentation (Fig. 19). The progradation of coastal deposits with conglomerate facies indicates that this event was followed by an uplift phase, which brought about filling of the basin. This tectono-sedimentary episode is interpreted as being synchronous with the continental rifting and the emplacement of the OCT at the foot of the margin (d'Acremont et al. 2005; Autin et al. 2010b; Leroy et al. 2010b). Analogies could be drawn with the developmental stage of "sag basins" on the Atlantic margins. These "syn-OCT" deposits were, in turn, uplifted and affected by late tilting. This major uplift during the syn-rift/post-rift transition appears to be expressed symmetrically on both margins (Leroy 2010a, b). These earlymost post-rift deposits were then uplifted and affected by late tilting. This contrasts with more recent deposits, probably dating from the Late Miocene/Quaternary, which were not uplifted, as opposed to the Dhofar deposits on the northern margin. The generalised uplift of Plio-Quaternary deposits on the northern margin can be observed as far away as the Oman mountains. Being apparently related to the movements and deformation of the Arabian plate (e.g. Zagros collision), this phenomenon could be independent of the mechanisms related to the opening of the Gulf of Aden.

The western part (west of Alula Fartak)

Several investigations were carried out in the basins of Qamar and Sayhut-Al Masilah in Yemen (Fig. 2), which make up the northern margin of the Gulf of Aden.

The Jiza-Qamar basin

Syn-rift and post-rift deposits are only encountered in the south of the Jiza-Qamar basin onland whereas they are thicker offshore (Brannan et al. 1997). The syn-rift deposits are represented by a thick inner-shelf carbonate succession dating back to the Oligocene and comparable to that seen in the Ashawq formation in Dhofar (Fig. 20a). This succession is tilted toward the SE and displays normal faults trending N110° E. It has a strongly erosive contact with an overlying conglomerate succession, which was accumulated in an alluvial fan system. This continental succession, considered as post-rift, is slightly deformed and fractured. Stratigraphic fans suggest syn-sedimentary deformation. The emplacement of this continental succession and the truncation of the Oligocene shelf succession indicate a major uplift of the domain, with the creation of relief during this post-rift phase (Fig. 20a).

The Sayhut–Al Masilah basin

The Sayhut-Al Masilah basin is located on the coast east of the town of Al-Mukalla (Fig. 2) and is characterised by a thick predominantly siliciclastic succession that accumulated during the syn-rift phase (Watchorn et al. 1998).

The thick continental syn-rift succession, mainly made up of fluviatile siltstones, lies unconformably on top of the carbonate formations dated as Early to Middle Lutetian (Fig. 20b). This discontinuity therefore implies the existence of a relatively intense phase of erosion between the pre-rift and syn-rift deposits, which is not the case further east in Dhofar or Socotra.

This syn-rift succession exhibits an overall "transgressive" vertical trend. In the median part of this succession, coarse terrigenous facies decrease in abundance at the expense of evaporite deposits and lacustrine carbonates. The upper part of


N
S
Cenomanian
Basement
Umm Er Rhaduma Fm. (Palaeocene - Early Eocene)
Rus Fm. (Cuisian-Early Lutetian)
Dammam Fm. (Mid Eocene)
Mughsayl Fm. - Late Oligocene calciturbidites
200 m
a
SW
NE
Mughsayl Fm. (Late Oligocene)
Ashawq Fm. (Early Oligo.)
Calciturbidites
Carbonate platform
Crystalline basement
50 m
b
NW
SE
Mughsayl Fm. (Late Oligocene)
Ashawq Fm. (Early Oligo.)
Reefal Platform
Calciturbidites
50 m
c
SOUTH
NORTH
Ayaft Fm. - Fan-delta conglomerates
Contemporaneous reefal carbonates
40 m
d
e
East
West
Cretaceous + Palaeocene
Crystalline basement
Reef carbonate &
breccia with basement clasts
Miocene
Crystalline
basement
2 m

Fig. 18 **a** Faden Matlo tilted block, showing the pre-rift sequence tilted by a normal fault and capped by breccias and the Late Oligocene syn-rift formation (Mughsayl Fm.). **b** Syn-rifts deposits, initially shelf deposits, then turbiditic, associated with the emplacement of a fault with very gentle dips (<20°). **c** Early Oligocene reefal shelf deposits (Ashawq Fm) and Late Oligocene calciturbidites (Mughsayl Fm.). **d** Lateral transition in post-rift facies deposits (Razin et al. 2010) from alluvial-fan type deposits to reefal carbonates, corresponding to deposits that could have been laid down during the formation of the OCT. These sediments are called syn-OCT sediments (d'Acremont et al. 2005; Autin et al. 2010b; Leroy et al. 2010b). **e** *Left* Mega-block of crystalline basement in the Mughsayl Fm. carbonates. Middle: Alluvial-fan type conglomerates. *Right* Contact between the crystalline basement, reef carbonates and Miocene breccias with basement clasts

this succession shows a combination of extremely varied facies, typical of marine/continental transition domains. The presence of Cenomanian, Maastrichtian and Paleocene pebbles within the fluviatile conglomerates is taken as an argument in favour of the functioning of faults with a strong vertical throw leading to the denudation of these successions on the borders of subsiding continental basins.

This predominantly terrigenous continental succession is overlain, without any transition, by a carbonate formation showing a peri-reefal facies. The presence of *Miogypsina* sp. is indicative of an Aquitanian/Burdigalian age. The base of this formation records a maximum flooding during this syn-rift phase on the scale of the entire domain. The upper part of the formation is characterised by the presence of large progradational clinoforms of reefal facies that record an already regressive stage (syn-rift/post-rift transition). The dating of this carbonate formation tends to constrain the age of the underlying terrigenous formation, which could therefore be attributed to the Oligocene. The upper boundary of the carbonates is not observable. More recent sub-horizontal formations overlie the tilted syn-rift successions in angular unconformity (Fig. 20b).

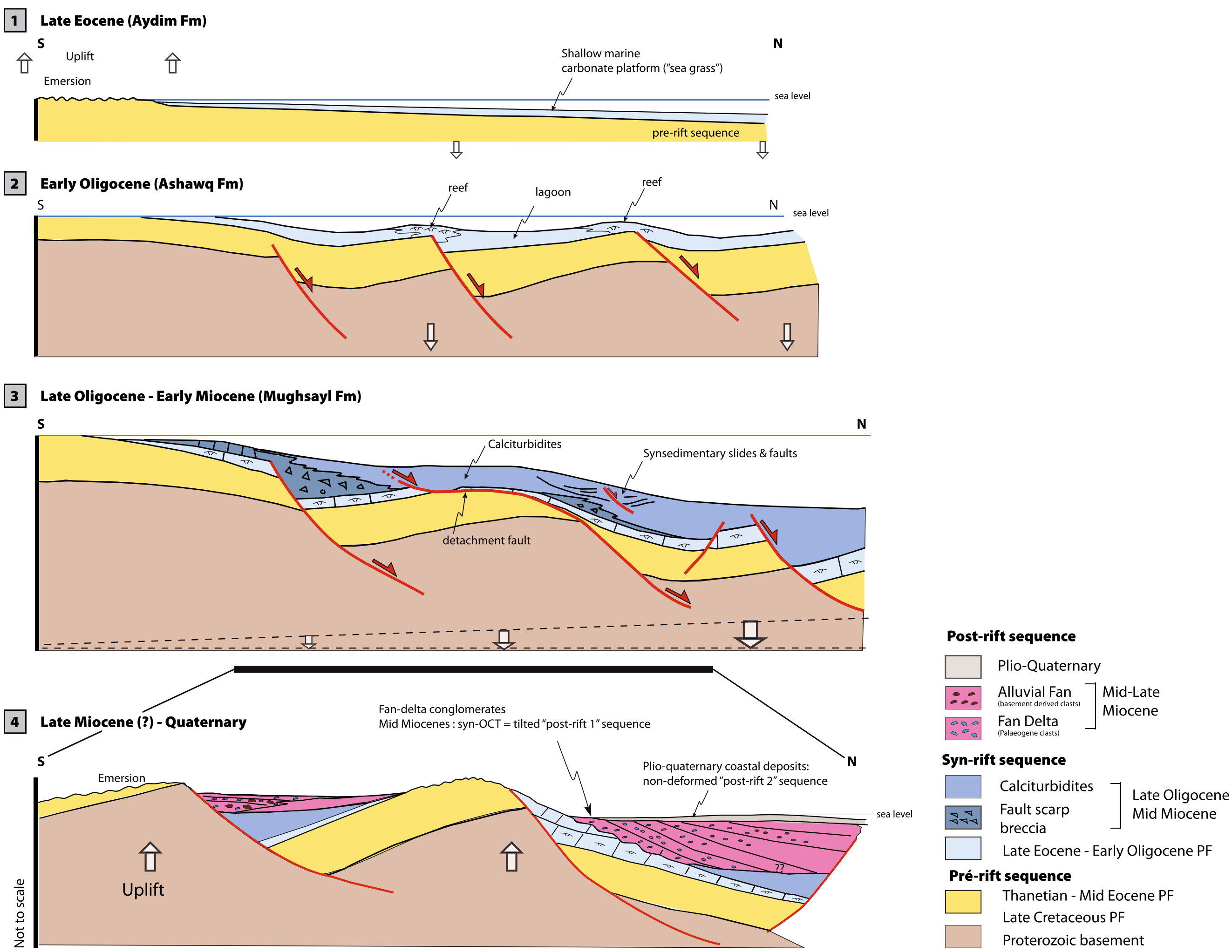


Fig. 19 Tectonic and sedimentary evolution patterns on Socotra Island; synthetic section during syn-rift and post-rift phases. See text for explanations

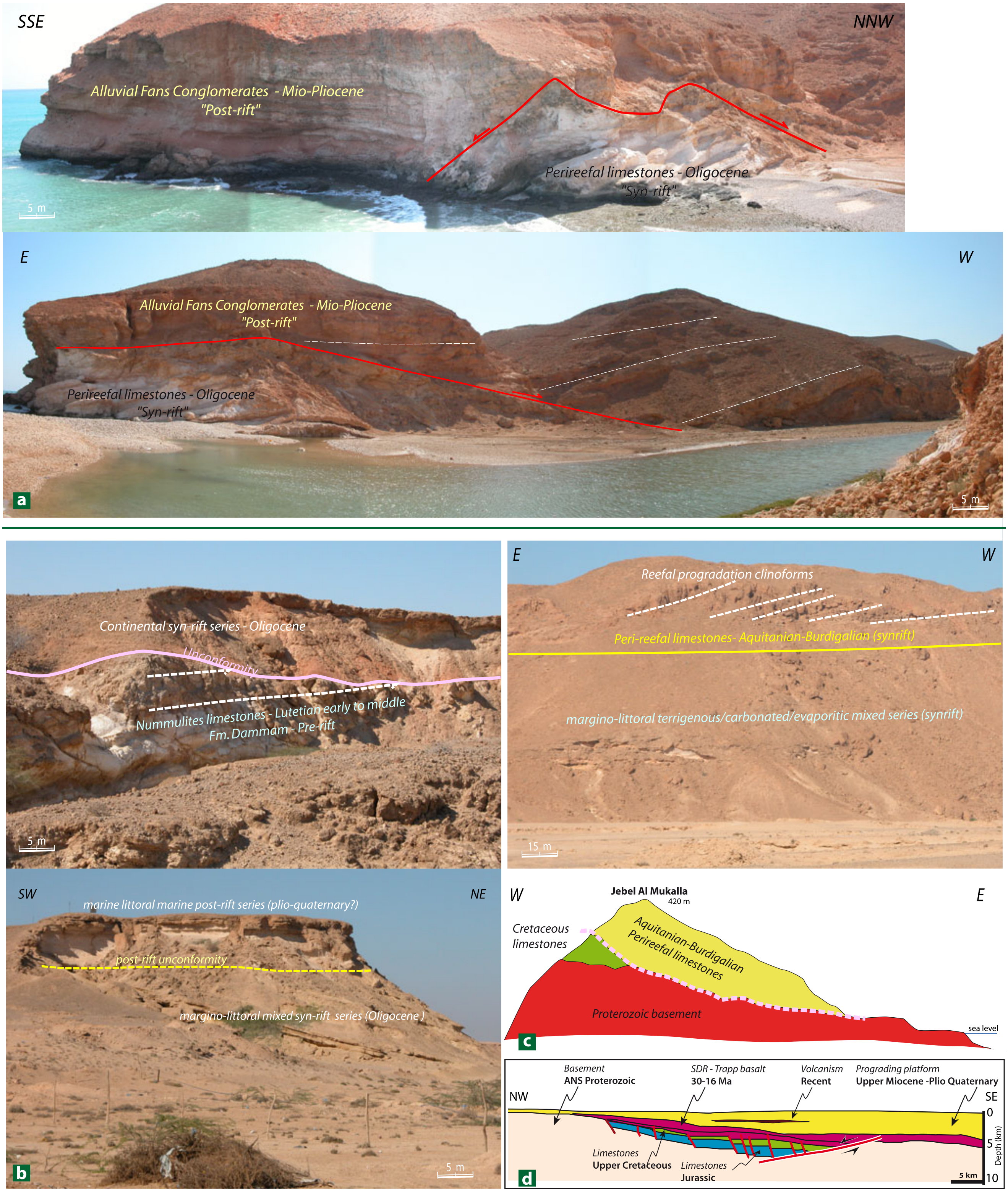


Fig. 20 **a** Stratigraphic and tectonic relationships between the Oligocene syn-rift peri-reefal carbonate deposits and the post-rift continental conglomerates on the southern border of the Qamar basin (Nasht'in). See Fig. 2 for location. **b** *Upper left* Unconformity of syn-rift Oligocene continental deposits on the pre-rift Middle Eocene succession (Dammam Fm.) in the Masilah basin. This unconformity indicates a major phase of uplift and erosion, before the preservation of syn-rift deposits in this domain. *Upper right* Peri-reefal limestone of Aquitanian/Burdigalian age, corresponding to the maximum marine flooding phase of the Masilah basin during the syn-rift stage. *Lower left* Unconformity of (probable) Oligocene syn-rift coastal margin (neritic) deposits with the post-rift coastal deposits (Plio-Quaternary) of the Masilah basin. **c** Section of Jebel Al-Mukalla. **d** Cross-section of the volcanic margin off Aden city (modified from Tard et al. 1991)

Al-Mukalla

In Jebel Mukalla (western edge of the Sayhut basin Fig. 2), reef limestones with Miogypsinidae dating from the Aquitanian/Burdigalian lie directly on the Proterozoic basement and culminate at an elevation of 420 m. This pattern indicates the importance of the uplift and the erosion of certain blocks during the rifting phase, before the major transgression of the Early Miocene. It also indicates a later uplift of several hundred metres, which could have occurred during the syn-rift/post-rift transition (Fig. 20c).

Aden-El Sheikh-Guban

From 30 Ma or even earlier and until to 16 Ma, a fairly typical volcanic margin developed with seaward dipping reflectors (SDR; Tard et al. 1991). SDRs are also observed in the southern part, west and east of the Shukra el Sheik rift zone (Fig. 2). This volcanic activity occurs in the context of the Aden Gulf opening and the SDR prisms make up the syn-rift sequence, comprised of an intercalated shallow-water sediments and subaerial basalts (Fig. 20d).

The ocean–continent transition

This study is based on the bathymetric, seismic reflection and refraction, gravimetric and magnetic data acquired during the cruises of Encens-Sheba (Leroy et al. 2004, 2006), Encens (Leroy et al. 2010b) and Encens-Flux (Leroy et al. 2007; Lucazeau et al. 2008; Figs. 7a and 21). These data allow us to identify the location of the OCT (Leroy et al. 2010b), which belongs to the deep margin domain where: (1) the geophysical and geological characteristics are typical neither of continental nor oceanic crust; (2) a thick horizontal layer of sediments (classified as "syn-OCT sediments") was deposited during the continental break-up over the sedimentary fans that are typical of the syn-rift period and (3) there are no clearly identifiable oceanic magnetic anomalies. The transition from a continental to an oceanic domain corresponds to an increase in lower crust seismic velocity ranging from 6 to 7.6 km s–1. It is also associated with a surface heat flow ranging from 40 to 120 mWm^{-2} and free-air gravimetric anomalies ranging from −40 to 40 mGal (Fig. 22; Leroy et al. 2010b). The formation of the OCT occurred when the continental crust ruptured, leaving only an incomplete continental crust, without any lower and/or upper crust. The OCT therefore extends from this point to the spreading system where oceanic crust is being formed.

The Encens-Sheba zone: between the Alula–Fartak and Socotra–Hadbeen fracture zones

The first-order segment between the Alula–Fartak and Socotra–Hadbeen fracture zones is divided into three second-order segments with different structures and morphologies (Fig. 21).

We first present a tectonic and stratigraphic model of the evolution of the western segment, called Ashawq–Salalah, from rifting to the present day, using the results of pre-stack depth migration seismic reflection profile (Fig. 23; Autin et al. 2010b). Then the second-order segments are presented together with their respective conjugate margins (Figs. 22 and 24; Leroy et al. 2010b).

Structure and evolution of the "Ashawq–Salalah" second-order segment

To interpret the chrono-stratigraphical development, we take account of the existing dating results and observations on land, while the modelling of gravity and wide-angle seismic data allow us to constrain the deep structure of this segment. The proposed evolution for the Ashawq–Salalah segment from rifting to oceanic accretion is also correlated with rocks of the outcropping proximal margin (Roger et al. 1989; Leroy et al. 2007). This correlation is based on the assumption that when onland sedimentation is poorly developed or lacking, offshore depositional sequences (usually detrital, i.e. resulting from erosion) should be very thickly developed on the deep margin, and vice versa. Moreover, the syn-rift successions are likely to be more recent nearer to the distal margin. Based on to these assumptions, and due to the lack of offshore drill core samples, this correlation allows us to date the structures observed offshore (Fig. 23; Autin et al. 2010b). During the syn-rift period (end of Priabonian/Early Burdigalian), the faults reached a depth corresponding to the fragile/ductile boundary (Fig. 23). Lithospheric thinning was very sudden, accommodated by several tilted blocks and syn-rift deposits recording the acceleration of deepening (15–20 km of thinning over a distance of 50–100 km). Fault evolution indicates a localisation of deformation and crustal thinning in the distal margin graben (DIM in Fig. 22a). This extremely thin continental crust conditions the future emplacement of the OCT (Fig. 22), which is narrow (15 km wide). The OCT upper crust is of oceanic type (4.5 km/s), while the lower crust is of continental type (>6.5 km/s; Fig. 22a Watremez et al. 2011).

Coeval with the OCT emplacement (Middle Burdigalian), the uplift associated with a phase of erosion on land could have triggered a submarine slide on the top of the horsts located in the southernmost continental domain. This slide

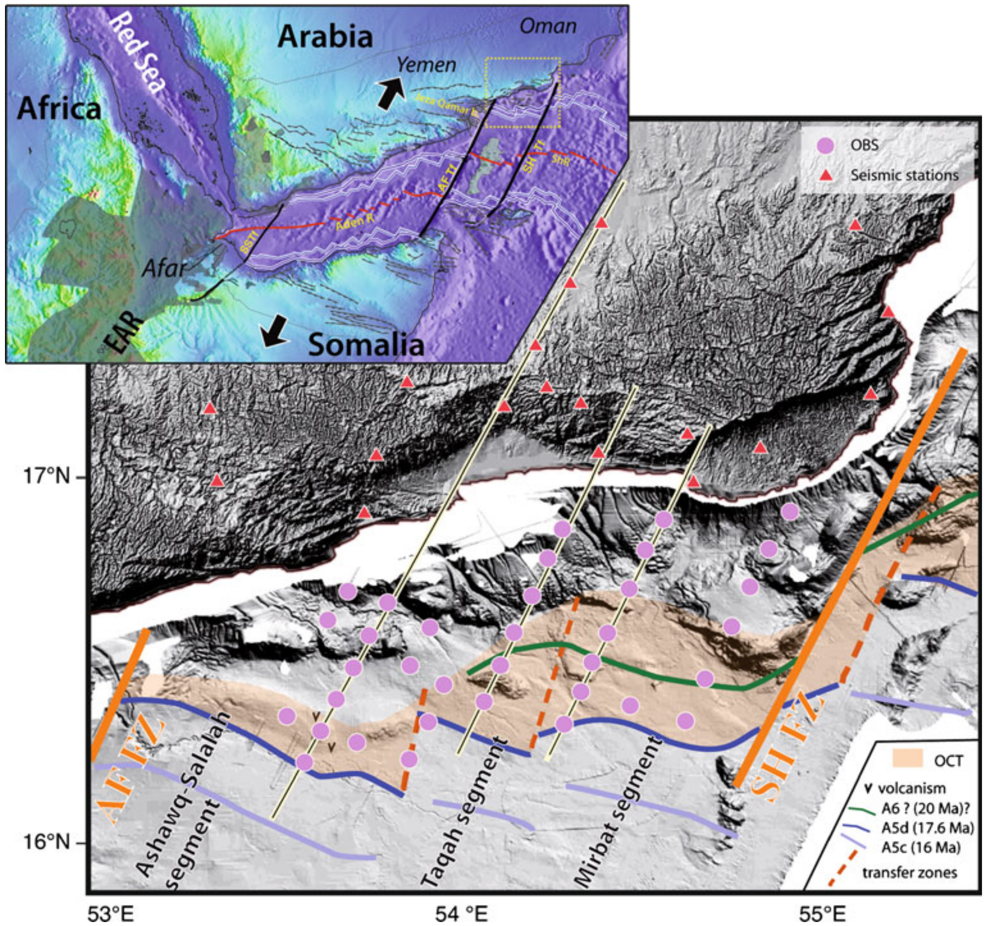


Fig. 21 *Inset* Location of study area. *Grey* shading indicates volcanic activity related to the Afar hotspot. *AFTf* Alula Fartak transform fault, *Aden R* Aden ridge, *ShR* Sheba ridge, *Jeza Qamar b*. inherited basin. The *red line* indicates active oceanic spreading ridge, while the *blue line* indicates the location of isochrones 5c (16 Ma; *light blue*) and isochron 5 days (17.6 Ma, *dark blue*) corresponding to locations of the paleo-ridge axis. Map of the Encens acquisition cruise in the northeast of the Gulf of Aden. Shaded bathymetric map showing the multi-beam bathymetry recordings during the Encens and Encens-Sheba cruises. The Ashawq-Salalah, Taqah and Mirbat segments are separated by *dotted lines* (accommodation zones). The *purple circles* represent the OBS and the *red triangles* the seismic stations used throughout the Encens experiment. The three profiles highlighted in *yellow* are wide-angle seismic profiles; reflection and heat flow data are presented in Fig. 21. *Salmon pink* ocean–continent transition, *AFFZ* Alula Fartak fracture zone, *SHFZ* Socotra Hadbeen fracture zone

may be rooted in the pre-rift evaporitic Rus formation. In the DIM graben, the location of strain shifts southwards, towards the location of the OCT ridge, where the continental break-up will occur, and create a paleo mid-oceanic ridge (Figs. 22a, 23 and 24). We presume that this OCT is composed of, at least partially, of exhumed serpentinized mantle, given that the crust in this zone is extremely thin (a few kilometres; Leroy et al. 2010b; Watremez et al. 2011; Fig. 22a). Towards the east, the oceanic crust thickness decreases from 10 to 5.5 km. Longitudinal thinning is probably related to variations in magmatic inputs along the mid-oceanic ridge developing at that time.

Reflections in the deep crust of the transitional oceanic mantle could represent traces of continentward-directed detachments in the mantle, which first decouple the crust and the mantle during the extension process, and then the serpentinized upper mantle and the non-serpentinized mantle. During the post-rift period [Late Burdigalian (17.6 Ma)], towards the end of its formation, the OCT zone was affected by magmatic activity in the thinnest part of the lithosphere. Such magmatic activity (flows, sills and volcano sedimentary prism), explains the late growth of the OCT ridge (Fig. 23).

In relation to this volcanic activity, a 5-km thick body with an average velocity (7.6–7.8 km/s) can be identified at the interface between the crust and the mantle, under the thinnest part of the margin, the OCT and the oceanic crust. Based on the presence of a young volcano revealed by heat flow measurements (Lucazeau et al. 2008, 2009, 2010) and by multichannel reflection seismics carried out during the Encens cruise (Autin et al. 2010b). We interpret this as an underplated mafic body, or as partly intruded mafic material emplaced during the post-rift phase. The present nature of the OCT could therefore represent a combination of

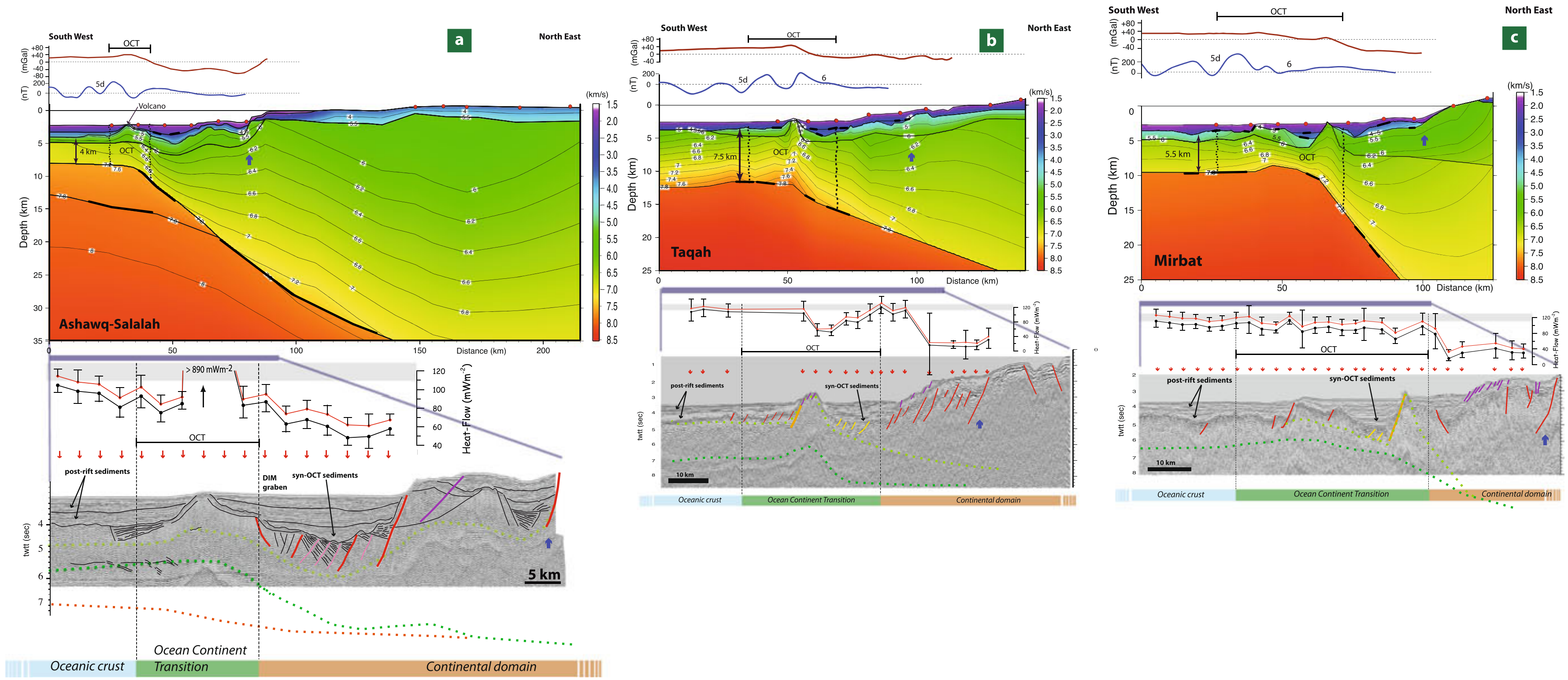


Fig. 22 **a** The Ashawq-Salalah segment during the Encens cruise. *Upper panel* Modelling of the seismic velocity structure. The velocity contours are in colour. They are represented in km s^{-1}. The *red circles* indicate the location of the instruments. OCT indicates the ocean–continent transition, as interpreted and localised between the two *vertical dotted lines*. The models were determined using a modelling method that combines direct/inverse P-wave travel times (chi^2=1.358, Nrais=5,101; highlighted *black lines* represent the wide-angle reflection points). Magnetic anomalies are represented in *blue* with the interpreted anomalies and free-air gravimetric anomalies in *red*. Magnetic and gravimetric data were acquired with the other data presented here. *Central panel* Deep seismic reflection profile with matching heat flow as well as interpretation The *red arrows* indicate the location of heat flow measurements on the seismic profile. *Black circles* indicate uncorrected heat-flow and *red circles* indicate the same after correction for the effect of topography, sedimentation and refraction. *Error bars* represent a confidence interval of 68%. This seismic profile shows the interpretation of the faults (faults active at the beginning of the syn-rift period are in *red*, and those active at the end of the syn-rift period in *green* (Autin et al. 2010b)); faults active during the exhumation phase are in *yellow*, those active during the post-rift phase in *orange* and those associated with active submarine slides in *violet*, along with the main sedimentary units. *Blue arrow* location of the major fault described in the text. *Coloured dotted lines* are derived from the velocity model, converted into two-way travel time; they correspond to the main interfaces underneath the acoustic basement. **b** The Taqah segment. *Upper panel* Seismic velocity model (chi^2=1.919; Nrais=2,689; reflection points, *black lines*), magnetic anomalies (*blue line*) with interpretation, and free-air gravimetric anomaly (*red line*). *Central panel* Heat flow and deep seismic profiles, with interpretation. **c** The Mirbat segment. *Upper panel* Modelling of the seismic velocity structure. Seismic velocity model (chi^2=2.407; Nrais=3,546; reflection points, *black lines*), magnetic anomalies (*blue line*) with interpretation, and free-air gravimetric anomaly (*red line*). *Central panel* Joint profile of deep seismics and heat flow data, with interpretation

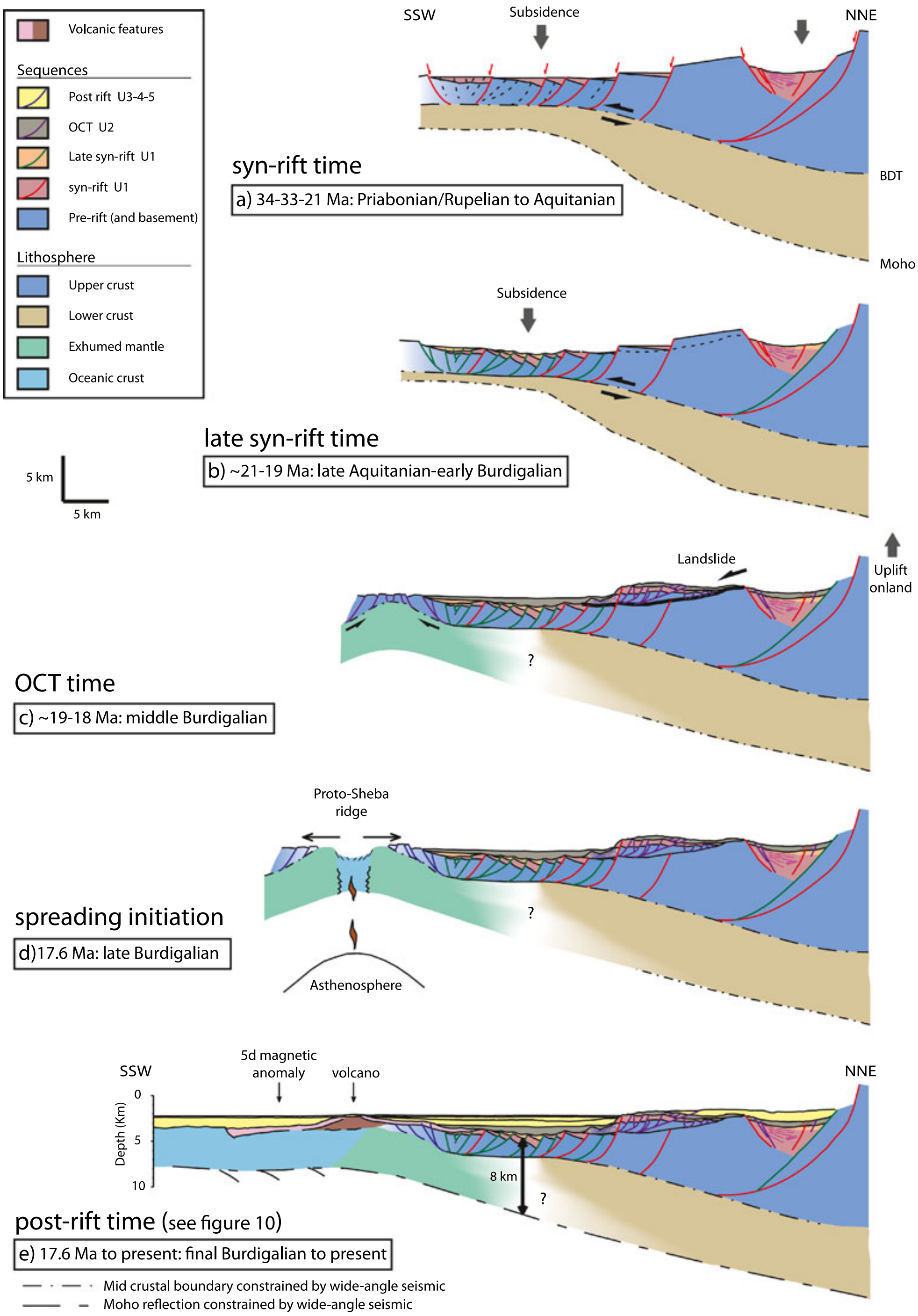


Fig. 23 Structural evolution of the Ashawq-Salalah segment based on the interpretation of the pre-stack depth migration of the ENC34 profile (Autin et al. 2010b). **a** *Syn-rift stage* normal high-amplitude faults, rooted in the brittle/ductile boundary (BDT), cause the pre-rift successions to tilt and delineate syn-rift grabens which are called perched grabens in the North and DIM in the south. **b** *Late syn-rift stage* location of the extensional zone: the continuous thinning of the brittle crust leads to the formation of normal faults that are less closely spaced. **c** *OCT stage* differential vertical movements between the distal and proximal parts of the margin lead to submarine slides of the most distal graben (the southernmost). The OCT may be formed by mantle exhumation; the latter may be partly serpentinized in the region of the continental break-up. Syn-OCT sediments are deposited in the DIM and perched grabens. They pre-date the initiation of oceanic spreading. **d** *Initiation of oceanic spreading* Rifting occurs when the mantle reaches the surface, which thus localises the emplacement of the oceanic accretion system. **e** *Post-rift stage* The ENC34 volcano builds up gradually through several volcanic events associated with resulting phases of uplift. The boundary between the serpentinized mantle and the ductile crust is difficult to image under the DIM graben

serpentinized continental mantle and late magmatic intrusions. Although the upper mantle velocities and densities are lower (7.8–7.9 km/s and 3.10–3.25 g/cm^3) than usual for the upper mantle (8 km/s and 3.3 g/cm^3), such values have nevertheless been observed on other rifted margins (Watremez et al. 2011).

On this segment of the margin, magmatism modifies the initial structural development of the OCT ridge, thus masking the history of the final stages of the continental break-up and formation of the OCT as usually observed on rifted continental margins which little or no magmatic activity. The presence of post-rift volcanism, in the form of melting anomalies, could influence the late-stage evolution of rifted continental margins.

Does the same apply to adjacent segments? Is this magmatism associated with local phenomena or is it related to processes initiated during extension? We will address these questions in the following.

Structure and evolution of rifted continental margin segments between the Alula–Fartak and Socotra–Hadbeen fracture zones (Ashawq–Salalah, Taqah and Mirbat segments)

Continental rifts and continental margins display major variations from one segment to another. This has been attributed to one of several processes such as normal-fault geometry, variable extension along the rift axis, pre-existing lithosphere composition and structural heterogeneity, oblique rifting and the presence or absence of volcanic centres.

In the study area, the non-volcanic oblique margin includes several transform faults with major offsets controlling the first-order segmentation of the mid-oceanic ridge. This segmentation starts during the late-stages of the syn-rift period (e.g. d'Acremont et al. 2010; Leroy et al. 2010b).

Using 2-km-spaced seismic reflection lines and new geophysical and geological observations in the eastern part of the Gulf of Aden, we demonstrate that the segments between the major fracture zones show relevant internal variability. Thus, over a small distance (~10 km), the OCT develops from a narrow and magmatic domain (~15 km) into a broader zone (~50 km wide) where the continental mantle is probably exhumed (Figs. 22 and 24). We consider that two mechanisms can induce the structures observed on the margins of the present-day oceanic accretion system. (1) Oblique rifting close to transform faults shows major offsets and (2) local magma-induced lithosphere weakening, in a setting of subcontinental mantle exhumation. Therefore, we suggest that this variability is due to the distribution of magmatism as well as differences in extension rates in the various sectors of the rift, along an oblique rift system. Magmatism could be associated with the AFFZ, which could stimulate the production of magma along its most recent border (d'Acremont et al. 2010), or with the channelling of material from the Afar hotspot, firstly along the developing OCT, and then along the ridge axis, as assumed by Leroy et al. (2010a). Two low-velocity anomalies are imaged by telesimic tomography in the onland continuation of the Alula–Fartak and Socotra–Hadbeen fracture zones, situated at a depth of 60–200 km. Partial melting (3–6%) at the core of these two negative anomalies is assumed by the data (Basuyau et al. 2010). The presence of this partial melting is interpreted as a possible interaction between the Afar hotspot and the Sheba ridge. However, the joint inversion of gravity and teleseismic data allows us to image only the present-day lithosphere heterogeneities. We therefore have no constraints on the history of the development of such structures (Fig. 25; Basuyau et al. 2010).

Thus, rift obliquity, along with the associated segmentation and interaction with an hotspot, could significantly control the type of rifting, the continental break-up and the evolution of numerous continental margins that display little or no volcanic activity.

Spreading and post-rift history

The geophysical dataset used to study the spreading history of the overall Gulf of Aden is that of research cruises (Tadjouraden (Audin et al. 2004), Encens-Sheba (Leroy et al. 2004), Encens (Leroy et al. 2010a), Encens-Flux (Lucazeau et al. 2010)) and unpublished industrial data. The magnetic anomalies has been identified jointly with the seismic reflection profiles in order to accurate the location of the first oceanic crust formed. The magnetic anomalies identification method is described in Leroy et al. (2000).

The "Encens-Sheba" zone

Between the Alula–Fartak and the Socotra–Hadbeen fracture zones, a melting anomaly developed at latitude 13°20′ N, causing a oceanic ridge jump that occurred just before A5 (10 Ma, Fig. 26; d'Acremont et al. 2010). Moreover, tectonic changes (between 9 and 6 Ma) induced the formation of transform faults and pull-apart basins are formed within the Alula–Fartak and Socotra–Hadbeen major transform zones (Fig. 26; d'Acremont et al. 2010).

Multibeam bathymetry, gravimetric and magnetic data show that the structures and contrasted seafloor depths between the eastern and western domains were initiated at least 10 Ma ago, during A5. These variations persist throughout the history of oceanic spreading with the emplacement of propagators towards the east (PF; Fig. 26). Moreover, an intensification of volcanic activity

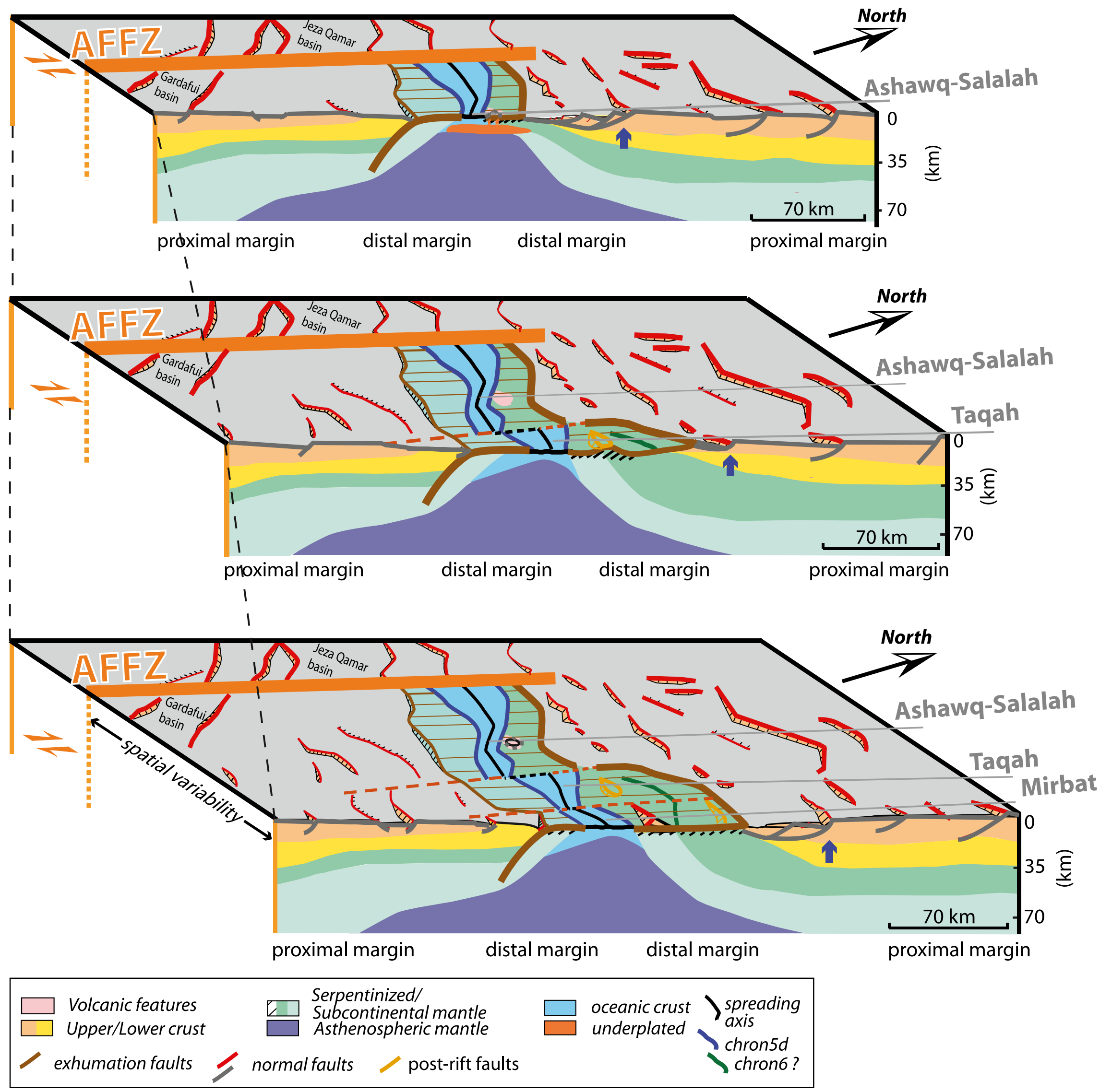


Fig. 24 Schematic representation of reconstructed segments on the southern conjugate margin, showing tectonic style before oceanic spreading along the Ashawq–Salalah profile, near the Alula Fartak (AFFZ), Taqah and Mirbat profiles. Three-dimensional block diagrams showing location of the AFFZ with respect to the various segments and the distribution of faults near the section line. The sediments are not represented

is observed, with a paroxysm around 11 Ma ago (Fig. 26). The outlines of the discontinuities and magnetic anomalies, as well as the variations in crustal thickness as calculated from gravimetric data, suggest that the OCT and oceanic crust segments are controlled by the segmentation of the conjugate margins. This segmentation has changed over time, first following the continental break-up, the spreading now taking place to the east of segment 1 (Fig. 26) following a ridge jump. Both phenomena result from the intensification and focusing of the magmatic activity linked to the presence of a melting anomaly, and the proximity of both the Alula–Fartak transform fault (which shows a 180-km horizontal offset) and the continental margins.

Eleven million years ago, between the A5c and A5 magnetic anomalies, oceanic spreading was asymmetrical. This can be explained by a ridge jump of the paleo-segment 1, toward the south in the direction of the melting anomaly. From 9 Ma, the propagators and ridge jumps ceased as a result of global kinematic changes and of decreasing influence of the mid-ocean ridge melting anomaly. This was followed by the development of a new transform fault zone (Socotra TF; Fig. 26) and of segment 2 (Fig. 26). The geological consequences of recent kinematic reorganisation induced a clockwise rotation of the stress field, a change in orientation of the AF and SH transform faults and their opening along a WNW-ESE trend. We suggest that a new

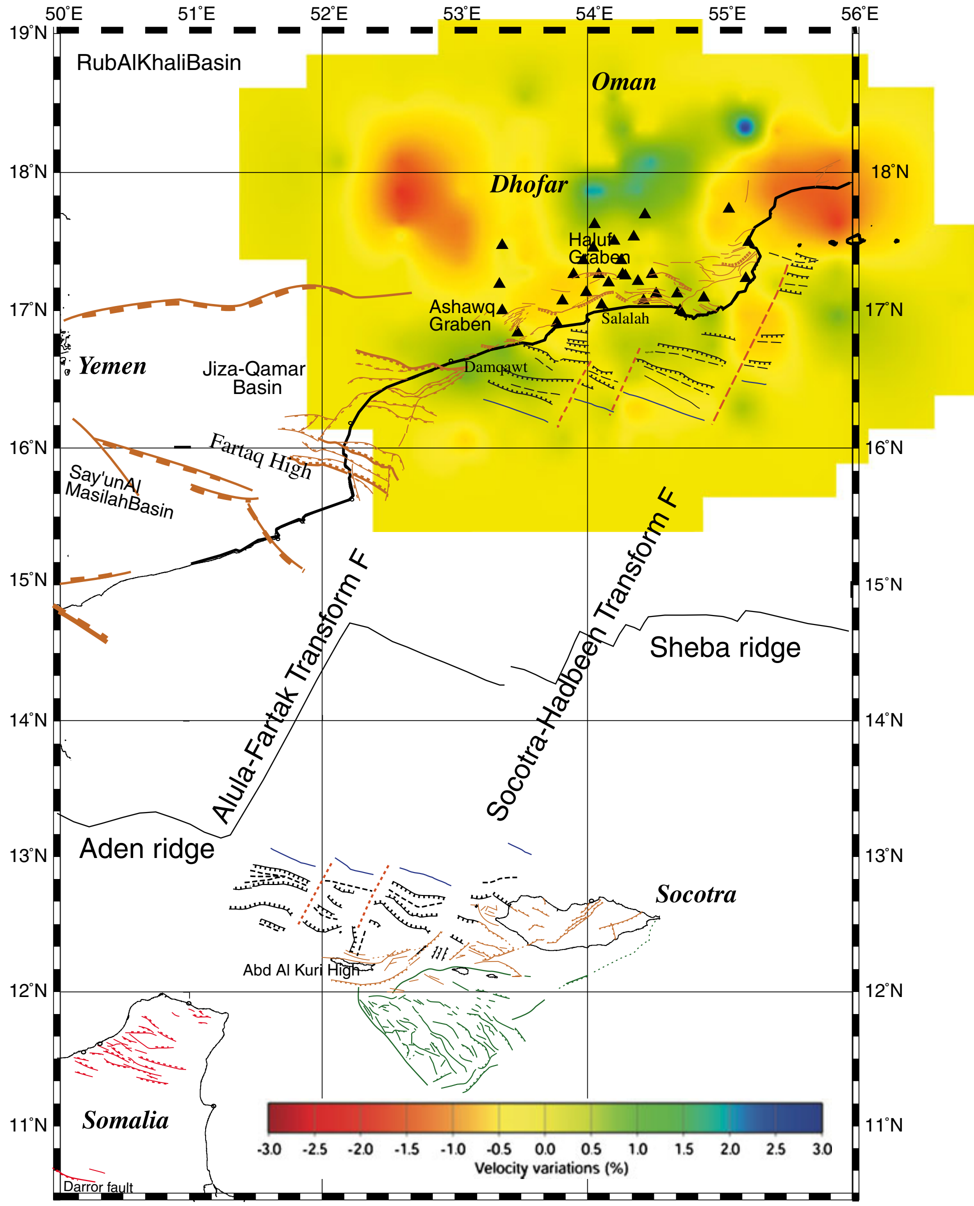


Fig. 25 Modelling of P-wave velocity discontinuities, at 170 km depth (Basuyau et al. 2010). Tectonic structures extracted from d'Acremont et al. (2005). Note that slow zones are localised along the prolongation of the Alula–Fartak and Socotra–Hadbeen fracture zones, underneath the province of Dhofar at 60–200 km depth. These velocity anomalies could correspond to deep partial melting zones with 2–6% of partial melting

spreading centre developed in the AF transform fault zone, between the Aden and Sheba ridges (Fig. 26). Therefore, it seems that the tectonic and magmatic evolution of the mid-oceanic ridge differs from one segment to another adjacent (d'Acremont et al. 2010; Leroy et al. 2010a).

Aden and Sheba ridge system

By pooling marine geophysical data from academic and oil industry surveys, we can obtain a high-resolution magnetic coverage of the area near the coasts of Yemen and Somalia. By analysis of the magnetic data, combined with reflection seismics and gravimetric profiles, we are able to determine the precise location of the OCT throughout the Gulf of Aden using the same definition as for the Dhofar northern margin (Fig. 27). Data interpretation allows us to reconstruct the history of the opening of the Gulf of Aden at 17.6 Ma, without any westward propagation into the African continent, from the edge of this craton to the Shukra el Sheik fracture zone, (Fig. 27). This age corresponds to anomaly A5d, which is identifiable from the edge of the Arabian–African craton in the east, to the Shukra-el-Sheik fracture zone in the west (Figs. 2 and 27). These observations imply a continuous opening, over a distance of more than 1,100 km, which took place 18 Ma ago.

The area located west of the Owen fracture zone (Fig. 2; east of the Arabia–India–Somalia triple point) was studied by Fournier et al. (2008). In this region, south of the Owen basin the mid-oceanic ridge could have been initiated 20 Ma

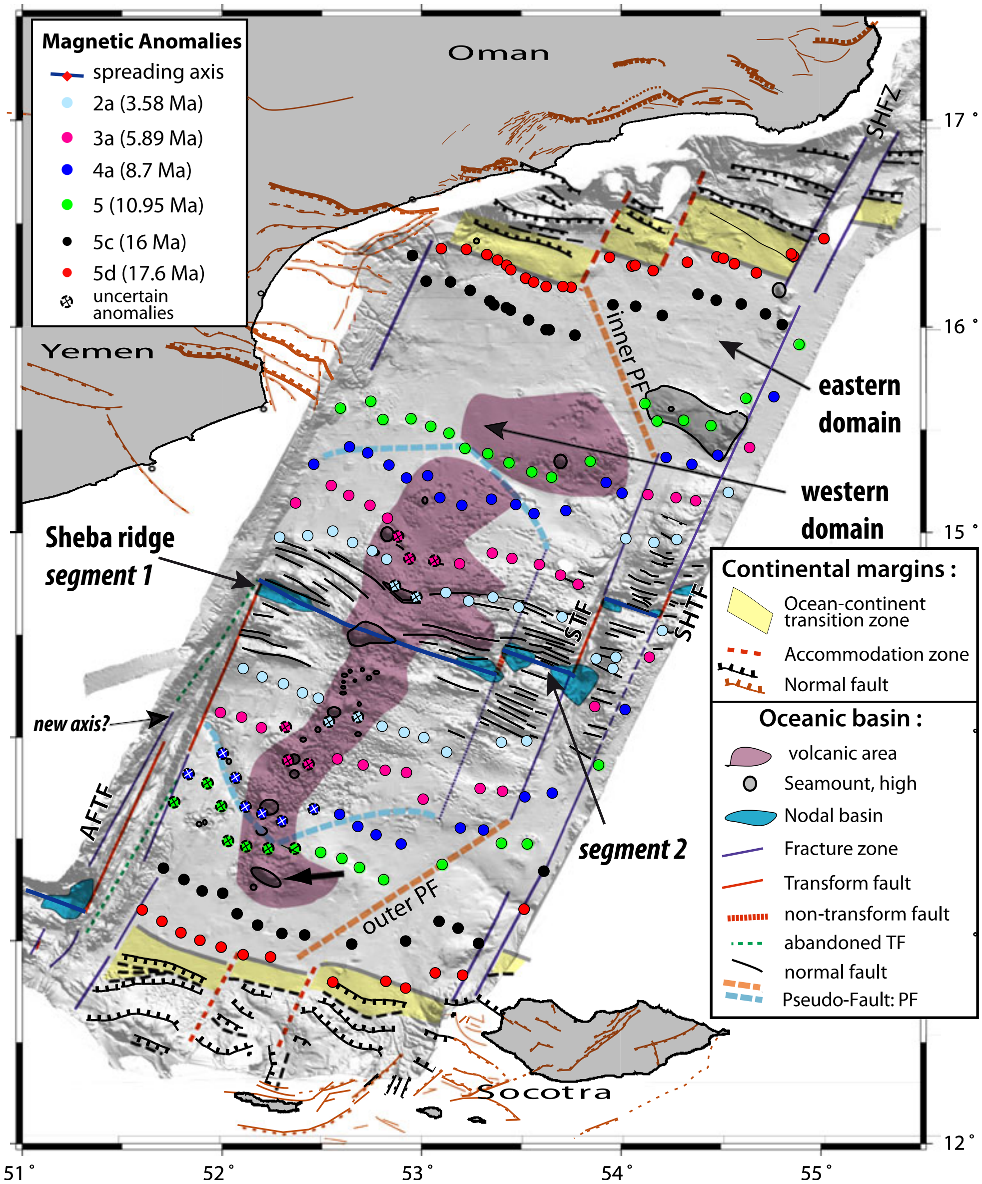


Fig. 26 Structural pattern and magnetic anomalies identified during this study (d'Acremont et al. 2010), represented on *shaded* bathymetric map of the Encens-Sheba cruise (Leroy et al. 2004). The identified magnetic anomalies are represented by *circled crosses*. The interpretation of continental margins is after d'Acremont et al. (2005) for offshore data; and after Beydoun and Bichan (1969), Platel and Roger (1989), Platel et al. (1992b), Birse et al. (1997), Brannan et al. (1997), and Bellahsen et al. (2006), for inland data. *AFTF* Alula–Fartak transform fault, *STF* Socotra transform fault, *SHTF* Socotra–Hadbeen transform fault, and *SHFZ* Socotra–Hadbeen fracture zone. *Black arrow* indicates position of the melting anomaly (d'Acremont et al. 2010; Leroy et al. 2010a, b)

ago, in an old oceanic lithosphere dated probably Late Cretaceous (Platel et al. 1992a) to Tertiary (Edwards et al. 2000). The rheology should be different compared to that of the African–Arabian craton. Since the orientation of the main structures and spreading-type appear rather similar to those found in the Carlsberg ridge, we consider that this region is probably a part of the Carlsberg ridge system different from the Gulf of Aden ridge systems. We propose that the Gulf of Aden sensu stricto is located in the west of the boundary of the African–Arabian shelf (Fig. 27).

Figure 27 shows clearly that the Alula–Fartak transform zone constitutes a boundary between two domains showing different behaviours. However, it does not represent a zone where ocean floor spreading has ceased, as previously suggested by certain authors (Manighetti et al. 1997; Huchon and Khanbari 2003; Fournier et al. 2010).

In terms of width, the ocean–continent transition shows a variation from east to west. The OCT is 100 km wide in the east, where the presence of serpentinized exhumed mantle is assumed to be located (Leroy et al. 2010b), and where the Sheba ridge is developing north of the reactivated Gardafui basin (Figs. 2 and 27 see reconstruction on Fig. 28). In the western part of the Gulf, towards the middle, the OCT is narrow (15 km), and the Aden ridge is developing to the south of the reactivated Jeza-Qamar basin. However, the OCT seems to cut across the reactivated syn-rift basins in the west of AFFZ (Masilah-Darror, Balhaf-Berbera Fig. 28). Nevertheless, as shown on Fig. 27, the orientation of the OCT boundaries in the central part of the Gulf is concordant with the structural trend of the syn-rift basins (N100° E). Between the Xiis-Al-Mukalla and Shukra-El-Sheik fracture zones, the continental mar-

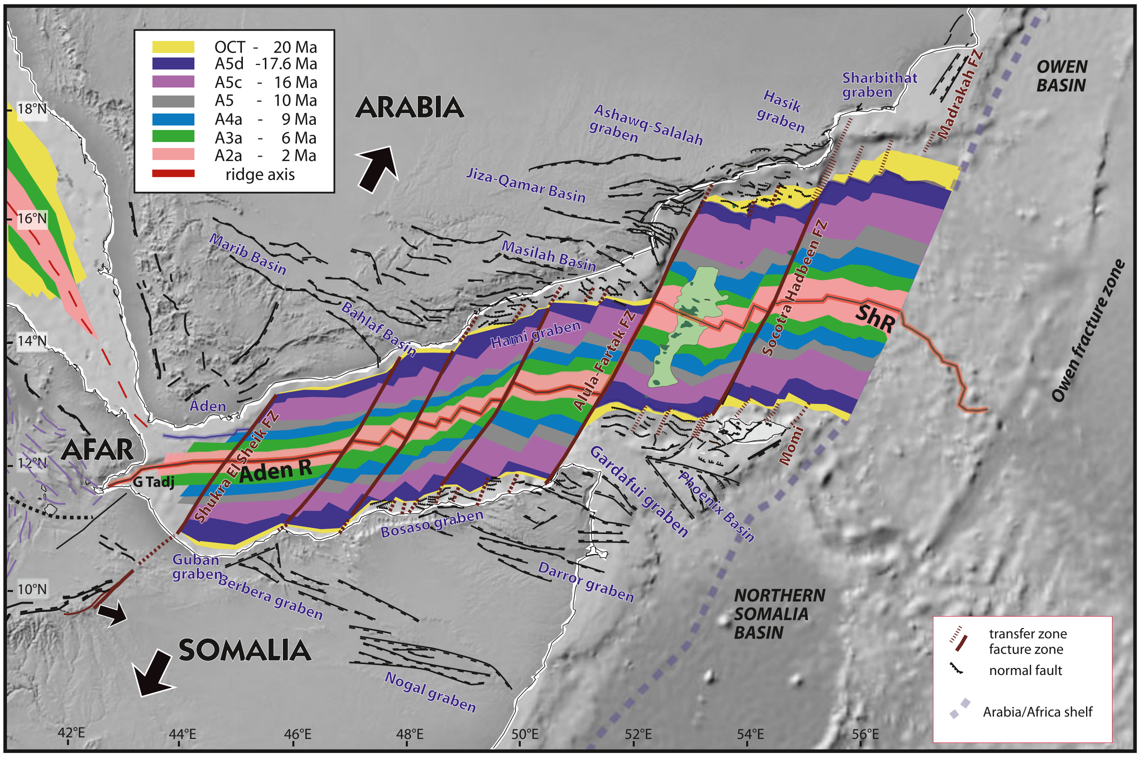


Fig. 27 Map of the Gulf of Aden isochrones, showing the synchronous opening of the Gulf of Aden 17.6 Ma ago (5-day magnetic anomaly) from the edge of the African–Arabian craton in the east to the Shukra-el-Sheikh fracture zone

gins appear to be predominantly influenced by syn-rift volcanism (Fig. 28). A rapid transition is observed with the Burdigalian oceanic crust (18 Ma).

In the westernmost part of the Gulf of Aden, west of the Shukra-El-Sheik fracture zone, the continental margin appears clearly volcanic (Tard et al. 1991). The boundary

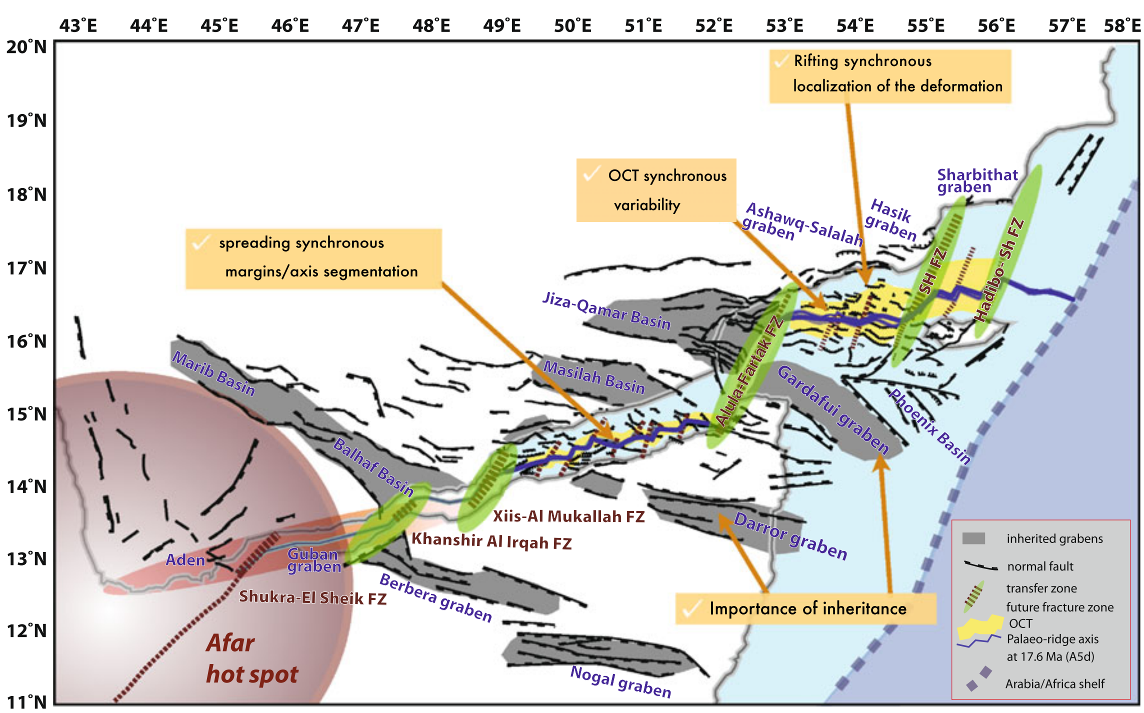


Fig. 28 Reconstruction map of the Gulf of Aden, just before oceanic spreading (18 Ma)

with oceanic crust, dated at 9–6 Ma (Fig. 27), appears to have occurred abruptly, as in the case for many volcanic margins (Gernigon et al. 2004; Geoffroy 2005). However, the youngest lava flows of the SDRs yield an age of 16 Ma (Tard et al. 1991), so there could be a gap between the end of volcanic margin formation and ocean accretion in this zone.

While the influence of the Afar hotspot is unequivocal in the west of the Gulf of Aden during rifting, it also appears to be predominant between the Shukra-el-Sheik and Xiis-Al-Mukalla fracture zones. In the non-volcanic margins located east of the Xiis-Al-Mukalla fracture zone, the Afar hotspot influence could have become active as early as the end of continental rifting, after emplacement of the OCT (Lucazeau et al. 2008; Leroy et al. 2010a, b), thus favouring localisation of the continental break-up (e.g. Hopper et al. 1992; Buck 2004). The hotspot influence persists to the present day in the manner of an interaction with the mid-oceanic ridge (Leroy et al. 2010a).

Discussion and conclusion

The contemporariness of mantle plume, continental flood basalts and rifting has long been noted (Morgan 1972; Courtillot et al. 1999; Geoffroy 2005). Detailed studies have been carried out on the role played by the ascent of material derived from a plume beneath the continental lithosphere. In triple junction zones the lithospheric extension is preferentially interpreted by an actively driving due to the convective plume own energy (e.g. Sengör and Burke 1978; Bott 1982). Other studies suggest a combination of active and passive processes. Nevertheless, plumes activity often guides the location of the initial rupture, requiring additional stress concentration on the plate boundaries (Hill 1991; Courtillot et al. 1999). Several mechanisms have been evoked, including the effect of slab-pull in subduction zones and the ridge-push of oceanic ridges and their related dynamic topography, but the ultimate mechanism controlling plate movements remains still under debate. The geometry of the continental break-up can also be influenced by volcanic events and by the numerous pre-existing varied wavelengths heterogeneities occurring in the lithosphere. We have a fair opportunity for analysing the history of rifting processes as a whole, in the Gulf of Aden establishing a chronology between the plume activity and the plate dislocation. Moreover, the segments of rifts/continental and margins/oceanic ridges where sedimentary rates are rather low in this area are easy to correlate. The newly acquired data and the multidisciplinary onshore–offshore studies, which have been conducted in the Gulf of Aden allow us to precise the tectonic history.

Inherited structures, reactivation and localisation of the deformation

The pre-existing features, either magmatic, tectonic or sedimentary, reactivated in the Gulf of Aden have influenced the rifting present development. Indeed, the location of the syn-rift basins appears to be controlled, or fairly guided, by pre-existing structures. From the Karoo [Permo-Trias] up to the Burdigalian oceanic crust formation, the region of the Gulf of Aden was affected by a discontinuous tectonic activity, which controlled the basins opening.

During the Jurassic, the future Gulf of Aden was a high zone and was bordered by sedimented margins (Fig. 4). At the entire Arabian plate scale (Fig. 4), during the Cretaceous, the E–W to SE–NW graben and associated normal faults orientation indicate a regional N–S to NE–SW extension. The development of these intraplate graben shows that tectonic stress has been transmitted and partitioned throughout the entire Arabian-African plate Consequently, this also indicate that the lithosphere did not have a purely rigid comportment during extension. During the Cretaceous period, no volcanism was apparently associated with this extension, and this argues in favour of a rifting mechanism driven by remote stress fields (Turcotte and Emerman 1983). The emplacement of the Afar hotspot, expressed a maximum activity 30 Ma ago, it has changed the mechanical properties of the lithosphere by heating up and weakening the Afro-Arabian continental lithosphere. This Afar-focused process possibly triggered the migration of the extensional regime onto the rifts of the Red Sea and the Gulf of Aden. We may consider the plume as one of the dominant triggrering factors responsible for the separation of the Arabian plate from Africa (Bott 1982; Malkin and Shemenda 1991; Cloetingh et al. 1995; Zeyen et al. 1997; Courtillot et al. 1999; Jolivet and Faccenna 2000; Bellahsen et al. 2003).

During the various extensional episodes, new faults were formed, and then reactivated, while already active ones developed continuously in response to local stress fields. The latest rifting stage, leading to oceanic spreading, appeared not to have been similarly guided by the reactivation of pre-existing tectonic features in the Gulf everywhere. We can definitely evoke a structural influence of the pre-existing tectonic features for the eastern part of the Gulf (asymmetry of the rifted zone clearly observable on a map; see Figs. 4 and 28). In the central part it is more difficult to find any evidence of tectonic inheritance.

The Sheba Ridge developed to the north of the Gardafui basin (e.g. d'Acremont et al. 2006), while the Aden Ridge developed between the two inherited basins of Jeza-Qamar/Masilah in the north and Darror/Bosaso in the south (Fig. 28). This hypothesis have been tested in oblique-rifting analogue models, and show that a pre-existing lithospheric weakness is not required to localise the Oligocene grabens along the oblique rift (Autin et al. 2010a). The presence of inherited structures appears to control only the location of the continental break-up (Figs. 4 and 28), which will not occur in the centre of the inherited basins (d'Acremont et al. 2005). It is also the case in well-studied northern Atlantic Margin (Cowie et al. 2005) and basin migration during rifting is a mechanism proposed for slow lithospheric extension (van Wijk and Cloetingh 2002).

The geometry of the new faults follows the inherited basins trends, and their segmentation increase eastwards. At the west of the AFFZ (Figs. 6 and 27), the fault wave-lengths are shorter. This mechanism is also related to the deformation style of numerous rifts, for example in Thailand (Morley et al. 2007). Analogue modelling provide us with an overview of the possible evolution of conjugate margins in the Gulf of Aden which have developed in the framework of an oblique extension (Autin et al. 2010a). Pre-existing and localised lithospheric weakness parallel to the rift is definitely not a pre-requisite for the opening of the eastern part of the Gulf. In contrary, a control by lithospheric weakness zones could be invoked west of the Alula–Fartak fracture zone. This process could be related to an assumed channelling of mantle plume hot material from the Afar hotspot coeval with the OCT formation (Leroy et al. 2010a).

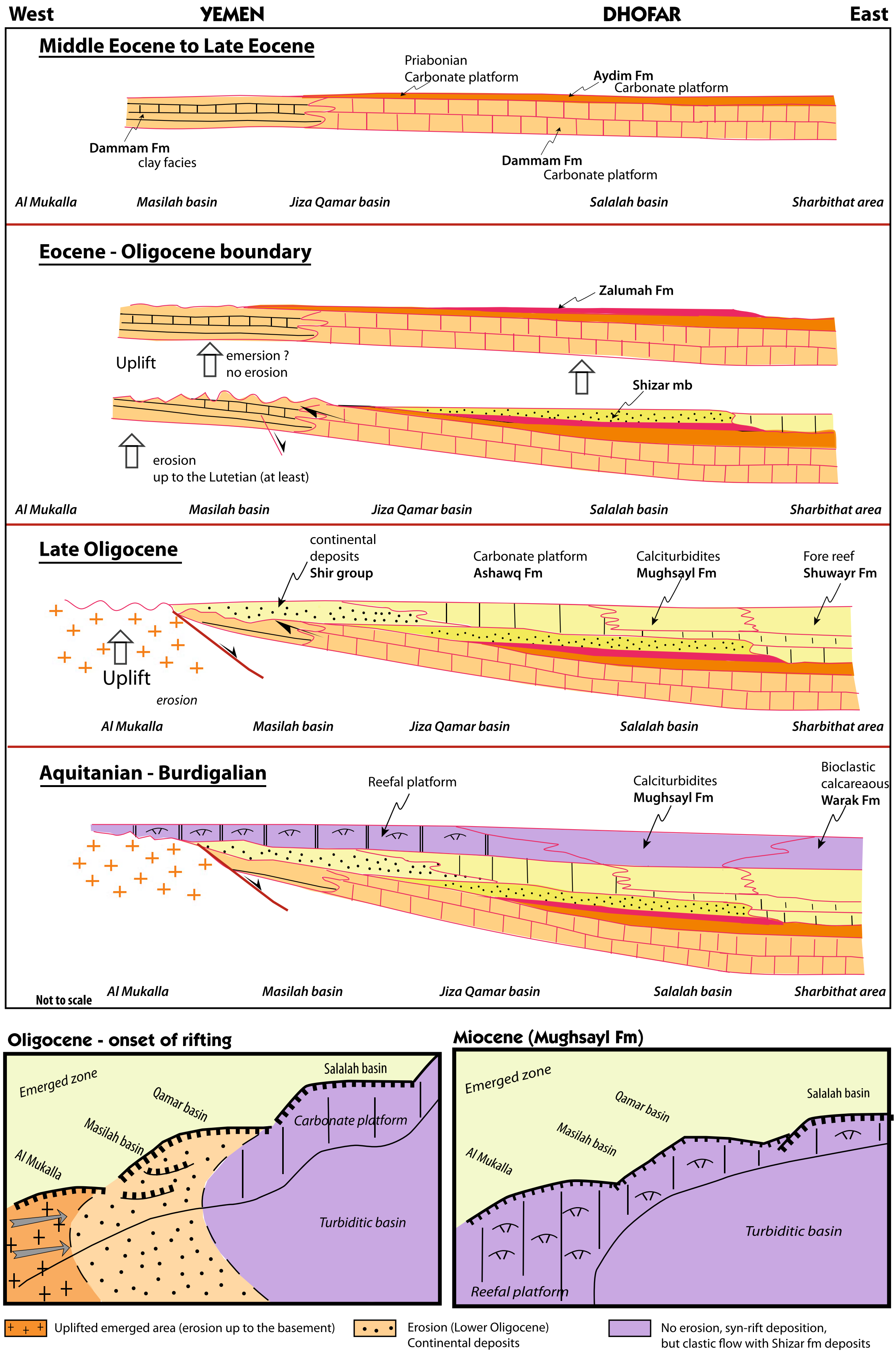
In addition, a detailed study of the Gulf of Aden Precambrian basement shows that the location and directions of the main transform faults are most probably inherited from the Neo-Proterozoic. Thus, both the HTZ Socotra Island and the Socotra–Hadbeen FZ might be inherited from the Proterozoic up to the Oligocene rifting time (Denèle et al. 2009; Leroy 2010a; Denèle et al. 2011). Consequently a preferential orientation of olivine crystals caused by the previous orogenic phases is responsible for a mantle anisotropy, which is of great impact on the rift zones and transform zones localisation. While rift initiation is guided by this anisotropy, stresses acting at the plate boundaries have been required to allow the continuation of the opening process (Tommasi et al. 2009). As proposed by Thomas (2006) the ancient suture zones represent thus a non-negligible inherited guide for the localisation of new continental rifts and/or transfer faults.

The timing of rifting and spreading

Since the emplacement of the Afar hotspot, the western domain has been under its direct influence. Since, at least, the Early Oligocene up to 16 Ma ago (Burdigalian) the margin is volcanic (Tard et al. 1991), as indicated by the SDRs (Figs. 27 and 28). The central domain has been uplifted and eroded from the beginning of rifting onwards, whereas the eastern domain was subject to continuous sedimentation. During the Oligocene, tectonic subsidence impacted the entire domain and favoured the preservation of syn-rift sedimentary successions. Syn-rift series display clear paleo-environment variations, expressed from the west by continental terrigeneous and evaporitic systems, and to the east by a carbonate shelf and turbiditic sedimentation. This east/west polarity is compatible with the commonly accepted idea of rift propagation (Bosworth et al. 2005; Fournier et al. 2010). As all the syn-rift series has been dated of the same age (biostratigraphic dating), the propagation under a mantle plume influence is not necessary (Fig. 29). The rapid tectonic subsidence increase during rifting is either expressed by the development of open marine turbiditic basins (Mughsayl formation) in the east, or by the development of evaporitic facies, firstly lacustrine, then finally marine, to the west (Fig. 29). The marine flooding maximum occurred during Early Miocene time in the western domain, whereas deep facies persisted in the eastern domain.

During the transition between the syn-rift and post-rift stages, the entire margin has been uplifted. These observations evidenced field elements for along-strike comparison of the northern margin during the rifting process (Fig. 28). In the western domain, the polyphased uplift, is responsible for the erosion and clastic supply on the margins, which can be correlated with the dynamics of the Afar hotspot. In the eastern domain, the complete denudation of the basement in the proximal parts of the margins occurred extremely rapidly during the formation of the OCT, and is coeval with the sag phase in the distal part of the margins (Autin et al. 2010b; Leroy et al. 2010b). Correlations between onshore and offshore deposits show that the proximal margins have been intensively eroded, delivering sediments in the OCT zone. Landslides can be frequently observed, both onshore and offshore, outlining the amplitude and velocity of the vertical movements (e.g. Autin et al. 2010b). Therefore, arguing that the syn-rift successions along the Gulf of Aden are coeval, irrespective on whether they are volcanic, volcano-sedimentary or sedimentary, we suggest that the continental rifting in the Gulf of Aden is synchronous, but diversely expressed during 15 Ma (from 35 Ma up to 20–18 Ma, which is the age of the oceanic spreading onset; Fig. 27).

At the end of the rifting period, the continental break-up resulted respectively in the creation of two Ridges corresponding to the domains in the Gulf of Aden, the Sheba oceanic Ridge in the east and the Aden Ridge in the west (Fig. 27). All these events occurred at the same period around 18 Ma, throughout the Gulf up to the Shukra El-Sheik fracture zone. As we already suggested, the hotspot activity had a strong impact on the lithosphere creating weakness zone at lithospheric scale (Bellahsen et al. 2003; d'Acremont et al. 2003). The thermo-mechanical effect of



West
YEMEN
DHOFAR
East
Middle Eocene to Late Eocene
Priabonian
Carbonate platform
Aydim Fm
Carbonate platform
Dammam Fm
clay facies
Dammam Fm
Carbonate platform
Al Mukalla
Masilah basin
Jiza Qamar basin
Salalah basin
Sharbithat area
Eocene - Oligocene boundary
Zalumah Fm
Uplift
emersion ?
no erosion
Shizar mb
erosion
up to the Lutetian (at least)
Al Mukalla
Masilah basin
Jiza Qamar basin
Salalah basin
Sharbithat area
Late Oligocene
continental
deposits
Shir group
Carbonate platform
Ashawq Fm
Calciturbidites
Mughsayl Fm
Fore reef
Shuwayr Fm
Uplift
erosion
Al Mukalla
Masilah basin
Jiza Qamar basin
Salalah basin
Sharbithat area
Aquitanian - Burdigalian
Reefal platform
Calciturbidites
Mughsayl Fm
Bioclastic
calcareaous
Warak Fm
Not to scale
Al Mukalla
Masilah basin
Jiza Qamar basin
Salalah basin
Sharbithat area
Oligocene - onset of rifting
Emerged zone
Salalah basin
Qamar basin
Masilah basin
Al Mukalla
Carbonate platform
Turbiditic basin
Miocene (Mughsayl Fm)
Emerged zone
Qamar basin
Salalah basin
Masilah basin
Al Mukalla
Turbiditic basin
Reefal platform
Uplifted emerged area (erosion up to the basement)
Erosion (Lower Oligocene)
Continental deposits
No erosion, syn-rift deposition,
but clastic flow with Shizar fm deposits

Fig. 29 Longitudinal evolution of sedimentary systems and erosion zones along the northern margin of the Gulf of Aden, from Al Mukalla (Yemen) in the west, to Sharbithat (Oman) in the east. *Lower panel* Schematic paleogeographic map during the syn-rift stage

the hot spot effect created strong heterogeneities and had a significant influence on the opening of the Gulf of Aden sensu stricto (Fig. 27). The Afar-related weakness zone does not allow any propagation. The opening occurs over more than 1,100 km large in the zone Gulf of Aden where the continental lithosphere finally break-up (Figs. 27 and 28). A similar and simple oceanic opening stage, without propagation, is also observed in Southern Atlantic along with the 2,000 km long southern segment (Moulin et al. 2010; Blaich et al. 2011).

Guided by the segmentation of the margin and different processes of subsidence, which increases eastward, the segmentation of the mid-oceanic ridge evolved extremely rapidly. After 1 Ma of activity, the mid-oceanic ridge segmentation follows to oceanic accretion processes, which is related to the global kinematic changes as well as the occurrence of volcanic episodes along the axis of one of the segments and off-axis. This explains how kinematic regime changes can modified the architecture of both Ridges and transform zones during the oceanic opening history, and are responsible either for the abandoning of some of the transform faults (Momi-Madrakah for example; Figs. 2 and 6) or for the initiation of new ones (STF; Fig. 26). One important observation is that only the first-order segmentation persisted all along the tectonic history.

Between 9 and 6 Ma (A4a and A3a), a major plate reorganisation occurred in the eastern part of the Gulf (Fournier et al. 2008; d'Acremont et al. 2010). In its western part, oceanic accretion started west of the SESFZ at the same time (between 10 and 6 Ma; Fig. 27). The oceanic spreading asymmetry between the AFFZ and SHFZ fracture zones is then explained by the ridge jumps induced 10 Ma ago by the occurence of a melting anomaly associated with major off-axis volcanic activity (d'Acremont et al. 2010; Leroy et al. 2010a). Mid-oceanic ridge jumps are associated with magmatism and have been observed near many hot spots in the world (e.g. Mittelstaedt et al. 2011).

Post-rift activity

Post-rift volcanic activity is closely related to melting anomalies leading to intense volcanic activity developed since 10 Ma between the AFFZ and SHFZ fracture zones. Probably initiated during the continental break-up (about 20 Ma), this episode persists up to the present-day. Tele-seismic tomography provides additional evidence for the presence of hot material at the base of the lithosphere (Basuyau et al. 2010). Low-velocity zones correlated with partial melting zonation have been identified in the prolongation of the AFFZ and SHFZ fracture zones. Recent volcanic activity has also been identified onshore in Yemen and Somalia, as well as on the Aden Ridge (e.g. Leroy et al. 2010a). The above evidence suggest an influence of the Afar hotspot in the Gulf of Aden reaching as far east as the Alula–Fartak transform fault, thanks to the hot material channelling derived from the Afar plume. This material has been conveyed along tectonic corridors such as Aden–Sheba oceanic ridges system and the Alula–Fartak transform zone, which exhibit a major offset (more than 180 km horizontal offset). This model of interaction between the mantle plume and the oceanic ridge, involving channelling processes, implies a lateral extension of the hotspot influence to a longitude of at least 54° E. Thus, we can account for a consequent magma supply to the off-axis melting anomaly, the abnormal volcanic activity in segment 1 of the Encens-Sheba zone and the volcano located on the OCT which remained active until very recently (Lucazeau et al. 2009; Autin et al. 2010b; d'Acremont et al. 2010; Leroy et al. 2010a).

By modelling the high heat flow values measured on the OCT and the sharp transition with the continental domain expressed by low heat flow values, we can infer the presence of a thermal anomaly, which persists after the continental break-up (Lucazeau et al. 2008; Lucazeau et al. 2010). This thermal anomaly, located within the upper mantle, may be related to the existence of small-scale convection cells generated by thickness variations between the continental crust and the very thin crust, process which probably started during and after the rifting phases.

The post-rift continental margin tectonic activity is enhanced by gravity-driven instabilities (Bache et al. 2011). Two episodes of debris flow deposition in the basin are observed on the seismic profiles. One of these episodes, has been dated at 10 Ma, and affected the margin over a large area extending from the major scarps. Another episode is much more recent, and is concentrated on the deeper parts of the margin. These deposits result from the post-rift activity along the margin and is definitely combined with a tectonic reactivation, probably as one of the consequences of the magmatic activity and/or small-scale convection cells. Two distinct deformational regimes, tectonic inheritance, pre-rift rheology (Cloetingh et al. 1995) or exhumation processes of continental mantle at the OCT (Van der Beek et al. 1995; Leroy et al. 2010b) can be responsible for the post-rift activity variations along the margin. While some authors have evoked the impact of climate (Petit et al. 2007), especially regarding the presence of monsoons dynamics since 15 Ma, this cannot be responsible as a unique parameter for the post-rift margin uplift (van der Beek et al. 1995; Bache et al. 2011).

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