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The Red Sea and Gulf of Aden Basins

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Abstract

We here summarize the evolution of the greater Red Sea–Gulf of Aden rift system, which includes the Gulfs of Suez and Aqaba, the Red Sea and Gulf of Aden marine basins and their continental margins, and the Afar region. Plume related basaltic trap volcanism began in Ethiopia, NE Sudan (Derudeb), and SW Yemen at ~31 Ma, followed by rhyolitic volcanism at ~30 Ma. Volcanism thereafter spread northward to Harrats Sirat, Hadan, Ishara-Khirsat, and Ar Rahat in western Saudi Arabia. This early magmatism occurred without significant extension, and continued to ~25 Ma. Much of the Red Sea and Gulf of Aden region was at or near sea level at this time. Starting between ~29.9 and 28.7 Ma, marine syn-tectonic sediments were deposited on continental crust in the central Gulf of Aden. At the same time the Horn of Africa became emergent. By ~27.5–23.8 Ma a small rift basin was forming in the Eritrean Red Sea. At approximately the same time (~25 Ma), extension and rifting commenced within Afar itself. At ~24 Ma, a new phase of volcanism, principally basaltic dikes but also layered gabbro and granophyre bodies, appeared nearly synchronously throughout the entire Red Sea, from Afar and Yemen to northern Egypt. This second phase of magmatism was accompanied in the Red Sea by strong rift-normal extension and deposition of syn-tectonic sediments, mostly of marine and marginal marine affinity. Sedimentary facies were laterally heterogeneous, being comprised of inter-fingering siliciclastics, evaporite, and carbonate. Throughout the Red Sea, the principal phase of rift shoulder uplift and rapid syn-rift subsidence followed shortly thereafter at ~20 Ma. Water depths increased dramatically and sedimentation changed to predominantly Globigerina-rich marl and deepwater limestone.

Within a few million years of its initiation in the mid-Oligocene the Gulf of Aden continental rift linked the Owen fracture zone (oceanic crust) with the Afar plume. The principal driving force for extension was slab-pull beneath the Urumieh-Doktar arc on the north side of the narrowing Neotethys. Drag of Arabia by the northward-moving Indian plate across the partially locked northern Owen fracture zone and the position of the Carlsberg oceanic ridge probably also influenced the geometry of the Aden rift. The trigger for the onset of rifting, though, was the impingement of the Afar plume at ~31 Ma. The Red Sea propagated away from the plume head, perpendicular to the extensional stresses then operating in Arabia, and arrived at the bend in the African-Levant margin, which itself may have been a stress concentration ripe for rifting.

The local geometry of the early Red Sea rift was strongly influenced by pre-existing basement structures, and as a consequence followed a complex path from Afar to Suez. Each segment of the rift was initially an asymmetric half graben, with well-defined accommodation zones between sub-basins. In the Gulf of Aden, the positions of accommodation zones were strongly influenced by older Mesozoic rift basins. Early rift structures can be restored to their original contiguous geometries along both the Red Sea and Gulf of Aden conjugate margins. In both basins, present-day shorelines restore to a separation of 40–60 km along most of their lengths. The initial rift basins were 60–80 km in width.

Oceanic spreading initiated on the Sheba Ridge east of the Alula-Fartaq fracture zone at ~19–18 Ma. After stalling at this fracture zone, the ridge probably propagated west into the central Gulf of Aden by ~16 Ma. This matches the observed termination of syn-tectonic deposition along the onshore Aden margins at approximately the same time.

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At ~14 Ma, a transform boundary cut through Sinai and the Levant continental margin, linking the northern Red Sea with the Bitlis-Zagros convergence zone. This corresponded with collision of Arabia and Eurasia, which resulted in a new plate geometry with different boundary forces. Red Sea extension changed from rift normal (N60°E) to highly oblique and parallel to the Aqaba–Levant transform (N15°E). North of Suez in Egypt the rift system became emergent, perhaps due to minor compression of the Sinai sub-plate, and the marine connection to the Mediterranean Sea became restricted but not terminated. Red Sea sedimentation changed from predominantly open marine to evaporitic, although deep water persisted in many regions. A third phase of magmatism commenced, locally in Ethiopia but predominantly in western Saudi Arabia and extending north to Harrat Ash Shama and Jebel Druse in Jordan, Lebanon, and Syria.

At ~10 Ma, the Sheba Ridge rapidly propagated west over 400 km from the central Gulf of Aden to the Shukra al Sheik discontinuity. Oceanic spreading followed in the south-central Red Sea at ~5 Ma. This corresponded in time to an important unconformity throughout the Red Sea basin and along the margins of the Gulf of Aden, coeval with the Messinian unconformity of the Mediterranean basin. A major phase of pull-apart basin development also occurred along the Aqaba–Levant transform. In the early Pliocene the influx of marine waters through Bab al Mandeb increased and Red Sea sedimentation thereafter returned to predominantly open marine conditions. By ~3–2 Ma, oceanic spreading moved west of the Shukra al Sheik discontinuity, and the entire Gulf of Aden was an oceanic rift.

During the last ~1 My, the southern Red Sea plate boundary linked to the Aden spreading center through the Gulf of Zula, Danakil Depression, and Gulf of Tadjoura. Presently, the Red Sea spreading center appears to be propagating toward the northern Red Sea to link with the Aqaba–Levant transform. Alkali basaltic volcanism continues within the Younger Harrats of western Saudi Arabia and Yemen and offshore southern Red Sea islands. Most of the Arabian plate is now experiencing N–S upper crustal compression, whereas the maximum horizontal stress is oriented E–W in NE Africa. Arabia and Africa, now on separate plates, are therefore completely decoupled in terms of regional, far-field stresses.

Keywords: Red Sea; Gulf of Aden; Afar; Gulf of Suez; Gulf of Aqaba; Continental rifting; Oceanic spreading; Plumes

1. Introduction

Rifting in the Red Sea and Gulf of Aden continues an episodic process that began in the Permian: the peeling away of strips of the Gondwana part of Pangaea along Paleo- and Neotethyan continental margins (Stampfli et al., 2001). It is in reality more complex than this, because collision of Arabia with Eurasia and the emergence of the Afar plume also played important roles in the kinematics and dynamics of the region (Fig. 1). Although the Red Sea and Gulf of Aden share in many respects a common tectonostratigraphic history due to their common role in the separation of Arabia from Africa, the kinematics and dynamics of these rifts were very different. The Gulf of Aden experienced oblique continental rifting, and became an oceanic rift as a result of the propagation of the Sheba Ridge from the Indian Ocean into the African continent. The Red Sea started with rift-normal extension, and switched to oblique-rifting much later. Its oceanic rift developed completely within continental lithosphere, without any connection to the world mid-ocean ridge system. By integrating studies of both rifts, an informative view of the onset of continental break-up can be obtained.

The Gulf of Aden and Red Sea basins tectonically link through Afar (Fig. 1). The geometry of this connection is dramatically illustrated by compilations of seismicity (Fig. 2; e.g., Ambraseys et al., 1994; Hofstetter and Beyth, 2003). The Red Sea rift initially included the present Gulf of Suez and the Bitter Lakes and Nile Delta region on the continental margin of North Africa (Bosworth and McClay, 2001). Mid-way through the Red Sea's evolution, the Gulf of Aqaba–Levant transform boundary was initiated, by-passing the Gulf of Suez. An understanding of the rifting of Arabia from Africa must be based on analysis of the total rift system: the Gulf of Aden, Afar, the Red Sea, and the Gulfs of Suez and Aqaba. We briefly review the present understanding of this greater Red Sea–Gulf of Aden rift system and present a synthesis of its tectonic development over the past 31 My. We use the geologic time scale of Gradstein et al. (2004) for Series/Epoch and Stage boundaries, but revert to Berggren et al. (1995) for the ages of biozones internal to Stages.

2. Afar

Initial continental rifting of the Gulf of Aden and southernmost Red Sea began nearly synchronously with magmatic activity in the region surrounding Afar (evidence is discussed in Section 3). This locality consequently played a central role in the development of "hot spot" models of continental rifting (e.g., Dewey and Bird, 1970; Morgan, 1971; Burke and Dewey, 1973; Falvey, 1974; Burke and Şengör, 1978). There remains considerable disagreement, however, regarding the timing and tectonic significance of uplift of the greater Ethiopian dome, and the control exerted by plume forces on subsequent rifting. It is appropriate therefore to begin discussion of the Red Sea–Gulf of Aden rifts with Afar. We first discuss the Ethiopian Afar Depression and its surrounding plateaus, and then shift to the Arabian sector of Afar which was formerly contiguous with its African counterpart. Further insight into the evolution of Afar and its relationship to the East African rift system can be found in Chorowicz (this issue).

2.1. African margin

Afar has been considered the classic example of a rift–rift–rift triple junction (McKenzie and Morgan, 1969;

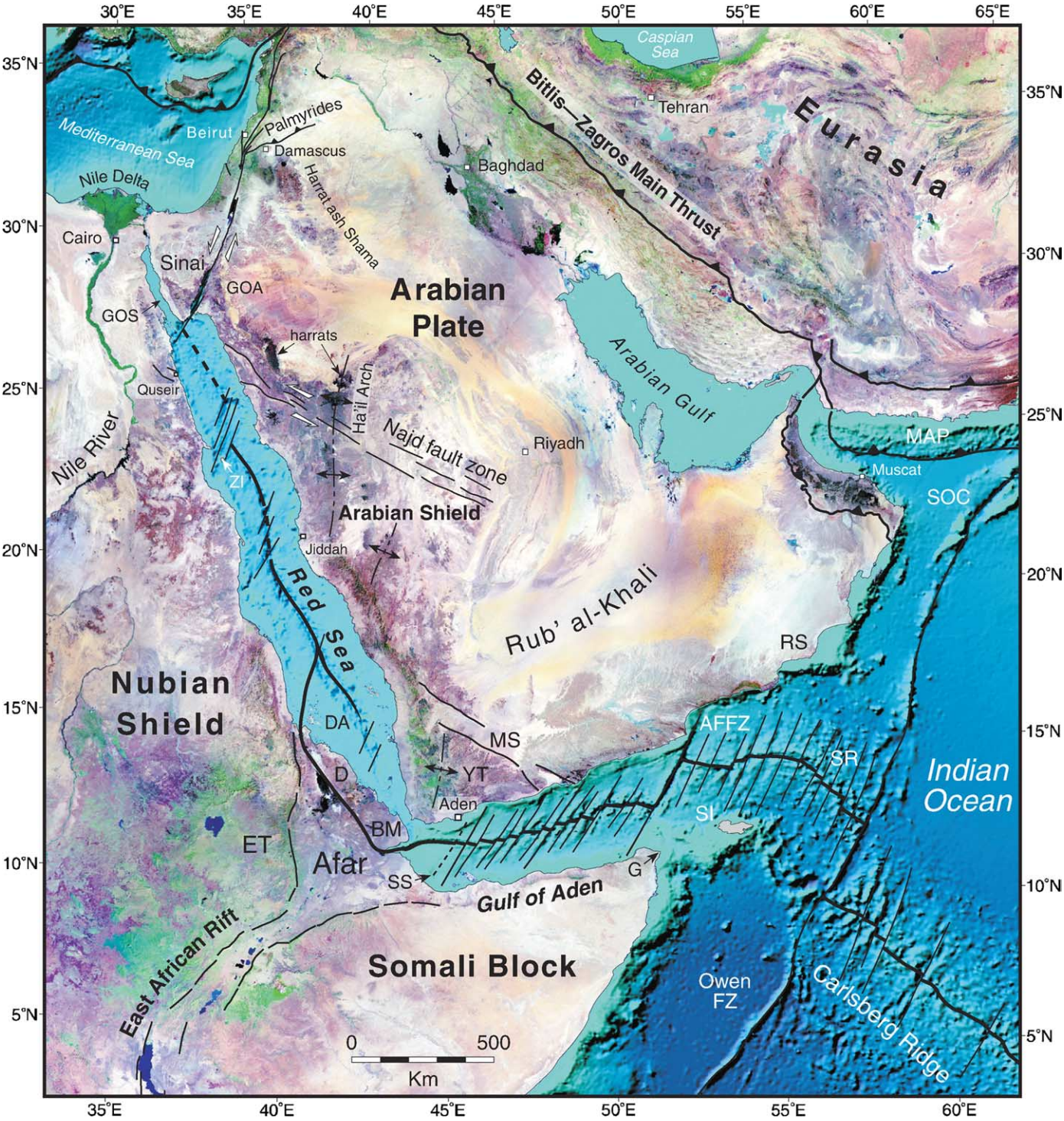


Fig. 1. Landsat imagery (onshore) and Seasat-derived bathymetry (offshore) of the Red Sea–Gulf of Aden rift system and environs. Individual Landsat 5 scenes were color-matched and mosaiced by the Stennis Space Center of NASA. Seasat bathymetric interpretation is from Smith and Sandwell, 1997. Major elements of the Neoproterozoic Pan-African Najd fault system, the Aqaba–Levant intra-continental transform boundary, the Mediterranean–Bitlis–Zagros convergence zone, and East African rift are highlighted. Albers conical equal area projection. AFFZ = Alula-Fartaq fracture zone; BM = Bab al Mandeb; D = Danakil horst; DA = Dahlak archipelago; ET = Ethiopian trap series; G = Cape Gwardafuy; GOA = Gulf of Aqaba; GOS = Gulf of Suez; MAP = Makran accretionary prism; MER = main Ethiopian rift; MS = Mesozoic Marib-Shabwa (Sab'atayn) Basin; SI = Socotra Island; SOC = Semail oceanic crust; SR = Sheba Ridge; SS = Shukra al Sheik discontinuity; RS = Ras Sharbithat; YT = Yemen trap series; ZI = Zabargad Island.

Dewey and Bird, 1970). Afar is formed by the confluence of the Main Ethiopian Rift, western Gulf of Aden, and southern Red Sea (Fig. 1). Most of the region is covered by a thick succession of flood basalts (Fig. 3). The physiographically low-lying part of the Afar triple junction lies in Africa, and covers an area of ~200,000 km^2. This central

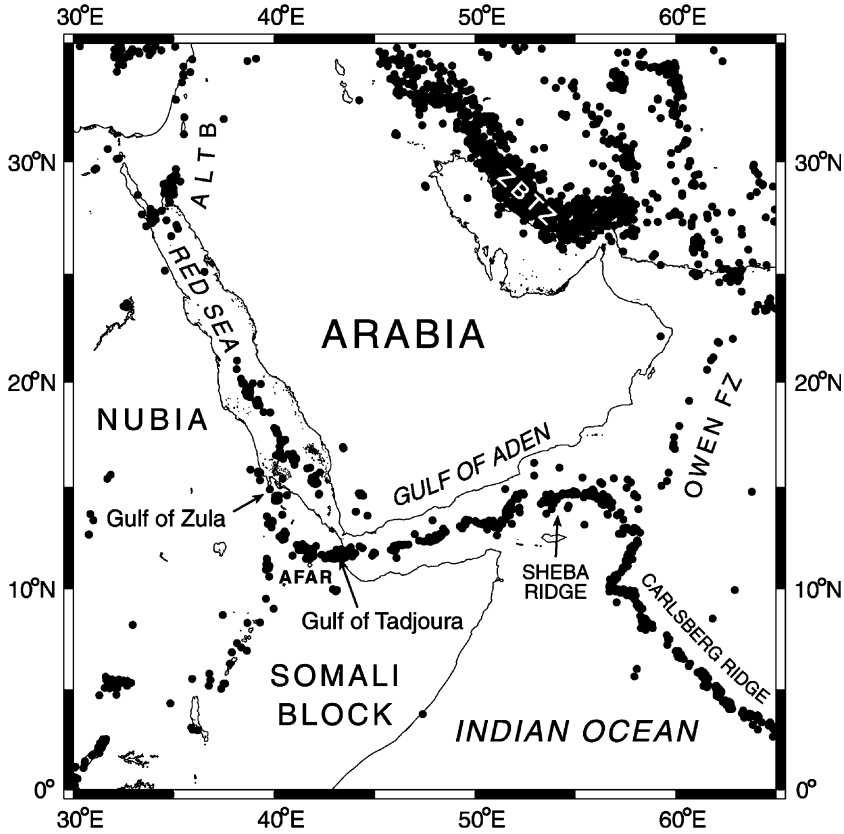


Fig. 2. Seismicity ($M \geqslant 5$) of the Arabian plate and surrounding regions. ZBTZ = Zagros-Bitlis thrust zone; ALTB = Aqaba–Levant transform boundary. After Khanbari, 2000.

area forms the Afar Depression, and is flanked on the west and southeast by the Ethiopian and Somali Plateaus, and on the east by the Danakil and Ali-Sabieh (Aisha) blocks (Fig. 3c). Elevations in the adjacent plateaus reach ~3000 m, and in the Danakil Alps exceed 2100 m. This is in marked contrast to the depression, where elevations range from about +800 m to more than −100 m at Dallol (CNR–CNRS Afar team, 1973; Mohr, 1983a). This low elevation region is interrupted by a number of topographically high shield volcanoes.

The Afar depression is floored principally by Pliocene and younger volcanic rocks, most importantly by the Afar Stratoid series (CNR–CNRS Afar team, 1973; Varet, 1978). Integration of outcrop data from adjacent marginal areas and the Danakil Alps (Gass, 1970; CNR–CNRS Afar team, 1973; Varet, 1978; Zanettin et al., 1978; Zanettin, 1993; Tefera et al., 1996) suggests a six-fold division of the geologic history: (1) Neoproterozoic basement, (2) Pre-rift Mesozoic strata and Early Tertiary volcanic rocks, (3) Oligocene plume volcanism, (4) Syn-rift Miocene volcanic rocks, (5) Syn-drift Pliocene and Pleistocene volcanic rocks, and (6) Quaternary volcanic rocks and lake sediments.

2.1.1. Neoproterozoic basement

Neoproterozoic crystalline basement rocks of the Nubian shield are exposed along the periphery of the Afar Depression, and within the Danakil and Ali-Sabieh blocks (Figs. 1 and 3c; Kazmin, 1971; Vail, 1976, 1985; Kazmin et al., 1978). These rocks were assembled and metamorphosed from 800 to 650 Ma during closure of the Mozambique Ocean, suturing the components of East and West Gondwana along the East African Orogen (Stern, 1994; Kusky et al., 2003). This and related Gondwana tectonic events that lasted until ~450 Ma are commonly referred to as the Pan-African Orogeny, but are in actuality a series of orogenies (reviewed in Kröner, 1993; Stern, 1994). The basement lithologies and their geologic history are similar to the much more extensive exposures north along the margins of the Red Sea (Section 4.1.1).

Basement structures in the Afar region, as in other segments of the rift system (McConnell, 1975), influenced late Tertiary rifting. N–S trending Neoproterozoic shear zones in the northern Danakil region were reactivated (Ghebreab, 1998; Ghebreab et al., 2002), as were a variety of Neoproterozoic structures in western Afar (Collet et al., 2000). An old, regional NNW–SSE trending lineament referred to as the Marda fault (Black et al., 1974; Purcell, 1976) is aligned with some Pleistocene volcanic centers and the main trend of the Red Sea itself. Kazmin and Garland (1973), relying principally on stratigraphic thickness changes, interpreted the Ethiopian escarpment bounding western Afar to have been an active high-angle fault boundary since the Neoproterozoic. However, citing

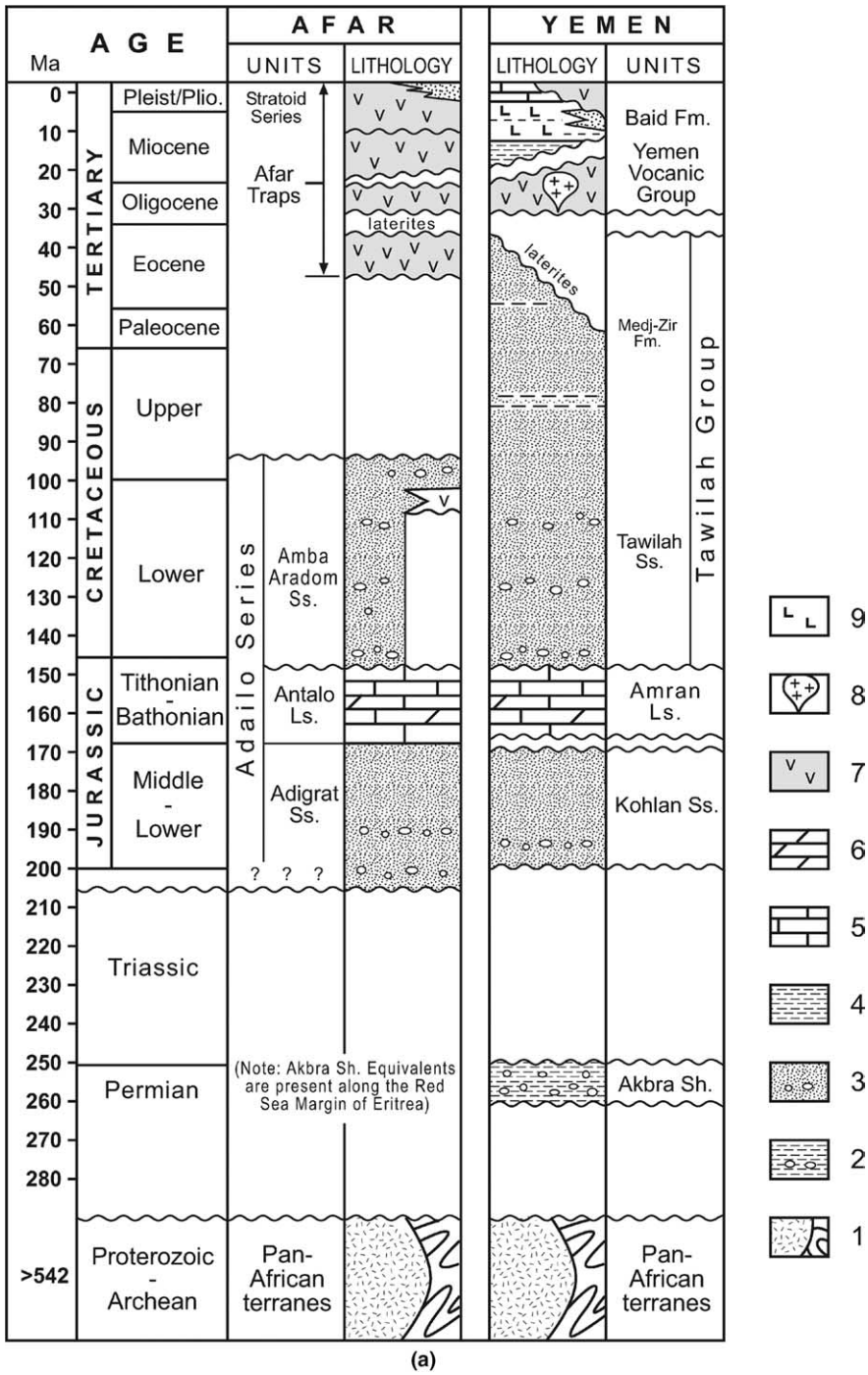
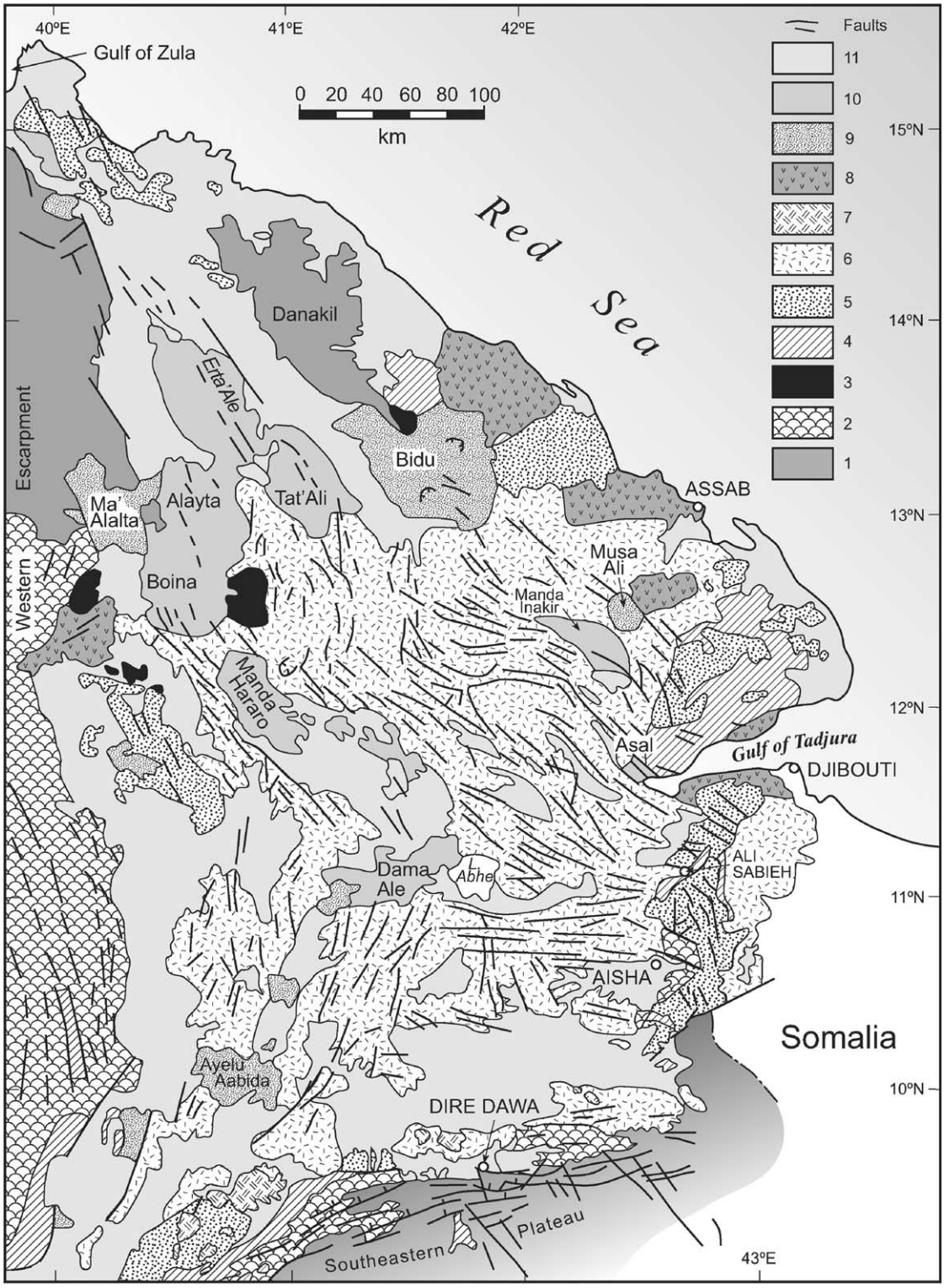


Fig. 3. Geology of the Afar region: (a) generalized stratigraphy of the Afar Depression and correlations with western Yemen (after Canuti et al., 1972; Davison et al., 1994; Sagri et al., 1998)—1, crystalline basement; 2, pebbly mudstone; 3, sandstone and conglomerate; 4, shale; 5, limestone; 6, dolostone; 7, volcanic rocks; 8 granitic intrusives; 9, evaporite; (b) volcanic stratigraphic terminology of the Afar Depression (after Barberi et al., 1975; Zanettin, 1993); (c) tectonic map of Afar (compiled by Berhe, 1986, largely after CNR–CNRS, 1975)—1, metamorphic and Mesozoic sedimentary rocks; 2, western escarpment: Ashangi, Aiba, and Alaje basalts; eastern escarpment: Adolei and Alaje basalts; 3, Miocene granite; 4, Mabla/Arba Guracha rhyolites and Achar/Fursa basalts; 5, Dalha/Lower Afar Stratoid basalt; 6, Afar Stratoid series; 7, rhyolite domes and flows; 8, transverse volcanic structures; 9, marginal volcanic complexes; 10, axial volcanic ranges (Erta'Ale, Tat'Ali, Alayta, Boina, Manda Hararo, Dama Ale, Manda Inakir, Asal); 11, Miocene to Quaternary sedimentary rocks and alluvium.

obliquity between the main Ethiopian and Red Sea structural trends and nearby Precambrian faults and schistosity, Coleman (1974) and Mohr (1975) suggested that basement structures had minimal influence on rifting. With this we disagree, particularly at local scales on the basis of observations from the Red Sea discussed below (Section 4.1.1). Mohr (1975) also emphasized that a proto-Afar rift did not arise until the Mesozoic, and this observation is supported by stratigraphic data from the southern Red Sea and Gulf of Aden.

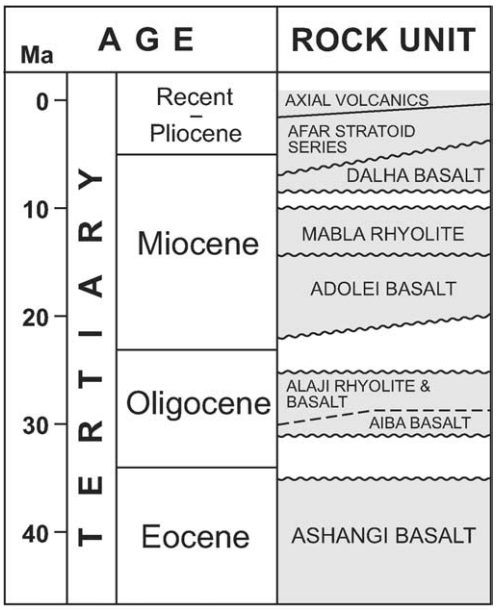


(c)

Fig. 3 (*continued*)

2.1.2. Pre-rift sequence

The pre-rift stratigraphic sequence of the Afar region is best exposed in the Danakil Alps (Fig. 3a and c), where the composite section reaches over 4000 m (Bunter et al., 1998; Sagri et al., 1998). The oldest strata are generally Early Jurassic or possibly Triassic overlying crystalline basement, with higher units extending into the Early Cretaceous. This Mesozoic section represents a major transgressive–regressive cycle, which has been assigned to the Adailo series (Fig. 3a; Bunter et al., 1998). The base of the section is



(b)

Fig. 3 (*continued*)

comprised of fluvial, deltaic, and marginal marine sandstones and conglomerates of the Adigrat Sandstone that attain a thickness of ~1600 m. These siliciclastic rocks are overlain conformably by over 2400 m of the Bathonian to Tithonian Antalo Limestone (see discussion of ages in Beyth, 1972). The carbonates were deposited in a low-angle ramp environment, with several flooding surfaces and numerous cycles of relative sea-level change (Sagri et al., 1998). A return to continental conditions occurred in the latest Jurassic to Early Cretaceous, with deposition of the fluvial Amba Aradom Sandstone (Merla et al., 1979). These upper siliciclastic rocks reach a thickness of about 500 m (Bunter et al., 1998). On the Ethiopian plateau west of Danakil (Mekele area), Antalo Limestone and older strata are preserved in a WNW-trending graben that predates the Amba Aradom Sandstone and is therefore probably part of the Mesozoic rift system that affected the Gulf of Aden margins in Yemen and Somalia (Section 3.1.2; Black et al., 1974; Bosellini et al., 1997). At the southeastern margin of Afar in the Ahmar Mountains 50 km west of Harrar, alkali basalt flows are interbedded between the Antalo Limestone and the Amba Aradom Sandstone (Mohr, 1971; Canuti et al., 1972). Canuti et al. interpreted these flows to be of Aptian to Cenomanian in age. As noted by Mohr (1975), this was the first significant volcanism in the Horn of Africa since the Precambrian.

The pre-rift strata of the Danakil area are cut by numerous extensional faults. Fields of fault blocks are rotated by as much as 60° (vertical bedding is occasionally reported), and fault planes dip moderately to locally horizontally (Morton and Black, 1975). This degree of deformation is comparable to the most highly extended parts of the Gulf of Suez–Red Sea (Bosworth, 1995; Section 4.2.2). Based on geometric considerations, Morton and Black estimated that the crust had been thinned to 40% of its original thickness in the areas of 60° rotation (extended by ~150%), which equates to a stretching factor (β) of 2.5. Cross-sections presented by Morton and Black indicate that all of this extension occurred after eruption of Oligocene basalts.

The thickness of the Adigrat Sandstone and Antalo Limestone progressively increases from west to east, from the Ethiopian/Eritrean Plateau through the Danakil Alps toward the Red Sea. This has been cited as evidence for the existence of a Jurassic-age proto-Red Sea rift basin (Mohr, 1975 and references therein; Bunter et al., 1998), of which the Mekele graben may be a small vestige.

On the Ethiopian Plateau, the top of the Mesozoic pre-rift stratigraphy is marked by a major unconformity, over which were extruded the Ethiopian flood basalts or "trap series." This sequence reaches a preserved thickness of ~2 km and presently covers an area of ~0.6 million km^2 (Mohr, 1983b; Zanettin, 1993). Prior to erosion, the total thickness of the volcanics was ~4 km and they covered a much larger area. The basal units of the traps were considered to be ~60 Ma old (Varet, 1978), but there are no reliable radiometric dates (discussed in the next section) and they are probably younger. Rhyolites, ignimbrites, tuffs, and locally fluvial and lacustrine sediments are interbedded with the oldest basalts. This series of volcanic rocks, sometimes referred to as the Ashangi basalts (Fig. 3a; Zanettin et al., 1978), reportedly continued into the Late Eocene (i.e. pre~34 Ma). If an Eocene age is eventually established for these basal trap units, then they probably reflect the northern effects of an older plume that impinged beneath the southern part of the main Ethiopian rift at ~45–35 Ma (Ebinger et al., 1993).

2.1.3. Oligocene plume volcanism

Plume-related volcanism commenced in the Early Oligocene with extrusion of the Aiba basalt and its lateral equivalents (Fig. 3a; Zanettin et al., 1978). The Aiba series is separated from the pre-plume Ashangi series by a regional peneplain. In northern Eritrea, this peneplain is marked by a laterite horizon that is present at the same stratigraphic position in Yemen (Section 2.2.2). Coleman (1993) cited this as evidence that the region of Afar remained near sea level prior to eruption of the Aiba basalt. This is consistent with the presence of Early to mid-Oligocene marine pre-rift strata identified in offshore wells in the central Gulf of Aden and Eritrean Red Sea (Sections 3.2.1 and 4.2.1; Hughes et al., 1991). These data do not preclude pre-Aiba localized plume uplift in the central Afar region itself, where pre-Aiba laterites are not proven (Burke, 1996), nor do they preclude uplift synchronous with eruption of the Aiba basalt. Şengör (2001) has discussed the difficulties inherent in relying on the presence of laterite to indicate paleo-elevation or surface slope. Given the uncertainties regarding the timing of Afar trap volcanism, doming, and extension, Burke (1996) concluded that all three could have occurred simultaneously.

The age of the post-Ashangi traps was previously considered to span ~32 Ma to 20 or 15 Ma, based on stratigraphic relationships and K–Ar age dates (e.g., Mohr, 1975; Zanettin et al., 1978; Zanettin, 1993). As shown by Féraud et al. (1991) and Al-Kadasi (1995) for magmatism on the Arabian plate, mineral alteration and consequent ^{40}Ar loss can make K–Ar dates unreliable indicators of age of solidification. The lack of precision and accuracy in K–Ar data was similarly recognized for the Ethiopian traps by Hoffmann et al. (1997). Furthermore, long-ranging debates have occurred regarding the appropriate stratigraphic nomenclature of the Afar volcanic rocks (e.g., Mohr, 1973; Barberi et al., 1973), and even the appropriateness of attempting regional correlations of local series (e.g., Barberi et al., 1972a). Recent ^{40}Ar/^{39}Ar determinations have produced a very different view of the timing of volcanism in both NE Africa and Arabia, and have produced a better defined sequence of volcanism in Afar (Fig. 4).

Trachyte near the base of the section at Desi-Bati in the southern Ethiopian escarpment of Afar has been dated at 30.92 ± 0.11 Ma (Ukstins et al., 2002). The oldest rhyolite ignimbrite in the same section is 30.16 ± 0.13 Ma. Basalt and ignimbrite at the top of the escarpment cluster at ~25.0–25.3 Ma (Ukstins et al., 2002). These dates and numerous others (Zumbo et al., 1995; Chernet et al., 1998; George et al., 1998; Rochette et al., 1997; Coulié et al., 2003) demonstrate that the Afar plume traps were erupted over a total period of ~6 My. The bulk of the basalt traps were erupted over a very brief period, from ~30 to ~29 Ma (Hoffmann et al., 1997; Coulié et al., 2003).

The eruption at Afar of the Oligocene plume-related flood basalts occurred without significant identifiable extension (Barberi et al., 1972b, 1975), although localized faulting may have occurred.

Simultaneous with plume magmatism in Ethiopia (and adjacent Yemen, discussed below, Section 2.2.3) localized eruptions occurred further north in NE Sudan and western Saudi Arabia. At Derudeb in the southern Sudanese Red Sea Hills (Fig. 4), a felsic tuff and a rhyolite flow have been dated 29.9 ± 0.3 and 29.6 ± 0.3, respectively (^{40}Ar/^{39}Ar; Kenea et al., 2001). This corresponds very closely with the onset of silicic magmatism at Desi-Bati. Basaltic eruptions at Derudeb began somewhat before this (circa 31 Ma), also analogous to Afar.

In Saudi Arabia, the volcanic centers that were coeval with eruption of the Afar plume are referred to as the "Older Harrats" (e.g., Coleman et al., 1983; Brown et al.,

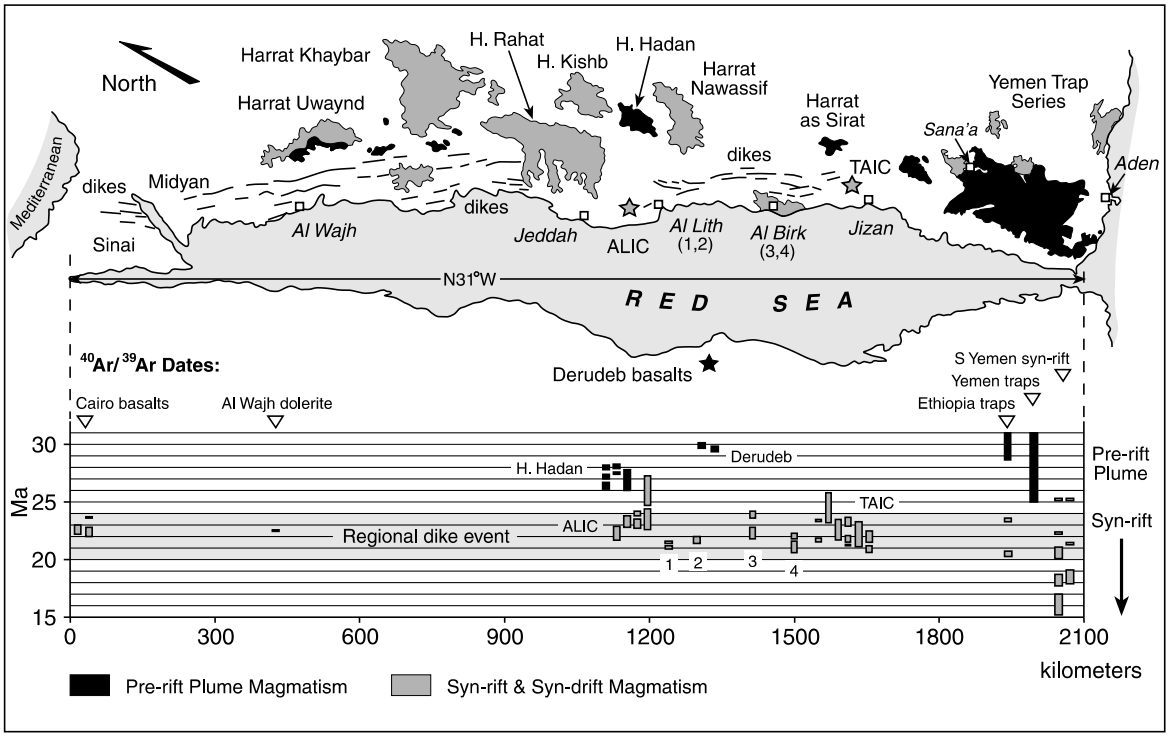


Fig. 4. Compilation of published ^{40}Ar/^{39}Ar dates for the Red Sea basin, and location of major Tertiary magmatism in Arabia and Sinai. Relative positions of Cairo (Kappelman et al., 1992; Lotfy et al., 1995) and Derudeb basalts (Kenea et al., 2001) and Ethiopia traps are also indicated, but minor dikes along the African margin are not. K–Ar analyses are not included, as in many areas they are shown to be unreliable. Two sequences of Oligocene–Early Miocene magmatism are distinguishable: (1) that associated with the impingement of the Afar plume at ~31 Ma, which is not associated with significant syntectonic sedimentation or extension, and (2) that associated with the onset of extension along essentially the entire rift at ~24–23 Ma. Heights of boxes indicate published error bars for individual samples except for the Ethiopia and Yemen traps, which are composites of numerous samples. Numbers 1–4 refer to gabbroic dike localities at Al Lith and Al Birk; ALIC = Al Lith igneous complex; TAIC = Tihama Asir igneous complex (Sebai et al., 1991). S Yemen syn-rift = ages of dikes and Dhala pluton reported by Zumbo et al. (1995) for southernmost part of the Yemen trap province. Other sources of data are listed in the text.

1989; Coleman, 1993). The Older Harrats were typically erupted from central vents and included olivine transitional basalts to alkali olivine basalts. The largest preserved eruptions occurred at Harrats Uwaynd, Hadan, and as Sirat (Fig. 4). $^{40}Ar/^{39}Ar$ dates are only available for Harrat Hadan, with six samples yielding ages from ~28 to ~26 Ma (Sebai et al., 1991; Féraud et al., 1991). Eruption of the Older Harrats was strongly influenced by N–S trending arches (Ha'il, Jizan; Fig. 1) that were inherited from Early Paleozoic and Late Cretaceous deformations (Greenwood, 1972; Coleman, 1974).

2.1.4. Syn-rift Miocene volcanism

A dramatic decrease in volcanism occurred in the Ethiopian Plateau from ~25 to ~20 Ma (Ukstins et al., 2002), although sporadic activity occurred as is evidenced by volcaniclastic rocks from Alem Ketema on the plateau north of Addis Ababa dated at ~23.4 to ~20.5 Ma ($^{40}Ar/^{39}Ar$; Coulié et al., 2003). Further east at Robit along the western escarpment of Afar, renewal of volcanism was marked by ignimbrite that is dated at 19.76 ± 0.06 Ma ($^{40}Ar/^{39}Ar$; Ukstins et al., 2002). Alkaline to peralkaline granites were also intruded into Precambrian basement, Jurassic limestone and the pre-plume traps along the margins of Afar (Varet, 1978).

Within the Afar Depression, Miocene volcanic rocks are assigned to the Adolei, Mabla, and Dalha series (Fig. 3a and b; Barberi et al., 1975; Varet, 1978; Vellutini, 1990). The Adolei basalts are found in a strongly tectonized area north of the Gulf of Tadjoura. The flows are highly altered and therefore difficult to date, but a few ages have been determined that extend from ~22 to ~14.6 Ma (K–Ar; Chessex et al., 1975; Barberi et al., 1975; Black et al., 1975). The overlying Mabla series was extruded from N–S trending fissures and aligned vents, and consists of rhyolite, ignimbrite and lesser basalt and pumice. Its age ranges from ~14 to ~10 Ma (K–Ar) according to Barberi et al. (1975), but Audin et al. (2004) reported significantly older $^{40}Ar/^{39}Ar$ ages (17 and 20 Ma) for two rhyolite samples from the Ali-Sabieh area (Fig. 3c). The top of the rhyolite sequence was significantly eroded prior to extrusion of the Dalha series, which also marked a return to basaltic volcanism. The Dalha series ages are from ~8 to ~3.5 Ma (K–Ar; Barberi et al., 1975), confirmed by $^{40}Ar/^{39}Ar$ ages of 7.7–4.8 Ma for four basalt samples (Audin et al., 2004). A more complete regional understanding of how Afar Depression volcanism correlates with the adjacent plateau volcanism will require additional $^{40}Ar/^{39}Ar$ age dating.

The Adolei, Mabla, and Dalha series and correlative Miocene units were strongly affected by extensional faulting during their extrusion (Barberi et al., 1972b, 1975; Zanettin et al., 1978). These authors dated the onset of major extension and rift formation to be ~25 Ma, which coincides roughly with the major dike event of the Arabian Red Sea margin and the beginning of syn-tectonic sedimentation throughout the Red Sea rift basin (Section 4.2.1). The Early Miocene Adolei basalts and their silicic time equivalents represent a major break in the style of magmatism in Ethiopia: plume-related undersaturated alkali basalts were replaced by syn-rift transitional basalts and associated peralkaline rhyolites and granites (Barberi et al., 1975). The change from Adolei basaltic to Mabla rhyolitic volcanism at ~14 Ma corresponds to initial collision of Arabia with Eurasia (Şengör and Yilmaz, 1981; Hempton, 1987; Woodruff and Savin, 1989; Burke, 1996), a major change in regional plate kinematics (Le Pichon and Gaulier, 1988), the beginning of rifting in the western branch of the East African Rift (reviewed in Burke, 1996; Chorowicz, this issue), and our favored time of onset of the Gulf of Aqaba transform boundary (Section 4.2.3). The end of the Mabla series at ~10 Ma is coeval with the rapid propagation of oceanic spreading throughout the west-central Gulf of Aden (Section 3.3).

A 70 km wide belt of Dalha series outcrops (age <8 Ma) along the Red Sea coast is essentially undeformed (Barberi et al., 1975). On the Yemeni side of Bab al Mandeb, near the town of Al Mokha, Huchon et al. (1991) described horizontal, ~10 Ma old basaltic flows resting unconformably over tilted, ~18 Ma old rhyolites equivalent to the Mabla series described above. These relationships suggest that the southernmost Red Sea had stabilized by the Late Miocene (by ~10 Ma), and that all significant extension had transferred west to the Danakil block and within the Danakil Depression.

2.1.5. Pliocene–Pleistocene volcanism

Pliocene–Pleistocene volcanic rocks cover most of the Afar Depression, and these units are mostly assigned to the Afar Stratoid series (Fig. 3a and b; Barberi et al., 1974, 1975; Varet, 1978; Berhe, 1986; Tefera et al., 1996). These rocks include basalts, comendites, and hawaiites, with occasional silicic lavas erupted from discrete centers in the upper part of the section. The Afar Stratoid series reaches a total thickness of ~1500 m (Varet, 1978). The oldest exposed flows of the Afar Stratoid series are ~4.4 Ma (K–Ar; Barberi et al., 1975) or ~7.4 Ma (K–Ar; Berhe, 1986) and may therefore extend down into the Late Miocene. They unconformably overlie the Late Miocene–Early Pliocene(?) Dalha series, and the overlap in ages of the two series may reflect both problems in the K–Ar dating technique and differing correlation schemes. More recent $^{40}Ar/^{39}Ar$ dates of the oldest Stratoid flows ("Somali basalts") are 3–2.7 Ma, with younger parts of the series up to 1.8 Ma (Audin et al., 2004). The youngest K–Ar dates from the Stratoid series are Late Pleistocene (~0.4 Ma; Barberi et al., 1972b, 1975; Civetta et al., 1974, 1975).

The Dalha-Afar Stratoid break in volcanism at ≈5 Ma, if it really exists, corresponds regionally with the onset of Red Sea oceanic spreading and the jump westward from 45°E of the spreading center in the Gulf of Aden (Sections 3.3 and 4.3). Barberi et al. (1975) considered the eruptive style and structuring of the Afar Stratoid series to represent the transition from continental rifting to oceanic spreading.

2.1.6. Quaternary geology and neotectonics

Basaltic flows, scoria cones, and alkaline to peralkaline silicic rocks were erupted locally in the Afar Depression over the past 1 My (Varet, 1978; Tefera et al., 1996). The style of eruption included both fissures and shield volcanoes. At ~200 ka, a marine gulf extended about 250 km SE of the present-day Gulf of Zula, covering the area between the Danakil Alps and the Ethiopian Plateau (Fig. 3c; CNR–CNRS Afar team, 1973). This resulted in the formation of several coral terraces and later the deposition of gypsum and halite. The highest Quaternary marine deposits are at ~90 m elevation.

Active rifting continued through the Quaternary west of the Danakil horst at Erta'Ale, Tat Ali, and Alayta and within central Afar at Manda Hararo, Manda Inakir and Dama Ale (Fig. 3c; Tazieff et al., 1971; Barberi and Varet, 1977; Varet, 1978). Rifting at Asal (Delibrias et al., 1975) connected these structures to the Gulf of Tadjoura, separating the Danakil block from the rest of Afar (Figs. 1, 3c; Manighetti et al., 1998, 2001). Basalts extruded along the axes of these rifts display symmetric magnetic anomalies, out to possibly anomaly 2A (~4 Ma) (Hall, 1970, cited in Barberi and Varet, 1977), and the ages of the basalts progressively increase away from the rift axes. Barberi and Varet (1977) considered these axial basalt ranges to be sub-aerial equivalents of oceanic spreading centers, linked by transform faults. Similarly, Tapponnier and Varet (1974) compared the Mak'arrasou region, between the Asal and Manda-Inakir rifts, to an emergent oceanic transform zone.

Fumarolic and eruptive activity continues in the present-day along the axial ranges, with historic eruptions recorded for Alayta, Dubbi, and Erta'Ale (Fig. 3b; Barberi et al., 1972a). A nearly permanent lava lake exists within the pit crater of Erta'Ale (Varet, 1971).

Formation of the Afar Depression was accompanied by a 13° anticlockwise rotation of the Danakil block since ~6–5 Ma (Tarling, 1970). By extrapolating to the complete closure, Sichler (1980) proposed a total 23° rotation since the inception of the separation between the Ethiopian Plateau and Danakil (Fig. 1). His "crank-arm model" (Souriot and Brun, 1992) explains the triangular shape of the depression, the Danakil block being pinned to Africa near the Gulf of Zula and to Arabia in the Bab al Mandeb region.

Faulting within the young axial rifts of Afar, as with Afar in general, displays predominantly dip-slip movement. Some field relationships and many seismic focal plane analyses, though, indicate a component of active rift-parallel strike-slip movement (e.g., Mohr, 1967; McKenzie et al., 1970; Gouin, 1979; Tapponnier et al., 1990; Lépine and Hirn, 1992; De Chabalier and Avouac, 1994; Hofstetter and Beyth, 2003). Outcrop fault kinematic studies confirm the existence of strike-slip faults striking at low-angles to the general trends of the Tendaho, Dobi-Hanle, Asal, and other active rifts in the area west of the Gulf of Tadjoura (Fig. 3b; Zan et al., 1990; Passerini et al., 1991; Abbate et al., 1995; Chorowicz et al., 1999; Audin et al., 2001). Dextral faults are present, but sinistral movement is predominant. Slickensides and slickenlines are best developed within the Dalha series basalts and Mabla rhyolites and more rarely within the Afar Stratoid series. Open en-echelon tension gashes affecting the $\leqslant$ 0.7 Ma Asal series of the Asal rift confirm at least some strike-slip movement in the Quaternary (Passerini et al., 1988). Transverse strike-slip movement, that would be analogous to a discrete, oceanic transform fault, is generally not observed (Abbate et al., 1995). Exceptions are recorded near Arta (Arthaud et al., 1980), in the onshore continuation of the offshore Maskali transform fault (Choukroune et al., 1988), along the western border of the Ali Sabieh block (Gaulier and Huchon, 1991), and along the Holol and Bia-Anot faults (Audin et al., 2004) that cross-cut the Ali-Sabieh block with a NE–SW "transform fault" trend.

These strike-slip components of motion have been diversely interpreted. Acton et al. (1991) proposed a microplate model while Tapponnier et al. (1990) emphasized the role of the overlap between the Manda Hararo-Goba'ad and the Asal-Manda Inakir rifts. Passerini et al. (1991) interpreted the local rift-parallel movement to be a response to complex interactions between the different rifts of the Afar triple junction. Abbate et al. (1995) concluded that rift-parallel compression and/or lateral plate boundary shear could contribute to produce the strike-slip faulting, and suggested that the Afar region is experiencing a small component of anti-clockwise rotation. All palinspastic restorations of the southern Red Sea–Gulf of Aden (see Section 5) indicate overlap in the Afar region and suggest, as noted above for Danakil, complex block movements must have occurred in this area (e.g., McKenzie et al., 1970; Chase, 1978; Le Pichon and Francheteau, 1978; Joffe and Garfunkel, 1987; Acton et al., 1991; Jestin et al., 1994; Manighetti et al., 1998; Collet et al., 2000; Eagles et al., 2002; Redfield et al., 2003).

Mesozoic pre-rift strata similar to those exposed in the Danakil Alps and Neoproterozoic basement are probably present beneath some areas floored by young volcanic rocks in the Danakil Depression (Morton and Black, 1975). This continental crust would be highly attenuated and is difficult to distinguish definitively from oceanic or transitional crust. A common view is that at least the narrow axial volcanic ranges are oceanic, and perhaps much broader regions as well (Barberi and Varet, 1975). Gravity and seismic interpretations indicate that the crust beneath the Ethiopian and Somalian plateaus is ~35–40 km thick, thins to <20 km in most of Afar, and is ~6–10 km along the axial volcanic zone west of the Gulf of Tadjoura (Searle and Gouin, 1971; Makris et al., 1972, 1975, 1991; Berckhemer et al., 1975; Ruegg, 1975). Mohr (1989) argued for most of the Danakil depression being floored by pseudo-oceanic crust but Makris and Ginzburg (1987) maintained that all of the anomalously thin areas are stretched continental crust with the exception of the Gulf of Tadjoura, where a wedge of oceanic crust is propagating westward

(e.g., Courtillot et al., 1980; Manighetti et al., 1997; Dauteuil et al., 2001).

2.2. Arabian margin

The geologic signature of the Afar plume extends into the Arabian plate in the area of what is now SW Yemen (Figs. 1, 5a, and 6b). The mid-Oligocene history of basaltic trap volcanism in Yemen is now known to have many close similarities to that in Ethiopia. Following the initiation of extension offshore Eritrea in the Late Oligocene (discussed in Section 4.2.1), Red Sea rift tectonics were superimposed on the flood basalt terrane, linking with the young Gulf of Aden rift and splitting Afar in two.

The Arabian margin in Yemen is characterized by the presence of the extensive Tihama coastal plain (Fig. 5a and 6b) bounded on the east by the "Great Escarpment" at the edge of the Yemeni highlands. The Tihama plain is up to 40 km wide, trends N–S, and rises gently from the Red Sea to around 200 m in elevation at the foot of the Great Escarpment (Huchon et al., 1991; Davison et al., 1994). The Great Escarpment extends for over 1000 km from Yemen into southern Saudi Arabia. It rises abruptly from ~200 m to elevations in excess of 1000 m. Inland of this dramatic escarpment the Yemeni highlands rise to greater than 2 km above sea level forming a broad plateau along the eastern margin of the Red Sea rift system. Maximum elevation of the plateau is 3660 m at Jebel Nabi Shuyab, west of Sana'a (Fig. 5a). The Tihama Plain is covered principally by Recent alluvial and fluvial deposits with localized outcrops of tilted fault blocks of the Red Sea rift (Davison et al., 1994) and active emergent diapirs of Miocene salt (Davison et al., 1996; Bosence et al., 1998). The Yemeni highlands along the margin of the Red Sea rift are characterized by outcrops of Neoproterozoic basement (in part reworked from Archaean basement), Mesozoic sedimentary strata, Oligocene volcanic rocks, Tertiary pre-rift sedimentary rocks, Oligo-Miocene syn-rift sedimentary rocks, dikes and granites, and Pliocene to Recent sediments (Figs. 3a and 5a; Davison et al., 1994). Six tectonostratigraphic intervals will be discussed for SW Yemen: (1) Precambrian basement; (2) Pre-rift Mesozoic sequences; (3) Pre-rift Oligocene plume volcanism; (4) End Oligocene to Miocene syn-rift units; (5) Syn-drift Pliocene–Pleistocene sediments; and (6) Quaternary geology and neotectonics.

2.2.1. Precambrian basement

The Precambrian crystalline basement of the Yemeni margin of the Red Sea is similar in origin to that of the Neoproterozoic Nubian Shield of the Ethiopian Plateau and Danakil Alps (Section 2.1.1.). Geochronological studies (Windley et al., 1996; Whitehouse et al., 1998; Whitehouse et al., 2001) have permitted the Yemeni Precambrian basement to be subdivided into six terranes that alternate between early Precambrian gneiss and Neoproterozoic island arcs accreted together during the Pan-African orogenies. Neodymium isotopic data suggest that a fundamental Neoproterozoic lithospheric boundary runs through Yemen, separating older crust to the east (the Afif terrane) with average model ages of 2.1 Ga, from Neoproterozoic ages to the west (Stern, 2002). This lithospheric boundary was interpreted to separate East and West Gondwana.

Basement rocks exposed in the footwall of the Great Escarpment in the northern part of the Yemeni margin are interpreted to be a continuation of the Asir terrane of Saudi Arabia and consist of folded gneiss intruded by late to post-tectonic plutons and post-tectonic intermediate to basic dikes. The Asir terrane was probably assembled during an early Pan-African orogeny (900–700 Ma), was subsequently deformed by the "Nabitah" orogeny (690–590 Ma), and then extended by crustal thinning and dike intrusion at 600–550 Ma (Genna et al., 2002).

2.2.2. Pre-rift sequence

The exposed pre-rift stratigraphic sequence of the Yemeni Red Sea margin (Fig. 3a) is similar to that of Eritrea and Ethiopia but much thinner, reaching a total of ~1200 m (Beydoun, 1964; Greenwood and Bleackley, 1967). At the base of the section the Permian Akbra Shale unconformably overlies the basement and may be up to 130 m thick. Lithologies are principally pebbly gray shale and siltstone interpreted to be fluvio-glacial in origin (Kruck and Thiele, 1983), similar to the Enticho and Edaga-Arbi sequences of the southern western Red Sea margin (Section 4.1.3). The overlying Mesozoic section comprises a transgressive–regressive cycle, as at Danakil (Section 2.1.2). The base of the cycle is represented by the Lower Jurassic Kholan Sandstone, correlative to the Adigrat Sandstone (Fig. 3a). The Kholan reaches ~150 m thick and changes from continental character at its base to marine near the top. The sandstone is overlain by the Middle to Upper Jurassic Amran Group, correlative to the Antalo Limestone. In western Yemen the Amran appears to span the Callovian to the earliest Tithonian (Al-Thour, 1997; Beydoun, 1997) and consists of 410–520 m of massive platformal limestone with interbedded shale (Davison et al., 1994). The top of the Amran Group shows evidence of emergence and is unconformably overlain by 150–400 m of medium to coarse-grained, cross-bedded sandstone of the Tawilah Group (Fig. 3a; El-Nakhal, 1988). The age of the Tawilah Group is reported to be Late Cretaceous to Eocene (Al-Subbary et al., 1994) or latest Jurassic to Eocene (Davison et al., 1994). The bulk of the Tawilah Group was deposited in continental fluvial environments, and is probably correlative with the latest Jurassic to Early Cretaceous Amba Aradam Formation of Ethiopia and Eritrea (Sections 2.1.2 and 4.1.3). In the upper part of the Tawilah Group, the Medj-Zir Formation includes shallow marine sandstone with agglutinated marine foraminifera of Paleocene age (Al-Subbary et al., 1998). These beds are correlative with the middle Usfan Formation of the Jeddah region (Section 4.1.3). The top

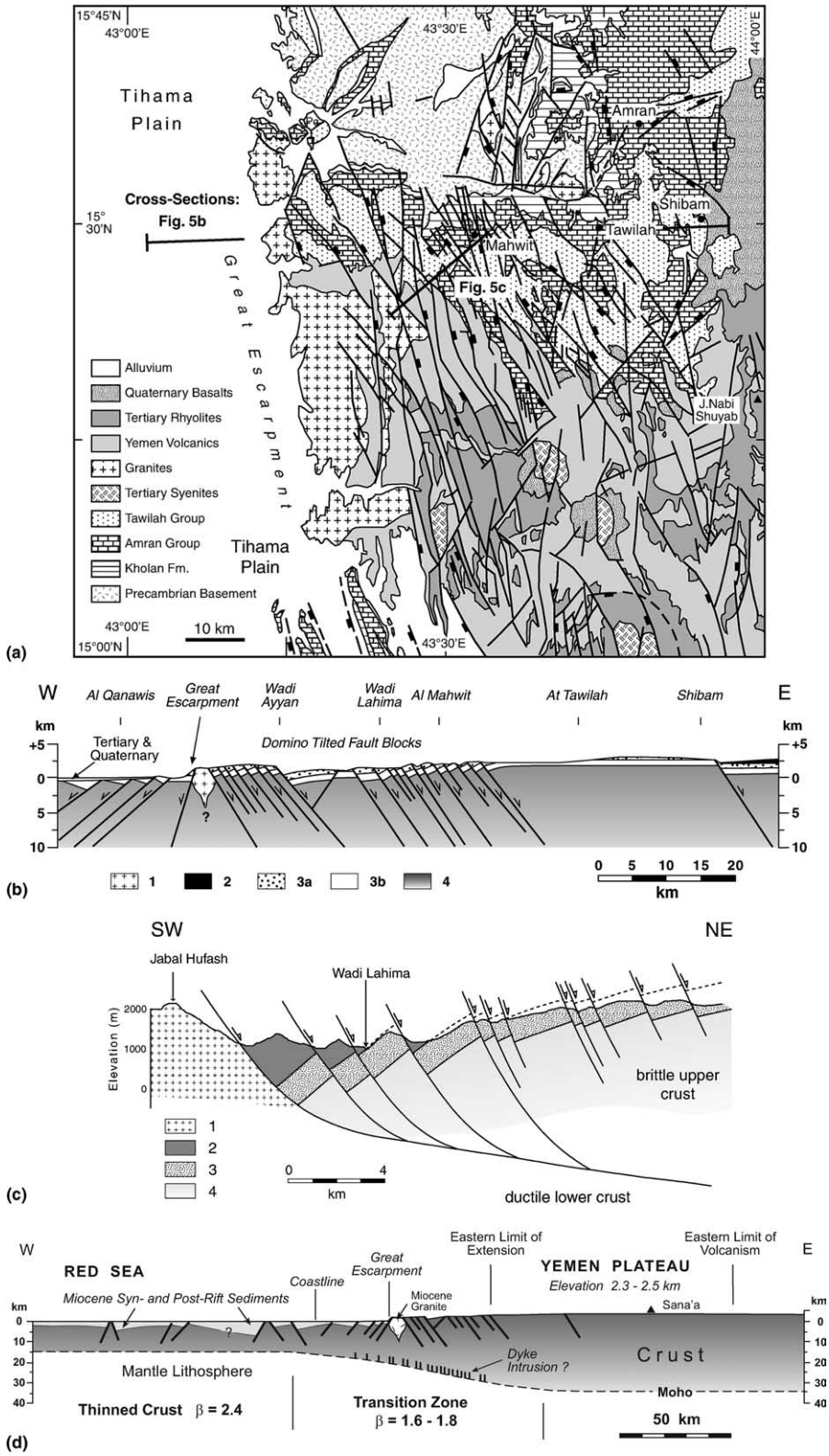
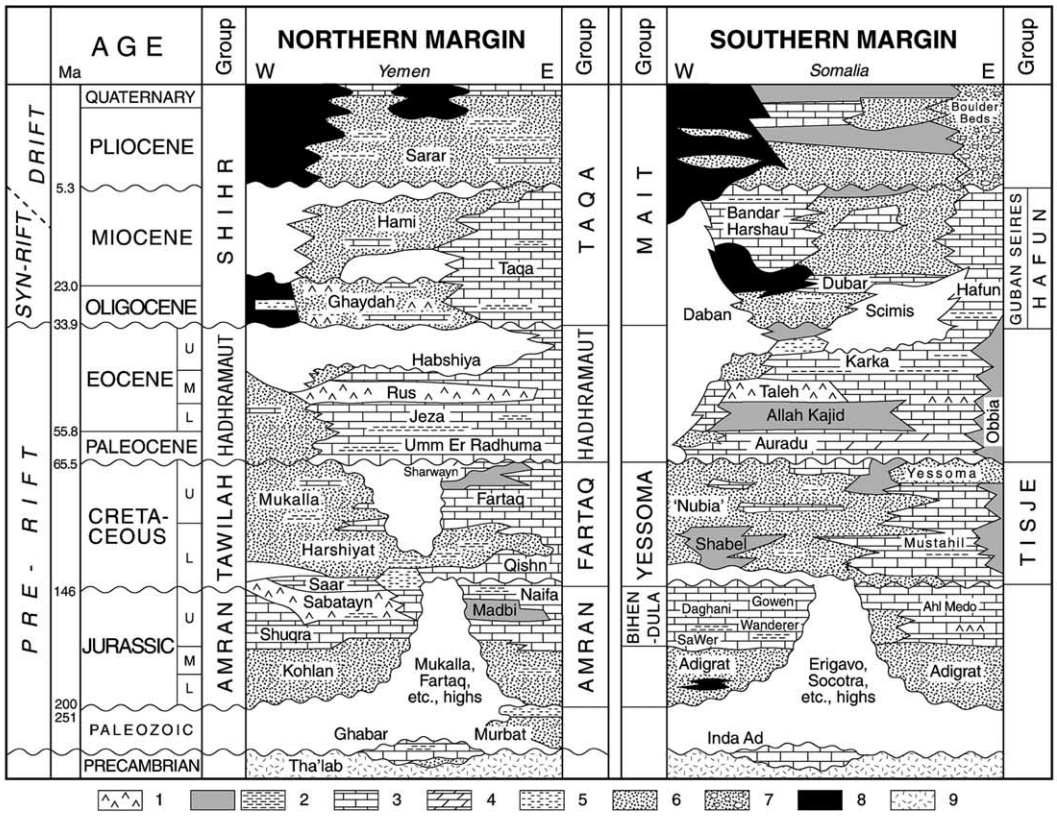


Fig. 5. Geology of the Yemeni Red Sea margin: (a) Tectonic map of SW Yemen. The elevation of Jabal Nabi Shuyab is 3660 m (after Davison et al., 1998). (b) Structural cross-section of the Tihama Plain, Yemeni Red Sea margin, with inferred planar, domino-style faulting. Location shown in a. 1, Miocene granite; 2, Oligocene volcanic rocks; 3a, Cretaceous Tawilah Sandstone; 3b, Jurassic Kholan Sandstone and Amran Limestone; 4, crystalline basement (modified from Davison et al., 1994). (c) Alternative interpretation for the relationship between the 21 Ma granitic intrusion of Jabal Hufash and the tilted fault blocks, emphasizing the gradient of extension toward the granite suggesting that the latter was emplaced passively in the necked area of stretched continental crust (after Geoffroy et al., 1998). Faults are interpreted to be listric and detached at the brittle–ductile transition. Legend as in b. Location shown in a. (d) Crustal-scale cross-section of the Yemeni Red Sea margin, from east of Sana'a to the center of the Red Sea, with inferred stretching factors (after Davison et al., 1994).


AGE
Ma
Group
NORTHERN MARGIN
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Yemen
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Group
SOUTHERN MARGIN
W
Somalia
E
Group
SYN-RIFT
DRIFT
PRE - RIFT
QUATERNARY
PLIOCENE
5.3
MIOCENE
23.0
OLIGOCENE
33.9
EOCENE
U
M
L
55.8
PALEOCENE
65.5
CRETA-CEOUS
U
L
146
JURASSIC
U
M
L
200
251
PALEOZOIC
PRECAMBRIAN
SHIHR
HADHRAMAUT
TAWILAH
AMRAN
Sarar
Hami
Taqa
Ghaydah
Habshiya
Rus
Jeza
Umm Er Radhuma
Sharwayn
Mukalla
Fartaq
Harshiyat
Qishn
Saar
Sabatayn
Naifa
Madbi
Shuqra
Kohlan
Mukalla, Fartaq, etc., highs
Ghabar
Murbat
Tha'lab
TAQA
HADHRAMAUT
FARTAQ
AMRAN
MAIT
YESSOMA
BIHEN -DULA
Boulder Beds
Bandar Harshau
Dubar
Hafun
Daban
Scimis
GUBAN SEIRES
HAFUN
Karka
Taleh
Allah Kajid
Auradu
Obbia
Yessoma
'Nubia'
Shabel
Mustahil
TISJE
Daghani
Gowen
Wanderer
SaWer
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Adigrat
Erigavo, Socotra, etc., highs
Adigrat
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1
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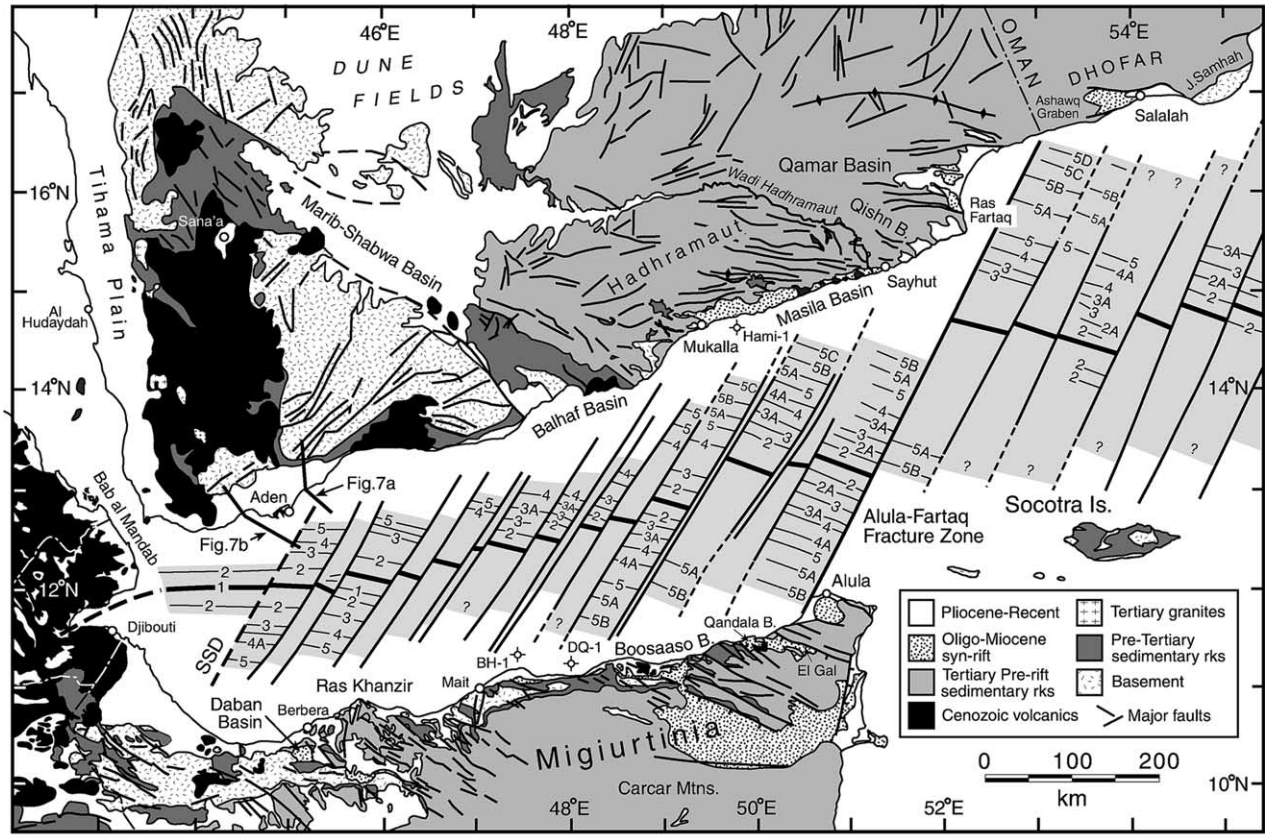
(a)


46°E
48°E
54°E
DUNE FIELDS
OMAN
DHOFAR
J.Samhah
Ashawq Graben
Salalah
16°N
Tihama Plain
Sana'a
Marib-Shabwa Basin
Qamar Basin
Wadi Hadhramaut
Qishn B.
Ras Fartaq
Hadhramaut
Sayhut
Masila Basin
Al Hudaydah
Mukalla
Hami-1
14°N
Balhaf Basin
Bab al Mandab
Aden
Fig.7a
Fig.7b
Socotra Is.
Alula-Fartaq Fracture Zone
12°N
Alula
Djibouti
Qandala B.
SSD
BH-1
DQ-1
Boosaaso B.
El Gal
Ras Khanzir
Mait
Daban Basin
Berbera
Migiurtinia
Carcar Mtns.
48°E
50°E
52°E
10°N
Pliocene-Recent
Oligo-Miocene syn-rift
Tertiary Pre-rift sedimentary rks
Cenozoic volcanics
Tertiary granites
Pre-Tertiary sedimentary rks
Basement
Major faults
0
100
200
km

(b)

of the Tawilah section is characterized by numerous paleosols and ferricrete horizons interpreted to indicate subaerial exposure and lateritic weathering prior to the deposition of the overlying Yemen Oligocene volcanic series. These laterites correlate to the pre-Aiba basalt laterites of the Ethiopian plateau (Fig. 3a; Section 2.1.3). Al-Subbary et al. (1998) and Davison et al. (1994) state that there is no evidence of widespread large-scale doming of the Yemeni margin prior to or during eruption of the Yemeni volcanics (but see the caveats discussed in Section 2.1.3).

2.2.3. Pre-rift Oligocene plume volcanism

Pre-rift volcanism on the Yemeni margin, the Yemeni Volcanic Group, produced >3 km section of basaltic to bimodal volcanism (Moseley, 1969) in a short interval between ~31 Ma and ~29 Ma (Fig. 4; Baker et al., 1994, 1996). The onset of basaltic volcanism in Yemen at around 30.9–30.5 Ma is coeval with that in Ethiopia (Ukstins et al., 2002; Coulié et al., 2003). Silicic volcanism characterized by massive ignimbrites subsequently initiated in Yemen ~29.9 Ma and continued until ~26 Ma (Fig. 4; Baker et al., 1996; Ukstins et al., 2002; Coulié et al., 2003).

The base of the Yemen Volcanic Group is referred to the Jihama Member and consists of volcanic ashes, lithic tuffs and lithi-crystal tuffs (Al-Subbary et al., 1998). The contact with the underlying Medj-Zir Formation of the uppermost Tawilah Group is diachronous but conformable, suggesting that no significant deformation or uplift was occurring at that time (Menzies et al., 1990).

East of the Great Escarpment the Yemeni Volcanic Group is cut and rotated by numerous normal faults (Fig. 5a). This section is then capped by a marked unconformity that extends from ~26 to ~19 Ma (Ukstins et al., 2002). Rifting therefore initiated in the latest Oligocene or Early Miocene. With the exception of dikes discussed below and minor plutonic bodies (e.g., the gabbro-syenite pluton near Dhala, southern Yemen, age ~22–21 Ma; Zumbo et al., 1995), there is little evidence of early Miocene syn-rift basaltic volcanism along the Yemen Red Sea margin (Menzies et al., 1997), reflecting the movement of Yemen away from the focus of the Afar plume in Ethiopia. However, massive alkali granitic bodies, dated at ~21 Ma (Huchon et al., 1991), outcrop both along the Great Escarpment and inland (Jebel Saber, near Taiz). Their chemistry suggests an origin mainly through fractional crystallization from basic magmas (Capaldi et al., 1987a; Chazot and Bertrand, 1995). The granites appear to have been emplaced passively, within the necked areas of the stretched Red Sea continental crust (Geoffroy et al., 1998). Equivalent granites are found near the eastern edge of the Ethiopian plateau and may have the same significance.

2.2.4. End Oligocene to Miocene syn-rift

Syn-rift sedimentary units of the Yemeni Red Sea margin are only found offshore and known from exploration wells (Heaton et al., 1995). Industry reflection seismic data suggest that up to 8 km of Miocene and younger siliciclastic rocks and interbedded evaporites occur in a number of offshore depocenters along the rift margin. The details of the Yemeni margin syn-rift stratigraphy are discussed with the rest of the Red Sea in Section 4.2.

Onshore, the Precambrian basement, pre-rift strata, and Yemen Volcanic Group are cut by abundant mafic and felsic dikes, regionally striking NNW–SSE to NW–SE (Mohr, 1991; Davison et al., 1994; Zumbo et al., 1995). The dikes are often highly weathered and altered. Two dikes striking N60°W from the very southern part of the Yemen volcanic province (Radfan area) yielded ages of 25.4 ± 0.1 Ma ($^{40}Ar/^{39}Ar$; Fig. 4; Zumbo et al., 1995). Four other dikes were also dated: 20.6 ± 0.5 Ma (N–S), 18.5 ± 0.6 Ma (N–S), 18.2 ± 0.5 Ma (N60°W), and 16.1 ± 0.9 Ma (N60°E). From these data Zumbo et al. concluded that no clear change in the stress field from latest Oligocene to Early Miocene could be discerned. In southern Saudi Arabia, the corresponding dike system strikes NNW–SSE and is dated between 24 and 21 Ma (Fig. 4; Section 4.2.1). Integrating both dike and small-scale fault kinematics analyses, Huchon et al. (1991) inferred a weak, early, E–W trending extension coeval with the emplacement of the Oligocene Yemeni Volcanics, followed in the Miocene by NE–SW extension responsible for Red Sea faulting and block tilting. N–S trending extension was also observed and attributed to the influence of the Gulf of Aden rifting.

Apatite fission track data for the Yemeni rift flank indicate the onset of significant exhumation at ~17–16 Ma, during the main phase of Red Sea continental extension (Menzies et al., 1992). This is comparable to other segments of the eastern and western Red Sea margins as discussed in detail below (Section 4.2.2).

The faulting along the Yemeni rift margin and in the exposed fault blocks in the Tihama Plain are typically domino-style, with major faults spaced 1–5 km apart, of consistent dip over large areas (SW or NE), and linked by fault relays and broad accommodation zones rather than distinct transfer faults (Fig. 5a and b; Davison et al., 1994). Toward the edge of the coastal plain and offshore the structure of the rift margin is dominated by widespread halokinesis in the Middle to Late Miocene massive halite (Heaton et al., 1995; Bosence et al., 1998). The prominent Great

Fig. 6. Geology of the Gulf of Aden region: (a) generalized stratigraphic columns and nomenclature of the Yemeni and Somali margins—1, gypsum; 2, shale and shaly intervals; 3, limestone; 4, dolostone; 5, siltstone; 6, sandstone; 7, conglomerate; 8, volcanic rocks; 9, crystalline basement (modified from Watchorn et al., 1998, and Fantozzi and Sgavetti, 1998); (b) simplified geologic map of Yemen and northern Somalia with major structural elements and location of oceanic crust (light shading offshore) and fracture zones in the Gulf of Aden (after Merla et al., 1973; Huchon and Khanbari, 2003; magnetic anomalies from Cochran, 1981; Sahota, 1990; Audin, 1999). BH-1 = Bandar Harshau-1 well; DB-1 = Dab Qua-1 well; SSD = Shukra al Sheik Discontinuity.

Escarpment delineating the eastern border of the Tihama Plain is formed in some areas by terraced tilted fault blocks and in others by resistant granite intrusives along the rift margin (Fig. 5a; Davison et al., 1994, 1998).

Stretching factors for the onshore Yemeni margin of $\beta = 1.6$–1.8 have been calculated using the geometry of fault block rotations (Fig. 5b and c; Davison et al., 1994). Seismic data across the Tihama Plain and offshore show a prominent seaward-dipping top salt reflector that probably masks similar domino style rift faults in the deeper section (Heaton et al., 1995; Davison et al., 1994). Near the coastline there are several emergent salt diapirs (Bosence et al., 1998). The near offshore is characterized by salt diapirs, extensional faults detaching in the salt layers, and salt canopies (Heaton et al., 1995). Further offshore salt cored detachment anticlines occur in the compressional zone of this detached terrane. Using gravimetric data Makris et al. (1991) estimated stretching factors of $\beta = \sim 2.4$ in the offshore region of the margin (Fig. 5c).

2.2.5. Syn-drift Pliocene–Pleistocene sediments

Pliocene–Pleistocene sediments are encountered in wells offshore western Yemen (Heaton et al., 1995). At the end of the Miocene and in the Pliocene a widespread open marine carbonate platform formed offshore with biostratigraphic affinities linked to the Indian Ocean (Heaton et al., 1995). The thickness of this section varies from approximately 500 m in the furthest offshore wells to less than 100 m in the near offshore (Heaton et al., 1995). Onshore wells show dominantly sand and shale fluvial and alluvial sequences in the Pliocene.

2.2.6. Quaternary geology and neotectonics

The Yemeni Red Sea region is still tectonically active with earthquakes (Choy and Kind, 1987) and subdued volcanism on the Yemeni plateau inboard of the rift margin uplift (Capaldi et al., 1983, 1987b; Chiesa et al., 1983). Quaternary alkaline volcanics occur on the islands of Hanish and Zubair offshore and in the Jizan volcanic field on the coastal plain as well as in the Hamdan field on the northern part of the rift margin (Volker et al., 1997).

3. Gulf of Aden

Continental rifting in the Gulf of Aden appears to have slightly pre-dated that in the Red Sea (evidence is discussed in Sections 3.2.1 and 4.2.1). Oceanic spreading is more advanced than in the Red Sea, the spreading center connects directly to the world system of oceanic plate boundaries via the Sheba Ridge, and the spreading center is highly dissected by well-defined transform faults (Fig. 1; Laughton, 1966a,b; Girdler and Styles, 1978; Cochran, 1981; Tamsett and Searle, 1990). Late Neogene kinematics of this plate boundary are therefore better constrained than those of the Red Sea (Le Pichon and Francheteau, 1978; DeMets et al., 1990; Jestin et al., 1994).

The general ENE–WSW trend of the Gulf of Aden is oblique to the NNE–SSW relative plate motion of Arabia with respect to Africa, as indicated by the bathymetric signature of fracture zones shown in Fig. 1. The cause for the orientation of this rift has therefore been of considerable interest (e.g., Malkin and Shemenda, 1991; McClay and White, 1995; Abelson and Agnon, 1997; Manighetti et al., 1997; Dauteuil et al., 2001; Huchon and Khanbari, 2003). The spreading center is known to have propagated toward Afar during its oceanic phase, but much less is known about its early continental history. Less exploration drilling has occurred in the Gulf of Aden than the Red Sea and Gulf of Suez, and the accompanying micropaleontologic age control is therefore less robust.

We discuss the history of the Gulf of Aden in four steps: (1) pre-rift setting; (2) Oligocene–Miocene continental rifting; (3) Miocene initiation of seafloor spreading; and (4) westward propagation of spreading to the Gulf of Tadjoura.

3.1. Pre-rift setting

Similar to the Yemeni Red Sea margin, rock units from Archean to Recent age are exposed along the Yemeni and Somali Gulf of Aden margins (Fig. 6a and b). Pre-rift strata are volumetrically much more important, and, except in the westernmost Gulf, plume-related and syn-rift volcanic units are less common. In addition to Neoproterozoic Pan-African basement fabrics, Mesozoic rift structures are an important pre-Cenozoic aspect of this margin.

3.1.1. Basement and Paleozoic–Mesozoic cratonal strata

The crystalline basement of the Gulf of Aden margins is similar to that exposed along the Red Sea, consisting of Archean and Proterozoic terranes amalgamated during the Neoproterozoic Pan-African orogenies. NW–SE trending transcurrent fault zones are important structural features (Fig. 6b). These faults formed as strike-slip structures late in the Pan-African cycle, and are elements of the Najd fault system that is most strongly developed in the Arabian shield (Fig. 1; Moore, 1979; Stern, 1985; further discussed in Section 4.1.1).

Paleozoic strata are missing from the Somali and Yemeni Gulf of Aden margins. The basement surface is generally peneplained and of low-relief, and this region was most likely emergent throughout the Paleozoic (Abbate et al., 1974; Beydoun, 1978). Erosion continued up to the Triassic. Basement units are unconformably overlain by Lower to Middle Jurassic Kohlan (Yemen) and Adigrat (Somalia) Formation conglomerate, sandstone, marl, and limestone, and Middle to Upper Jurassic Amran (Yemen), Bihen-Dula (western Somalia), and Ahl Medo (eastern Somalia) Group transgressive marl and limestone (Fig. 6a; Beydoun, 1964, 1966; Abbate et al., 1974; Beydoun and Sikander, 1992). The unconformably overlying Cretaceous Tawilah (western Yemen), Fartaq (eastern Yemen), Dhalqut (Dhofar Oman), Jessoma (western

Somalia), and Tisje (eastern Somalia) Groups consist of fluvial sandstone in the west (see Section 2.2.2) but are replaced by marine clastic and carbonate facies in the east (Fig. 6a; Beydoun, 1964, 1978; Al-Subbary et al., 1998; Watchorn et al., 1998). This facies distribution is a regional effect of the accelerated spreading in the Indian Ocean to the east during the Cretaceous (Bosellini, 1986).

Continental crust continues east from Alula and Cape Gwardafuy to Socotra Island, where Precambrian granite, pyroclastic rocks, gabbro, and peralkaline intrusives are exposed (Figs. 1 and 6b; Beydoun and Bichan, 1970). The basement there is overlain by Cretaceous sandstone and limestone, followed by Paleocene and Eocene limestone (Beydoun and Bichan, 1970). The southeast margin of the Socotra platform is marked by a major NE-striking fault, with oceanic crust overlain by a thin section of sedimentary rock lying to the south (Bott et al., 1992).

3.1.2. Mesozoic rifting

The Yemeni and Somali Gulf of Aden margins were affected by several important phases of Mesozoic continental rifting (Figs. 1 and 6b). In western and central Yemen, the Marib-Shabwa (or Sab'atayn) and Balhaf basins are oriented NW–SE (Bott et al., 1992; Brannan et al., 1997), reactivating Najd trend faults. To the east, the Sayut-al Masila, Qishn, and Jiza-Qamar basins are oriented more E–W (Redfern and Jones, 1995; Beydoun et al., 1996; Bosence, 1997). The preserved stratigraphic section at Marib-Shabwa is predominantly Late Jurassic. In eastern Shabwa, Early Cretaceous is also encountered, and increasingly dominates in the Balhaf and Sayhut-Al Masila basins. Jiza-Qamar syn-rift strata are predominantly of Cretaceous age (Brannan et al., 1997). In northern Somalia eroded remnants of several ESE–WNW trending basins are present with Jurassic and Early Cretaceous syn-rift fill (Fig. 6b). These are best preserved east and south of Ras Khanzir, in the Berbera basin or Nogal rift (Bott et al., 1992; Granath, 2001). The Berbera and Balhaf basins were contiguous in the pre-Oligocene plate configuration.

The Gulf of Aden region Late Jurassic rifting was part of the breakup of Gondwana that began in the Permian (Stampfli et al., 2001). Jurassic rifts are also present in the Sudan (Blue Nile Rift; Bosworth, 1992) and Kenya (Lamu embayment; Reeves et al., 1987), and this is the time of the separation between Madagascar and Africa (Somali basin; Rabinowitz et al., 1983; Bosellini, 1992; Guiraud and Maurin, 1992). During the Early Cretaceous, rifting spread throughout much of Africa (reviewed in Guiraud and Bosworth, this volume), and in the Late Cretaceous (circa 80 Ma) transcurrent faulting began between India and Madagascar (Norton and Sclater, 1979; Bosellini, 1992).

In the late Tertiary, Gulf of Aden rifting reactivated many of the Mesozoic rift faults, which helped control the initial break-up geometry of the margin. This is observed in industry seismic data where faults that bounded Mesozoic rifts cut through the overlying Oligocene–Miocene syn-rift stratigraphy (Fig. 7; Tard et al., 1991; Bott et al., 1992). The southern margin of the eastern Gulf of Aden (including the Socotra platform) is much broader than its northern margin (Figs. 1 and 6b). D'Acremont et al. (2005) attributed this asymmetry to the inherited structure of Mesozoic rifting.

3.1.3. Cenozoic pre-rift strata

Paleocene to Eocene strata in eastern Yemen and Dhofar, Oman, are assigned to the Hadhramaut Group and consist of limestone, marl, and gypsum that reach a maximum thickness of ~1770 m (Fig. 6a; Beydoun, 1964; Roger et al., 1989; Haitham and Nani, 1990; Bott et al., 1992). The only equivalent strata along the Yemeni Red Sea margin are the shallow marine limestone beds of the Paleocene Medj-Zir Formation (Section 2.2.2.). The contact between the Hadhramaut and underlying Cretaceous

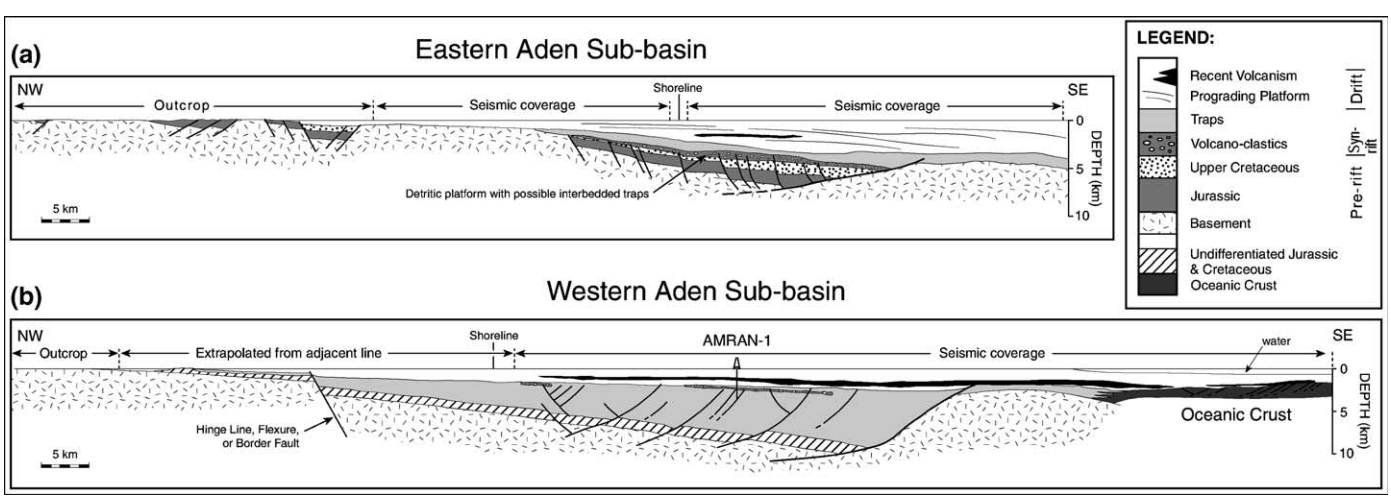


Fig. 7. Structural profiles across the Yemeni margin of the Gulf of Aden incorporating outcrop and reflection seismic data (after Tard et al., 1991): (a) cross-section east of Aden city from Precambrian basement to the edge of the continental margin; (b) cross-section west of Aden city from Precambrian basement to oceanic crust. Locations are shown in Fig. 6b.

Tawilah-Dhalqut Groups is interpreted to be disconformable (Beydoun, 1964) or unconformable (Watchorn et al., 1998; Robertson and Bamakhalif, 1998).

In Somalia, the Paleocene consists of Auradu Formation thick-bedded shallow water platform limestone and dolostone (Fig. 6a; MacFadyen, 1933). The overlying Eocene Allah Kajid, Taleh, and Karkar Formations consist of shale, carbonate, gypsum, and sandstone (MacFadyen, 1933; Bosellini, 1989, 1992).

Paleocene and Eocene strata of the Gulf of Aden margins generally thicken and become more basinal facies to the east (Beydoun, 1970), reflecting continued subsidence along the Indian Ocean margin. The composite Paleocene–Eocene stratigraphic column in Migiurtinia in NE Somalia (Fig. 6b) reaches ~1000 m (Fantozzi and Sgavetti, 1998).

3.2. Oligocene–Miocene continental rifting

The best exposures of Gulf of Aden syn-rift strata in Yemen are found in the Hadhramaut coastal region, in the Balhaf, Masila, Qishn, and eastern Qamar basins (Fig. 6b; Schüppel and Wienholz, 1990; Bosence et al., 1996; Nichols and Watchorn, 1998; Watchorn et al., 1998). Good exposures are also found in the Ashawq graben of the Dhofar region of Oman (Fig. 6b; Platel and Roger, 1989; Roger et al., 1989; Robertson and Bamakhalif, 1998). In Somalia, more extensive sections are present along the margin, particularly in the Boosaaso and associated basins in Migiurtinia in the extreme northeast (Fig. 6b; Azzaroli, 1958; Beydoun, 1978; Abbate et al., 1988; Fantozzi, 1996; Fantozzi and Sgavetti, 1998; Fantozzi and Ali Kassim, 2002). Less aerially extensive but vertically very complete outcrops are also found in the Daban basin in NW Somalia (Fig. 6b; Merla et al., 1973; Bruni et al., 1987).

3.2.1. Rift initiation

When continental rifting along the Gulf of Aden began is generally only approximately known. In Yemen, the syn-rift sediments are assigned to the Shihr Group, and are generally of Oligocene to Miocene age (Beydoun, 1964; Beydoun and Greenwood, 1968; Beydoun and Sikander, 1992). Onshore Yemen, the base of this section is interpreted to be Rupelian (Watchorn et al., 1998), or ~33.9–28.4 Ma. Definitive paleontologic or other dating data, though, were not reported. In Dhofar, Rupelian age strata are referred to the Ashawq Formation (Roger et al., 1989). The Ashawq Formation is divided into two parts, the pre-rift Shizar Member and the conformably overlying syn-rift Nakhlait Member (Robertson and Bamakhalif, 1998). Age assignment is based on benthic foraminifera assemblages (Roger et al., 1989), and is therefore imprecise, but if the Shizar is truly pre-rift, then rifting commenced sometime after (and not at) ~33.9 Ma, and only no later than ~28.4 Ma. However, some authors include the Shizar Member, and even the underlying late Priabonian (~35–~33.9 Ma) Zalumah Formation as syn-rift (Roger et al., 1989; Fournier et al., 2004). Syn-rift strata are also known in offshore exploratory wells in the central Gulf of Aden (Fig. 6b). In the Hami-1 well, offshore Massila basin (central Yemen), and the Bandar Harshau-1 well, ~48°E offshore Somalia, calcareous nannofossil-bearing marine syn-rift strata were encountered overlying pre-rift sabkha and brackish marginal marine deposits. Hughes et al. (1991) interpreted the lowermost syn-rift units to belong to Zone NP24-lower (Martini, 1971), which is between ~29.9 and ~28.7 Ma (late Rupelian; Berggren et al., 1995). These offshore results are compatible with what is presently known from the equivalent onshore outcrops in Yemen and Oman, if one accepts the Shizar Member of the Ashawq Formation as pre-rift.

In northern Somalia, the basal syn-rift strata are Oligocene continental and lagoonal deposits of the Scimis and Scushuban Formations (Azzaroli, 1958), also referred to the Daban series (Fig. 6a; Abbate et al., 1988). These units were deposited in fault-controlled basins, unconformably overlying the Late Eocene Karkar carbonates or older pre-rift section (Fantozzi and Sgavetti, 1998). In the Boosaaso and Qandala basins, strata bearing *Austrotrillina asmariensis*, Adams, (Rupelian–Chattian, ~33.9 to ~23.0 Ma) seal faults within the upper pre-rift sequence (Ali Kassim, 1991, 1993; Fantozzi, 1992; Fantozzi and Sgavetti, 1998). This demonstrates a phase of intra-Oligocene deformation and subsequent deposition, but with significant uncertainty in the precise timing.

These data suggest that rifting was established in the central and eastern Gulf of Aden by the mid-Oligocene (circa 30 Ma), with some pre-rift sedimentation continuing into the earliest Oligocene. No definitive ages are available for the areas west of ~48°E. There is no firm evidence that any extension pre-dated the onset of plume volcanism in the Afar region at ~31 Ma (Sections 2.1.3 and 2.2.3). It is also not established where the Gulf of Aden continental rift first initiated, nor whether its subsequent propagation represented a geologically discernible period of time.

3.2.2. Syn- to post-rift deposition

The syn-rift basins of the Yemeni and Somali Gulf of Aden margins are oriented WNW–ESE to E–W, separated by structural highs that locally expose basement. In Somalia, the basins are asymmetric half-graben, separated by complex accommodation zones that are interpreted to have influenced the later development of Gulf of Aden oceanic fracture zones (Fantozzi and Sgavetti, 1998). Fault throws of several kilometers have been documented (Fantozzi and Sgavetti, 1998). Outcrop fault kinematic analyses from syn-rift strata along the Yemeni margin indicated that the early regional extension direction for the Gulf of Aden continental rift was ~N20°E, highly oblique to the ~N70°E trend of the Gulf (Huchon and Khanbari, 2003). Similar observations were made in the Dhofar region (Lepvrier et al., 2002; Fournier et al., 2004).

In the Yemeni Hadhramaut region (Fig. 6b), the Shihr Group includes three distinct sequences (Watchorn et al., 1998): (1) basal, complexly interfingering alluvial siliciclastics

overlain by dominantly marginal marine carbonates. Gypsum is locally interbedded with limestone and marl in this sequence. Sequence 1 has elsewhere been referred to the Ghaydah and lower Taqa Formations (Fig. 6a) and the maximum thickness measured by Watchorn et al. was ~185 m; (2) laterally persistent alluvial dominated facies changing upwards to coastal salina-sabkha facies. This sequence is equivalent to parts of the Ghaydah and lower Hami Formations (Fig. 6a) and it is approximately 1000–1500 m thick; and (3) post-rift near-shore facies that largely mimic present-day sedimentary environments (alluvial terraces, braided stream complexes, re-deposited gypsum beds). Marine strata in this sequence are restricted to a narrow 100 m to 1 km wide strip adjacent to the present-day coastline. Sequence 3 is equivalent to the Sarar Formation (Fig. 6a). Sequences 1 and 2 were interpreted to be syn-extensional in origin, but interestingly are described as not showing growth relationships to faults nor syn-sedimentary deformation (Watchorn et al., 1998). The ages of sequences 1 and 2 are poorly known (Oligocene–Miocene), but the boundary between sequence 2 and 3 is constrained by $^{87/86}$Sr chemostratigraphy to be between ~21.1 and ~17.4 Ma (Early Miocene, early Burdigalian). These stratigraphic data indicate that the central Yemeni Gulf of Aden margin became stable in the Early Miocene, much earlier than the Afar (Sections 2.1 and 2.2) and Red Sea (Section 4) margins. Watchorn et al. interpreted the sequence 2–3 boundary to represent the rift-to-drift transition.

Offshore from the Hadhramaut region, maxima of about 2000 m and 1100 m of the syn-rift Shihr Group and post-rift Sarar Formation, respectively, were encountered in six exploratory wells (Haitham and Nani, 1990; Bott et al., 1992; Hughes and Beydoun, 1992). Thinner sections of the same age were seen in the two wells drilled offshore central Somalia (Fig. 6b).

Onshore Somalia, a 2500 m syn-rift section is present in the small Daban basin south of Berbera (Fig. 6b; Bruni et al., 1987). The outcrops consist of fan delta complexes interfingering with lacustrine deposits. Source of the siliciclastics was the up-thrown Somali Plateau to the west.

In the Boosaaso basin of Migiurtinia (Fig. 6b), the base of the syn-rift section contains lenses of coarse-grained sandstone and conglomerate bearing basement clasts (Fantozzi and Sgavetti, 1998). These beds are overlain by lagoonal complexes and fluvial facies that pass northward to a marine succession. Similar north–south assemblages of facies are found in the Qandala basin. By contrast, the syn-rift strata of the El Gal basin (Fig. 6b) display southeast depositional dip. The syn-rift section in Migiurtinia has been referred to the Guban series (or group), and is correlative to the Scushuban, Scimis, and Bandar Harshau Formations (Fig. 6a; Bott et al., 1992; Fantozzi and Sgavetti, 1998). Maximum marine incursion occurred in the Miocene. Despite the presence of relatively good exposures in Somalia, the age of the syn-rift section remains poorly constrained (definitively Oligocene and Miocene).

3.3. Miocene initiation of seafloor spreading

Based on bathymetric data, Laughton (1966a,b) first recognized the mid-oceanic ridge/fracture zone morphology of the central Gulf of Aden (Figs. 1 and 6b). Mathews et al. (1967) confirmed that the ridge continued east to the Owen fracture zone and coined the name "Sheba Ridge" for this structure. Magnetic data soon indicated the presence of spreading anomalies north and south of the ridge from the Owen to just west of the Alula-Fartaq fracture zones (Fig. 1), out to anomaly 5 or ~10 Ma (Laughton et al., 1970). Later interpretations (Girdler and Styles, 1978) suggested a more complex history, with spreading initiating in the mid-Oligocene (~30 Ma) and continuing to ~15 Ma, then recommencing at ~5 Ma. Cochran (1981) synthesized the historical and new data and concluded that spreading initiated throughout the Gulf of Aden, from the Owen fracture zone west to fracture zone M (~45°E; also referred to as the Shukra al Sheik discontinuity; Fig. 6b), at about anomaly 5 time. He also recognized that the magnetic quiet zone landward of the correlatable spreading anomalies is not necessarily an isochron (Cochran, 1981, 1982).

Magnetic anomalies near the magnetic quiet zone are difficult to interpret in many areas of the Gulf of Aden. Recent studies suggest that oceanic spreading probably initiated between the Alula-Fartaq and Owen fracture zones at anomaly 5d–5e, or ~19–18 Ma (Fig. 6b; Sahota, 1990; Sahota et al., 1995; Leroy et al., 2004). Anomaly 5c (~16 Ma) is also identified for a few 100 km west of the Alula-Fartaq fracture zone. Spreading therefore propagated from east to west, first splitting oceanic crust east of Ras Sharbithat-Socotra Island (Fig. 1; Stein and Cochran, 1985), and then continental crust throughout the Gulf of Aden. Propagation apparently occurred in a punctiform fashion, stopping for several million years at major lithospheric breaks (Manighetti et al., 1997).

The onset of organized spreading in the central Gulf of Aden by ~16 Ma corresponds well with the interpretation of the rift-to-drift transition at between ~21.1 and ~17.4 Ma along the onshore Yemeni margin (Watchorn et al., 1998; Section 3.2.2). There also appears to have been a rotation of the stress field along the Yemeni margin at this time, to N20°W directed extension (Huchon and Khanbari, 2003). Huchon and Khanbari interpreted this to have resulted from crack-induced tensional stresses associated with N110°W propagation of the Gulf of Aden oceanic spreading center.

3.4. Propagation of seafloor spreading to the Gulf of Tadjoura

The Shukra al Sheik discontinuity, or 'fracture zone M' (Cochran, 1981), lies at the approximate eastern boundary of the Afar plume (Fig. 6b). This discontinuity is not a fracture zone in the usual sense but rather corresponds to a major change in rheology of the continental lithosphere

(Manighetti, 1993; Manighetti et al., 1997; Hébert et al., 2001; Dauteuil et al., 2001), and the anticipated style of rift propagation might therefore be different. The gravity and bathymetric signatures of the area west of the discontinuity are very different than found in the central Gulf of Aden. An axial trough is present, but without offsets by fracture zones (Figs. 1, 6b). Aeromagnetic (Courtillot et al., 1980) and ship data (Girdler and Styles, 1978; Girdler et al., 1980) resulted in various interpretations for the possible age of oceanic crust west of the Shukra al Sheik discontinuity. More recent data have shown that no magnetic anomaly older than anomaly 2 can be reasonably identified west of the discontinuity (Audin, 1999; Hébert et al., 2001). The Sheba Ridge thus stalled at the Shukra al Sheik discontinuity for about 8 My and propagated westward quite recently (since ~2 Ma) into Afar.

Multibeam bathymetric data show that the Gulf of Aden axial trough west of the Shukra al Sheik discontinuity curves gently south and enters directly into the Gulf of Tadjoura (Dauteuil et al., 2001; Audin et al., 2001). Several large submarine canyons were also observed emanating from the Bab al Mandeb strait and entering into the axial trough. The Shukra al Sheik discontinuity itself is not seen in the bathymetric data, nor are any other fracture zone-like candidates, with the exception of the Maskali transform fault in the Gulf of Tadjoura (Manighetti et al., 1997; Dauteuil et al., 2001). A strong WNW–ESE trending fabric is evident, producing rhombic sub-basins (Dauteuil et al., 2001). A simple interpretation of these features is that they represent normal faults formed under NNE–SSW extension, arranged en echelon because they occupy a zone of weakness controlled by sub-crustal magmatism that trends oblique to the imposed extension direction (Dauteuil et al., 2001).

Integration of limited seismic refraction surveys (Laughton and Tramontini, 1969) and gravity data suggest that from the Gulf of Tadjoura east to 44°10′E the lithosphere is of continental type but locally thinned beneath the axial trough (Hébert et al., 2001). From 44°10′E east to 44°45′E (approximate location of the Shukra al Sheik discontinuity) it is of transitional character. East of 44°45′E the lithosphere becomes oceanic with a thermal subsidence period of ~10 Ma.

4. Red sea

Offshore exploration drilling results suggest that continental rifting in the southern Red Sea began a few million years later than in the Gulf of Aden (Hughes et al., 1991). The two arms of the rift system underwent significantly different styles of extension, and seafloor spreading in the Red Sea appears to have initiated ~5–10 My later than the principal Gulf of Aden events. Not all workers support these interpretations (reviewed in Cochran, 1981; Hempton, 1987). Le Pichon and Gaulier (1988) argued that spreading necessarily started at the same time in the Red Sea and the Gulf of Aden, because they postulated that the rift-to-drift transition coincides with an acceleration in plate motion, which must therefore be synchronous in both rifts. There is also debate regarding the impact of pre-rift setting on the early evolution of the basin, the style of syn-rift kinematics, and whether propagation of the basin to the north can truly be discerned.

In the following we discuss: (1) the pre-rift structural and stratigraphic setting of the Red Sea, Gulf of Suez, and Gulf of Aqaba; (2) Late Oligocene to Late Miocene rift evolution; (3) Pliocene to Recent drift phase; and (4) Quaternary geology and neotectonics.

4.1. Pre-rift setting

Unlike the Gulf of Aden, the Red Sea developed purely within an intra-continental setting. At a local (<100 km) scale, faulting in the Red Sea basin was strongly influenced by the geometry of old basement fabrics and terrane boundaries, long-lived discrete structures of many ages, and the composition and thickness of the pre-rift sedimentary cover. Regionally the orientation of the rift appears to have been controlled by far-field stresses discussed below.

4.1.1. Neoproterozoic basement lithologies and structure

The crystalline basement of the Red Sea is exposed along the uplifted Red Sea Hills, southern Sinai Peninsula, and Arabian Shield (Figs. 1 and 8). Lithologies consist of highly foliated granitic gneisses, metasediments, Neoproterozoic calc-alkaline volcanic flows and pyroclastic rocks, Neoproterozoic ophiolites, and several phases of syn-kinematic to post-kinematic granitic intrusions (Greenwood et al., 1976; Hadley and Schmidt, 1980; Stern, 1981; Greenberg, 1981; Stacey and Hedge, 1984). Regional basement fabrics are dominated by N- and NW-trending Pan-African foliations in the basement gneisses, pervasive NE- to ENE-trending vertical to sub-vertical mafic to andesitic post-Pan-African (latest Neoproterozoic to Early Cambrian) dikes, N–S fracture zones that dextrally offset some of the granitic intrusions, and WNW- to NW-trending basement shear zones that are part of the Najd fault system of the Arabian craton (Brown and Jackson, 1960; Moore, 1979; Stern et al., 1984; Stern, 1985; Stoeser and Camp, 1985; Abdelsalam and Stern, 1996; Younes and McClay, 1998; McClay and Khalil, 1998).

In addition to the sub-aerial continental margins, Neoproterozoic granitic gneisses and peridotites are exposed ~50 km offshore on Zabargad Island (Fig. 1; Brueckner et al., 1996). Granitic rocks have been encountered in offshore exploration wells drilled along the Egyptian and Saudi Arabian margins, to a distance of about 20 km from the shoreline (Girdler and Southren, 1987; Bosworth, 1993). Continental basement is also exposed in the Danakil horst of Afar (Figs. 1 and 3c), as emphasized by Freund (1970). The Red Sea therefore contains at least some stretched continental crust along its marine margins. Nevertheless, shoreline-to-shoreline palinspastic restoration of the Red Sea to a pre-rift configuration has been suggested based on plate scale considerations (e.g., McKenzie et al.,

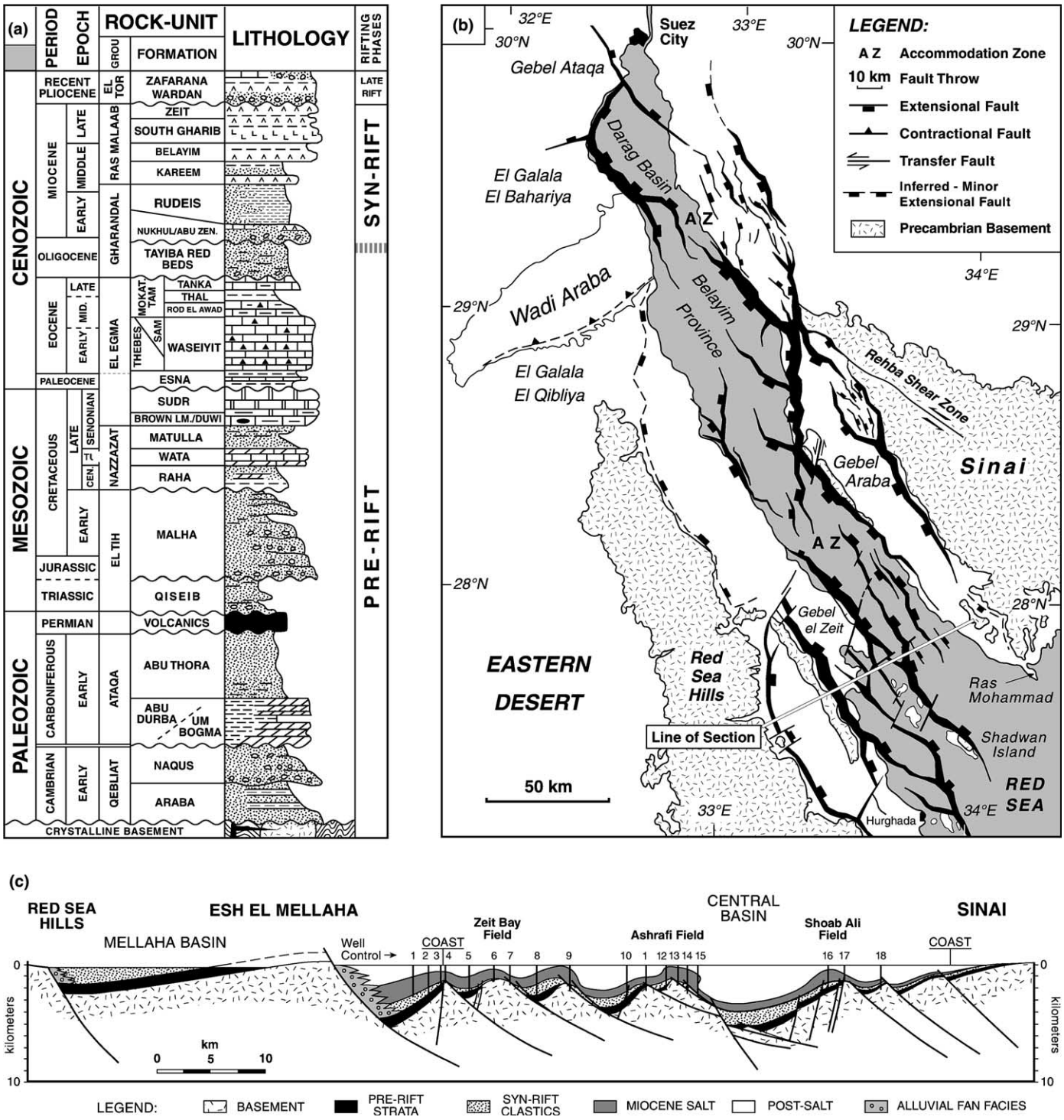


Fig. 8. Stratigraphy and structure of the Gulf of Suez: (a) detailed stratigraphic nomenclature and lithologies (after Darwish and El Araby, 1993; Bosworth and McClay, 2001); (b) principal basement faults, with width proportional to throw (after Khalil, 1998; Bosworth and McClay, 2001). Wadi Araba is a pre-rift, Late Cretaceous age anticline; (c) cross-section across the southern Gulf, illustrating rotated fault block geometry that is thought to be representative of the early phases of Red Sea rifting (after Bosworth, 1994).

1970; Le Pichon and Gaulier, 1988; Bohannon, 1986) and realignment of Precambrian structures and lithologic belts in the Nubian and Arabian shields (Sultan et al., 1992, 1993). Other authors have argued that a gap must exist, the width dependent on just how much stretched continental crust actually exists (Freund, 1970; Joffe and Garfunkel, 1987; Bosworth, 1993, 1994). Plate kinematic constraints from the Red Sea–Gulf of Suez–Gulf of Aqaba junction suggest that the northern Red Sea Saudi Arabian and Egyptian margins can only be restored to about

50–60 km from each other (discussed in Section 5; Courtillot et al., 1987). This remains an important controversy in the study of the Red Sea–Gulf of Aden rift system.

The average strike of the Red Sea Basin from Bab al Mandeb to Suez is ~N30°W (Figs. 1 and 4). The main fault block trends identified in outcrop along the coastal plains and in the subsurface offshore commonly follow this regional orientation (e.g., Garfunkel and Bartov, 1977; Tewfik and Ayyad, 1984; Miller and Barakat, 1988; Bohannon, 1986; Coleman, 1993). However, on Sinai and the Egyptian and northern Saudi Red Sea margins, for long distances the early rift border faults reactivated Najd faults, striking N60–45°W. The border faults maintained the overall N30°W trend by hard linking the Najd faults with the old N–S trending dextral faults, or via other pre-existing structures. The resulting zigzag pattern of faulting (Fig. 8b) is a fundamental attribute of the northern Red Sea and Gulf of Suez (El Tarabili and Adawy, 1972; Garfunkel and Bartov, 1977; Jarrige et al., 1986; Patton et al., 1994; Younes et al., 1998; Bosworth et al., 1998; McClay and Khalil, 1998). On a larger scale, N–S kinks in the N30°W trend of the Nubia Red Sea margin occur at about 15°N, 18°N, and 22°N (Fig. 1). The 15°N kink is a northward projection of Kazmin and Garland's (1973) Neoproterozoic-inherited western Afar border fault (Section 2.1.1). The bends at 18°N and 22°N correspond to northward projections of the Neoproterozoic Baraka and Onib-Hamisana sutures, respectively. The sutures were interpreted to be zones of lithospheric weakness that acted as stress guides, deflecting the early northwestward propagating Red Sea continental rift (Dixon et al., 1987; Kenea et al., 2001). Along the Yemeni southern Red Sea margin, Najd and N–S oriented faults are much less pronounced in the basement, and soft-linkage of syn-rift faults is more commonly observed (Davison et al., 1994).

4.1.2. Structures related to the evolution of Neotethys

Structures that originated during Mesozoic and early Cenozoic deformations exerted lesser, but nonetheless important, influence on the orientation of Red Sea syn-rift structures. The most important of these deformations occurred during the Santonian (~84 Ma) and Late Eocene (~37 Ma) phases of regional compression and transpression (reviewed in Guiraud and Bosworth, this volume). Both were related to events occurring to the north in the Tethyan domain, and for that reason are most strongly developed in the north, particularly in the Gulf of Suez. In the Red Sea basin, their effects are reported south to the latitude of Quseir (Fig. 1; Guiraud and Bosworth, 1997), and Late Cretaceous deformation can also be recognized in Afar (Abbate et al., 2004).

The Alpine deformation resulted in right-lateral transpression along the southern margin of the Neotethys Ocean. This produced the Syrian arc fold belt, which runs across northern Egypt, follows the Levant margin, and turns east in the Palmyrides in Syria (Fig. 1). This produced large E–W to NE–SW trending folds and reverse faults, with underlying uplifted basement blocks (Jenkins, 1990; Moustafa and Khalil, 1990; Moustafa and Khalil, 1995). The basement blocks acted as abutments, blocking the propagation of Gulf of Suez syn-rift faults, particularly at the Wadi Araba anticline (Fig. 8b; Patton et al., 1994; Lambiase and Bosworth, 1995). The Syrian arc also influenced the northern termination of the Red Sea rift system. South of the town of Suez, extension was confined to a single, narrow rift basin, referred to as the Darag basin (Fig. 8b). North of Suez, Neogene extension was spread over a broad area that reaches from the Bitter Lakes west to the vicinity of the Nile delta (Fig. 1), where the Neogene age Manzala rift is now buried by Pliocene, Pleistocene, and Recent sediments (Bosworth and McClay, 2001). Relay of movement between the northernmost Gulf of Suez and Manzala occurred through reactivation of the abundant E–W trending Syrian arc structures of the Cairo-Suez fault zone.

4.1.3. Pre-rift stratigraphy and proto-Red Sea embayments

The pre-rift stratigraphic section of the Red Sea basin varies dramatically along its length, in general decreasing in thickness away from the Tethyan margin north of the Gulf of Suez and the Indian Ocean margin south of Afar. The nature and thickness of the pre-rift strata influenced the composition of the syn-rift fill, the geometry of the initial extensional faulting, and the large-scale morphology of the rift sub-basins. Pre-rift strata were also critical in forming several of the proven hydrocarbon systems of the basin (reviewed in Patton et al., 1994; Lindquist, 1998).

Neogene extensional faulting has disrupted the stratigraphy of the Red Sea region. Outcrops and well-bores tend to provide stratigraphic information about the uplifted footwalls of these blocks, so that estimating the complete stratigraphic section that would be encountered in a basinal, down-thrown structural position is difficult. Nonetheless, this is a useful exercise that has significance to paleo-depositional studies, basin modeling, and reflection seismic interpretation. Fig. 9 is a compilation of outcrop and subsurface data from along the Red Sea rift that attempts to estimate such complete stratigraphic sections for rift-margin settings. Extreme, localized stratigraphic thickness values are not utilized, and the lithologic columns are for positions away from the effects of salt diapirism.

The metamorphic and granitoid pre-rift basement of the Nubian–Arabian shield was extensively peneplained prior to deposition of Cambrian siliciclastic rocks. At the latitude of the central Gulf of Suez, the Cambrian consists of continental to marginal marine sandstone of the Qebliat Group Araba and Naqus Formations (Nubia 'C' of oilfield terminology), which attains a total thickness of about 580 m (Figs. 8a and 9; Hassan, 1967; Said, 1971; Issawi and Jux, 1982). South of the Gulf of Suez, definitive lower Paleozoic strata are not encountered until the southern Saudi Arabian coastal plain, where the Cambrian to Early Ordovician Wajid (Saq) Sandstone attains a thickness of ~200 m (Hadley and Schmidt, 1980).

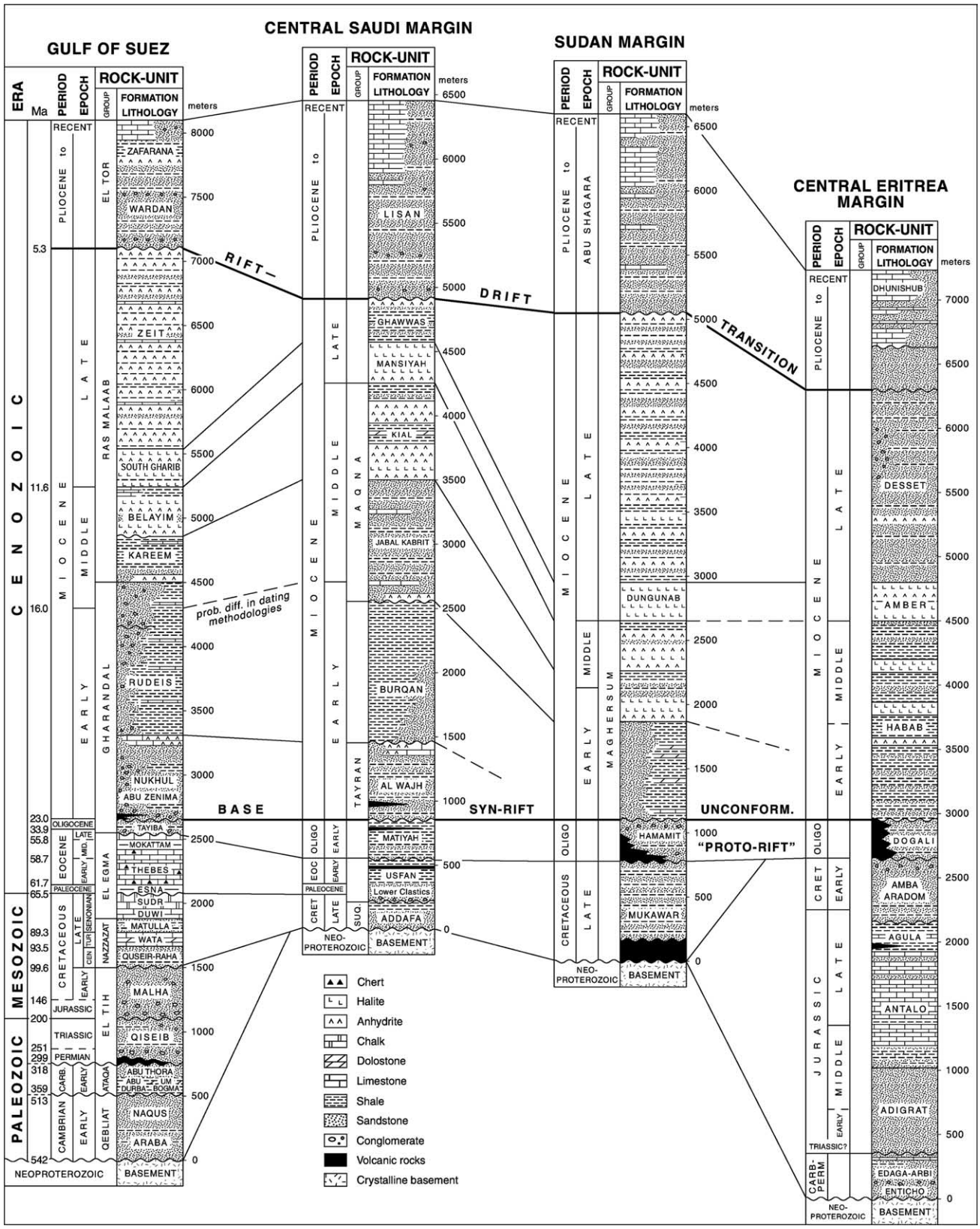


Fig. 9. Representative stratigraphic sections along the length of the Red Sea basin, hung on the base syn-rift unconformity, which roughly coincides with the Oligocene–Miocene boundary. Thickness is composite maximum expected in basinal settings, downthrown to major faults, and away from salt flowage structures. Includes information derived from measured outcrop sections and industry exploration wells. Thickness of the early Late Miocene massive salt (South Gharib Fm., etc.) is assumed in all columns to be 300 m, the approximate average for the southern Gulf of Suez in areas away from salt domes. Sources of data are listed in the text.

The Cambrian sandstones of the Gulf of Suez region are disconformably overlain by up to 240 m of Carboniferous marine dolostone, black shale, mudstone, and sandstone of the Umm Bogma, Abu Durba, and Abu Thora (Nubia 'B') Formations (Figs. 8a and 9; Kostandi, 1959; Hassan, 1967; Weissbrod, 1969; Kora, 1984; Klitzsch, 1990). The upper Carboniferous sandstones are intruded and capped by Permian basalts. Upper Paleozoic strata are not yet proven in the central Red Sea basin. They reappear along the Eritrean margin, where the Carboniferous–Permian(?) Enticho and Edaga-Arbi Formations (Fig. 9) consist of ~340 m of sandstone, shale, and pebbly mudstone that are interpreted to be of glacial or periglacial origin (Dow et al., 1971; Beyth, 1972).

Overlying the Gulf of Suez late Paleozoic sandstones and basalts are red beds and continental sandstones of the Permian(?)–Triassic Qiseib and the Jurassic to Lower Cretaceous Malha (Nubia 'C') Formations of the El Tih Group (Figs. 8a and 9), that attain a total maximum thickness of about 700 m (Abdallah et al., 1963; Barakat et al., 1988; Darwish, 1992; Bosworth, 1995). Equivalents of the Malha Formation, with a thickness of several hundred meters, are interpreted to occur along the Egyptian Red Sea margin as far south as Quseir (Fig. 1; Bosworth et al., 2002), although these unfossiliferous sandstones have also been interpreted as Late Cretaceous in age (Van Houten et al., 1984; Kerdany and Cherif, 1990). Offshore, at Zabargad Island, ~200 m of Lower Cretaceous strata are present (Zabargad Fm.) consisting of alternating beds of marine mudstone, sandstone and silicified limestone (Bosworth et al., 1996).

In the southernmost Red Sea basin, Jurassic strata are the thickest component of the pre-rift stratigraphy (Fig. 9), and are very similar to the rocks outcropping in Afar (Section 2.1.2). Up to ~680 m of fluvial sandstones of the Adigrat Formation were deposited during the Triassic(?) to Middle Jurassic, followed by a major transgression and deposition of ~900 m of the late Middle to Late Jurassic Antalo limestone and ~250 m of the Late Jurassic Agula shale (Beyth, 1972; Savoyat et al., 1989; Bunter et al., 1998). In the latest Jurassic or Early Cretaceous, regression led to a return to continental environments, and deposition of ~500 m of the Amba Aradom sandstone (Savoyat et al., 1989), the correlative of the Malha Formation in Egypt.

The Upper Cretaceous section in the Gulf of Suez and Egyptian Red Sea margin consists of shallow marine strata of the Quseir, Raha, Wata, Matulla, Duwi, and Sudr Formations (Figs. 8a and 9; Youssef, 1957; Ghorab, 1961; Kerdany and Cherif, 1990; Darwish, 1994). The lower units are predominantly shale and sandstone of the Cenomanian to early Senonian transgression, overlain by Campanian and Maastrichtian limestone and chalk. The Upper Cretaceous strata attain a total thickness of about 580 m in the Gulf of Suez. Along the Saudi Red Sea margin (Fig. 9), Campanian to Early Maastrichtian sandstone and siltstone are present at Midyan and Wadi Azlam (Adaffa Fm.; Clark, 1986) and in the vicinity of Jeddah (undiff. Suqah Gp., Hughes and Filatoff, 1995), where the thickness is >165 m. In the Jeddah region these are overlain by Late Maastrichtian bioturbated siltstone and sandstone of the lower Usfan Formation (Basahel et al., 1982). Along the Sudanese margin the Coniacian to Lower Maastrichtian sandstone and shale sequence is referred to the Mukawar Formation and reaches 190 m at its type section (Carella and Scarpa, 1962; Bunter and Abdel Magid, 1989). Cretaceous volcanic units have been encountered below the Mukawar Formation in offshore wells (Fig. 9; Hughes and Beydoun, 1992).

The Paleogene of the Gulf of Suez and Egyptian Red Sea margin (Figs. 8a and 9) consists of the ⩽35 m Esna Shale Formation overlain by thick, often cherty, limestones of the Thebes, Samalut, Waseiyit, Mokattam and other Formations (Said, 1960, 1990; Bishay, 1966; Zico et al., 1993; Darwish and El Araby, 1993). The limestones are locally capped by thin transitional Late Eocene–Oligocene marginal marine to fluvial sandstones and paleosols of the Tayiba Formation (Hume et al., 1920; Hermina et al., 1989; Abul Nasr, 1990). The total thickness of the Paleogene in the northern Gulf of Suez is about 500 m, thinning to ~100 m along the Red Sea Egyptian margin at Quseir.

Along the northern Saudi Arabian margin at Midyan, syn-rift strata rest directly on the Late Cretaceous Adaffa Formation and the Paleogene is absent (Hughes et al., 1999). Near Jeddah, the Late Maastrichtian Lower Usfan Formation continues un-interrupted into the Paleocene middle Usfan carbonate beds, recording a local marine incursion (Fig. 9; Basahel et al., 1982). Marine environments extended south into western Yemen, where the Medj-Zir Formation was deposited (see Section 2.2.2). The Usfan is overlain by continental and, in the upper beds, estuarine to marginal marine clastic units of the Early Eocene Shumaysi Formation (Moltzer and Binda, 1981; Basahel et al., 1982).

Paleogene strata are not yet documented along the Sudanese Red Sea margin, but the Hamamit Formation (Fig. 9) includes units that may range from Paleocene to Miocene in age (Carella and Scarpa, 1962; Bunter and Abdel Magid, 1989). Red sandstones and shales of fluvial and deltaic origin are the principal lithologies, with associated volcanic rocks in the younger sections. Along the Eritrean margin, upper parts of the predominantly Early Cretaceous Amba Aradom Formation may be stratigraphic correlatives of the lower Hamamit (Paleocene?) Formation (Bunter et al., 1998).

In summary, the pre-rift stratigraphic section in the Gulf of Suez, in composite, is ~2600 m thick. In the northernmost Red Sea, the section is normally about 1000 m, with loss of both the basal Paleozoic section and much of the Eocene carbonate. This lesser thickness is maintained along the Saudi Arabian and Sudanese margins. In Eritrea, the section expands to ~3000 m, predominantly due to increase in the Jurassic units. Marine incursions into the central-northern Red Sea region took place in the Early Cretaceous (Zabargad Fm.), Maastrich-

tian-Paleocene (lower to middle Usfan Fm.), and Early Eocene (upper Shumaysi Fm.; see review by Şengör, 2001). Intra-plate rifting probably controlled the first flooding, whereas the latter two were predominantly eustatic in nature. A proto-Red Sea also existed in the south during the Jurassic (see Afar Section 2.1.2), but Jurassic paleo-shorelines trended E–W in Egypt (Abdallah et al., 1963; Guiraud et al., 2001) and no marine incursion occurred from the north.

4.2. Syn-rift evolution

We differentiate three phases of Red Sea continental rifting: (1) Late Oligocene–Early Miocene rift initiation; (2) Early Miocene main syn-rift subsidence; and (3) Middle Miocene onset of the Aqaba–Levant transform. Sediments deposited during the three phases are present throughout most of the basin, from Yemen to Egypt (e.g., Montenat et al., 1986; Hughes and Beydoun, 1992; Coleman, 1993; Bunter et al., 1998).

4.2.1. Rift initiation

The earliest definitive syn-tectonic sediments in Egypt are Abu Zenima Formation red beds that are capped by a ~22 Ma basalt flow in Sinai (Plaziat et al., 1998) and Nukhul Formation dolomitized limestone bearing Aquitanian (~23.0–20.4 Ma) foraminifera in Wadi Nukhul in the central Gulf of Suez (Scott and Govean, 1985) and at Gebel el Zeit in the southern Gulf (Bosworth et al., 1998) (Figs. 8a and 9). Deposition during the Aquitanian was controlled by short-segmented, closely spaced faults (e.g., Gawthrope et al., 1997; Winn et al., 2001), and the rate of rotation of fault blocks peaked at this time (Bosworth et al., 1998). The Abu Zenima-Nukhul section includes chert-cobble conglomerate, channelized sandstone, marine shale, a variety of carbonate facies, and anhydrite (EGPC, 1964; Saoudi and Khalil, 1986) and attains a maximum thickness of ~500 m. In northern Saudi Arabia, Chattian age (~28.4–23.0 Ma) early rift deposits were described from Midyan (Dullo et al., 1983; Bayer et al., 1988; Purser and Hötzl, 1988), but their microfaunas have been re-interpreted as Aquitanian (Hughes et al., 1999). Several species of reworked Chattian foraminifera were also reported from Aquitanian Globigerina marls in a 5 m core at the base of a well at Hurghada Field in the southern Gulf of Suez (El-Shinnawi, 1975). The source strata of these forams have yet to be identified.

Further south in the Jeddah region of the Saudi Arabian margin, siltstones, sandstones, and interbedded basalt flows of the Matiyah Formation (Fig. 9) are paleontologically and radiometrically (K–Ar) dated as Early Oligocene and considered by Hughes and Filatoff (1995) to be early syn-rift or "proto-rift." We are unaware of evidence that this unit is syn-tectonic in origin, and we tentatively correlate it with the pre-rift Tayiba red beds of the Gulf of Suez (Fig. 9). The first definitive syn-rift strata are interbedded sandstones, shales, pyroclastic rocks, and basalts of the Jizan Volcanics and Al Wajh Formations (Hughes and Filatoff, 1995). Radiometric, strontium isotopic, and paleontologic data constrain these units to ~24–21 Ma.

In Sudan, the boundary between pre- and syn-rift strata appears to lie within the ~300 m thick Paleocene(?) to Early Miocene(?) Hamamit Formation discussed in Section 4.1.3 (Fig. 9), although as emphasized by Bunter and Abdel Magid (1989), the position of the unconformity has not been established. The overlying Maghersum Formation (Carella and Scarpa, 1962) or Group (Bunter and Abdel Magid, 1989), which is definitively syn-rift, attains over 2500 m in offshore wells. The oldest strata within the Maghersum are only dated generally as Early Miocene (e.g., Bunter and Abdel Magid, 1989).

The lateral lithostratigraphic equivalent of the Hamamit Formation in the Eritrean Red Sea margin is the Dogali Formation (Fig. 9), which also may in part be syn-rift (Bunter et al., 1998). The overlying definitive syn-rift Habab Formation generally begins in the Early Miocene (Fig. 9), but in the Thio-1 well the basal Habab Formation contains calcareous nannofossils of Martini (1971) Zone NP25 (late Chattian; Hughes et al., 1991). Thio-1 is located offshore from the Danakil horst, south of the Gulf of Zula. Zone NP25 is considered to extend from ~27.5 to ~23.0 Ma (Berggren et al., 1995; Gradstein et al., 2004), bracketing the minimum age of rifting.

All the basal syn-rift strata are associated with localized tholeiitic basaltic magmatism, mostly in the form of dikes that strike parallel to the Red Sea margins (~N30°W; Blank, 1978; Camp and Roobol, 1989; Coleman, 1993). These dikes are K–Ar dated between about 27 and 20 Ma (reviewed in Coleman, 1993; Bosworth and McClay, 2001), but as discussed above (Section 2.1.3), K–Ar dates can be unreliable and $^{40}Ar/^{39}Ar$ methodology is preferred. Unfortunately, few analyses are presently available. The largest concentrations of dikes occur along the Saudi Arabian and Yemeni margin, and on Sinai (Blank, 1978; Coleman et al., 1979, 1980; Baldridge et al., 1991; Davison et al., 1994). Sebai et al. (1991) published 23 ages from the Saudi Arabian dike swarms and associated flows and plutons, which cluster between 24 and 21 Ma (Fig. 4). As noted by these authors, the tholeiitic activity spans nearly 1700 km (more if dates were available for Sinai), and shows no discernible time migration. Coeval volcanism also occurred in the Cairo-Suez fault zone at this time. Three $^{40}Ar/^{39}Ar$ analyses there gave ages between 23.6 and 22.4 Ma (Fig. 4; Kappelman et al., 1992; Lotfy et al., 1995). Further north in Jordan, early volcanic activity in the Harrat ash Shama similarly cluster between 26 and 22 Ma (K–Ar; Ilani et al., 2001). When the Red Sea basin formed at ~24 Ma, it did so very rapidly, analogous to a large, propagating crack. The rift may have linked Afar to northeastern Egypt because of regional stress concentrations at the bend in the continental margin between North Africa and the Levant (Fig. 1; Burke, 1996).

The early Red Sea rift was segmented along-strike into distinct sub-basins with general half-graben form,

separated by transversely-oriented accommodation zones (Jarrige et al., 1990; Bosworth, 1994). The sense of asymmetry of the sub-basins commonly changed across accommodation zones. As in the Gulf of Aden, the location and orientation of the accommodation zones were strongly influenced by pre-existing basement structures, in particular the Najd fault system (Younes and McClay, 2002, and references therein).

The basal syn-rift strata of the Gulf of Suez and Red Sea were deposited on a low relief surface, and generally near sea level (Garfunkel and Bartov, 1977; Sellwood and Netherwood, 1984; Bohannon et al., 1989; Coleman, 1993). With the onset of extension at the Oligocene–Miocene transition (slightly earlier offshore Eritrea), local footwall uplift developed quickly. In the southern Gulf of Suez, rare cobbles of granitic basement are found within the basal Nukhul Formation (Bosworth, 1995; Winn et al., 2001), and granitic cobbles are common in the Al Wajh Formation at Midyan (Hughes et al., 1999). The Hamamit Formation in Sudan similarly contains coarse detritus from uplifted pre-rift lithologies (Bunter and Abdel Magid, 1989). This uplift was localized to the scale of individual fault blocks, and was distinctly different from the later regional rift shoulder uplift and exhumation discussed below.

The tops of the Nukhul Formation in the Gulf of Suez and the Tayran Group along the northern Saudi Arabian Red Sea margin (Figs. 8a and 9) are marked by laterally extensive anhydrite and carbonate beds (Saoudi and Khalil, 1986; Hughes and Filatoff, 1995; Hughes et al., 1999). These units are probably related to a fall in sea level at the end of the Aquitanian (Haq et al., 1987). At the same time, small fault blocks coalesced into stable, large half graben whose crests were capped by the first of several phases of Red Sea carbonate platform development (Bosworth et al., 1998; Bosworth and McClay, 2001). This event is taken to mark the end of rift initiation.

4.2.2. Main rift subsidence

Following the deposition of the Nukhul-Tayran evaporite and carbonate sequence, sedimentation abruptly shifted to predominantly open marine conditions. This was marked by deposition of thick Globigerina-bearing shale, marl, and deepwater limestone (Figs. 8a, 9) referred to the Rudeis Formation in Egypt (EGPC, 1964; Garfunkel and Bartov, 1977), the Burqan Formation in Saudi Arabia (Hughes and Filatoff, 1995), the lower Maghersum Group in Sudan (Bunter and Abdel Magid, 1989), and the Habab Formation in Eritrea (Hughes and Beydoun, 1992; Bunter et al., 1998). These formations attain thicknesses of ~1400 m, ~1100 m, ~800 m, and ~1500 m, respectively (Fig. 9). Block rotation rates decreased somewhat (et al., 1998), but subsidence rates increased markedly (Steckler, 1985; Moretti and Colletta, 1987; Moretti and Chénet, 1987; Evans, 1988; Richardson and Arthur, 1988; Steckler et al., 1988). Most workers place the onset of rapid subsidence within planktonic foraminiferal zone N5 (Blow, 1969) and calcareous nannofossil zone NN2 (Martini, 1971), or at approximately the Aquitanian–Burdigalian boundary (~20.4 Ma; Richardson and Arthur, 1988; Wescott et al., 1997; Hughes et al., 1999).

Basement exposures along the shoulders of the Red Sea and western Gulf of Suez are typically at elevations of 1000 m or more, and in Sinai the highest peaks exceed 2600 m. Apatite fission track analyses from the western margin of the Gulf of Suez indicate that significant rift shoulder basement exhumation, and by inference uplift, occurred at 22 ± 1 Ma, roughly concomitant with the rapid increase in subsidence rates (Omar et al., 1989). On Sinai, four fission track dates obtained by Kohn and Eyal (1981) also cluster at 20–22 Ma, although slightly older dates of 24.4 ± 2.8 and 26.6 ± 3.0 Ma were also obtained. Along the Saudi Arabian Red Sea margin, between Jeddah and Yemen, Bohannon et al. (1989) reported fission track interpretations that suggested rift shoulder erosion and uplift began at ~20 Ma, but with at least 2.5 km of the total occurring after 13.8 Ma. As mentioned in Section 2.2.4, exhumation driven by uplift at ~17–16 Ma was also identified in fission track studies for the Yemeni margin (Menzies et al., 1992, 1997). Along the conjugate margin in Eritrea, Abbate et al. (2002) found a broad range of fission track ages (~10 to ~400 Ma), but modeling suggested a major crustal cooling event driven by denudation at ~20 Ma. Ghebreab et al. (2002) similarly found cooling ages along the Eritrean margin north of Danakil to cluster between 23 and 17 Ma. These studies suggest that the rift shoulders of the Red Sea first showed demonstrable denudation, perhaps localized, at about the same time as the regional dike event (circa 24–23 Ma). With the onset of the main phase of extension and subsidence, about 4 My after the onset of regional rifting, rift shoulder denudation (uplift) developed throughout the basin.

Despite the consistency of apatite fission track studies around the margins of the Red Sea, some data suggest that a phase of basement unroofing began in the rift shoulder of the southern Red Sea of Egypt and in a small area of the NW Gulf of Suez at ~34 Ma (Eocene–Oligocene transition) (Steckler and Omar, 1994; Omar and Steckler, 1995). These authors and others consider this to represent the onset of Red Sea rifting, coeval with the beginning of rifting in the Gulf of Aden. There are no other documented stratigraphic or structural data in the northern Red Sea domain to support this interpretation (Bosworth and McClay, 2001). The Steckler and Omar fission track studies were located near areas of Late Eocene strike-slip faulting (Tethyan-related fabrics) discussed above (Section 4.1.2), and this is a more likely cause of this limited unroofing.

Concomitant with increased subsidence and rift shoulder uplift in the early Burdigalian (circa 20 Ma), rift faults continued their coalescence into more continuous structures (Bosworth et al., 1998; Bosworth and McClay, 2001). Large basins along the western and eastern rift margins were abandoned, and extension began to focus along the rift axis (Bosworth, 1994). These relict basins are now preserved within the uplifted rift flanks and include the Esh el Mellaha

and Duwi basins of Egypt and the Al Wajh and Yanbu basins of Saudi Arabia (Figs. 1 and 8b, c). Volcanism largely ceased in the Red Sea basin during this period (from ~19 to ~13 Ma) (Coleman et al., 1983; Coleman, 1993).

At ~17 Ma, a significant intra-formational unconformity, the "mid-Clysmic" or "mid-Rudeis" event, developed in the central Gulf of Suez basin (Garfunkel and Bartov, 1977; Jarrige et al., 1990; Patton et al., 1994). The section overlying the unconformity is typically much sandier than that below. Numerous structural interpretations have been attached to this change in sedimentation, which was locally associated with uplift followed by cessation of movement on some faults (reviewed in Bosworth and McClay, 2001). This may have been an early far-field effect of the imminent collision between Eurasia and Arabia along the Zagros–Bitlis suture. The mid-Rudeis event was of minor significance in the southern Gulf of Suez and areas further south.

Hughes and Beydoun (1992) have correlated the Burdigalian (~20.4–16.0 Ma) Rudeis Formation Globigerina marls throughout the offshore wells of the Red Sea basin, from Egypt to Eritrea. This deepwater facies is considered by most authors to represent the principal, syn-rift depositional sequence, with subsidence and formation of accommodation space driven directly by extension as envisioned by McKenzie (1978). A short pulse of evaporite deposition, principally anhydrite, typically marks the top of the Rudeis (Fig. 9). This evaporite forms an important lithologic marker throughout much of the Gulf of Suez and Red Sea, and defines the base of the Kareem Formation in Egypt (EGPC, 1964; El Gezeery and Marzouk, 1974) and the Jabal Kabrit Formation in Saudi Arabia (Hughes and Filatoff, 1995). The evaporite beds are overlain by mixed open and marginal marine clastic and carbonate sediments. The Kareem and Jabal Kabrit Formations reach 350 m and 1000 m in thickness, respectively. The Kareem Formation was originally thought to be Lower Miocene (EGPC, 1964; El Gezeery and Marzouk, 1974), but more recent biostratigraphic studies show it to be Langhian (~16.0–13.7 Ma) to Serravalian (~13.7–11.6 Ma), with Langhian faunas also present in parts of the Upper Rudeis (Fig. 9; Andrawis and Abdel Malik, 1981; Scott and Govean, 1985; Richardson and Arthur, 1988; Wescott et al., 1997). Hughes et al. (1992) placed the Gulf of Suez Rudeis/Kareem boundary within the Burdigalian (i.e., pre-16.0 Ma), and followed this interpretation for the Jebel Kabrit Formation (Hughes and Filatoff, 1995). Kareem-Jebel Kabrit correlative units in the Sudanese and Eritrean Red Sea margins are similarly reported to extend from the late Burdigalian to Middle Miocene (Bunter and Abdel Magid, 1989; Hughes and Beydoun, 1992), but detailed biostratigraphic data have not been published for these sections.

4.2.3. Onset of Aqaba–Levant transform boundary

The sinistral nature of the Aqaba–Levant fracture system was early recognized by Lartet (1869), and elaborated upon by Dubertret (1932). Utilizing a variety of stratigraphic, geomorphologic, and paleontologic evidence, Quennell (1951, 1958) distinguished two phases of offset: 62 km in the Miocene and 45 km in the Pliocene to Recent. The circa 24–21 Ma rift initiation dike swarms of Sinai and northwest Saudi Arabia discussed above are offset by the same amount as basement structures, or about 107 km (Quennell, 1951, 1958; Bartov et al., 1980; Eyal et al., 1981). Onset of movement is therefore demonstrably post-Aquitanian, although there is evidence that the position of the faulting probably occupied a zone of inherited weakness that dates back to at least the formation of the Neotethyan Levant margin, or perhaps earlier (De Sitter, 1962; Dubertret, 1970; Dixon et al., 1987; Guiraud et al., 2001).

During the Early Miocene, the Gulf of Suez and Red Sea basins were linked depositionally, and their structural evolution was indistinguishable (Section 4.2.2 and reviewed in Steckler and ten Brink, 1986; Montenat et al., 1988; Bosworth et al., 1998). There is no definitive evidence for movement or subsidence in the Gulf of Aqaba during the Early Miocene.

During the Middle Miocene, at ~14–12 Ma, a major structural event occurred in the southern Gulf of Suez and in Midyan (Fig. 1). In the southern Gulf of Suez, block rotation rates dropped dramatically, and an unconformity developed in many areas (Bosworth et al., 1998). Most of the blocks in the southern Gulf of Suez plunge to the north beneath this unconformity, indicating uplift along what would become the Aqaba transform boundary (Bosworth, 1995). In Midyan, folding and faulting of the stratigraphic section attributed to sinistral shearing began sometime after deposition of Langhian–early Serravalian age strata (Bayer et al., 1986, 1988). At ~13 Ma, a new phase of volcanism abruptly commenced in Harrat ash Shama in Jordan, after quiescence of about 9 My (Ilani et al., 2001). These features suggest that the Gulf of Suez was abandoned as a strongly active rift in the Serravalian, at the end of deposition of the Kareem/Jabal Kabrit Formations, with motion transferred to the new Aqaba transform boundary. The Red Sea switched from rift-normal movement to highly oblique extension parallel to the transform, and Sinai experienced a minor counter-clockwise rotation. This probably produced minor compression and uplift in the northernmost Gulf of Suez (Patton et al., 1994). Coupled with a minor drop in sea level (Haq et al., 1987), this partially isolated the northern Red Sea from influx of marine waters of the Mediterranean, and sedimentation abruptly changed to widespread evaporite of the Belayim Formation (Egypt; EGPC, 1964; Richardson and Arthur, 1988), Kial Formation (Saudi Arabia; Hughes and Filatoff, 1995), and upper Maghersum Group (Sudan; Bunter and Abdel Magid, 1989). These units reach ~400 m, ~700 m, and ~800 m, respectively. Normal marine conditions persisted in the Darag basin at the north end of the Gulf of Suez (Fig. 8b; Hassan and El-Dashlouty, 1970), so the Mediterranean connection was not completely severed.

Steckler and ten Brink (1986) suggested that the cause for the switch in plate boundary configuration in the Middle

Miocene (or possibly earlier, at the mid-Clysmic event) was the inability of the Red Sea rift to propagate northward through the stronger lithosphere of the Mediterranean continental margin. This model also explained why the new Aqaba–Levant transform then remained inland from the Levant margin. Steckler and ten Brink were not aware of the significant Early Miocene rifting that is now buried beneath the modern Nile Delta (Manzala rift of Bosworth and McClay, 2001), nor the fact that the Aqaba–Levant transform does splay into the Mediterranean basin in the vicinity of the Sea of Galilee, and that significant strike-slip deformation is occurring offshore from Israel and Lebanon. The greater strength of the Mediterranean continental margin lithosphere probably did play a role in determining the post-Early Miocene Red Sea plate boundary geometry, but not in as complete a fashion as originally envisioned.

The onset of movement on the Aqaba–Levant transform did not stop subsidence in the Gulf of Suez, but extension was severely reduced. Structural restorations (Bosworth, 1995) and subsidence modeling (Steckler et al., 1988) both indicate ~35 km of total extension has occurred across the southern Gulf of Suez. This equates to $\beta = \sim 1.6$ regionally, but at the rift axis $\beta = \sim 2$. Seismic refraction experiments have similarly observed a 50% reduction in crustal thickness along the axis of the basin (Gaulier et al., 1988).

Along the Eritrean margin deposition of open marine shale locally persisted until the Late Miocene (Savoyat et al., 1989; Hughes and Beydoun, 1992). The connection to the Gulf of Aden/Indian Ocean through Bab al Mandeb must therefore have remained open. By the early Late Miocene (circa 10 Ma), though, massive halite deposition predominated throughout the Red Sea basin (Fig. 9). This is referred to the South Gharib (Egypt; Hassan and El-Dashlouty, 1970; Fawzy and Abdel Aal, 1986), Mansiyah (Saudi Arabia; Hughes and Filatoff, 1995), Dungunab (Sudan; Bunter and Abdel Magid, 1989) and Amber (Eritrea; Savoyat et al., 1989; Hughes and Beydoun, 1992) Formations. Most published data and stratigraphic analyses suggest that the massive salt is Tortonian in age (e.g., Richardson and Arthur, 1988; Wescott et al., 1997), coincident with a very large eustatic drop in sea level (Haq et al., 1987) that far exceeded the magnitude of the previous Serravalian fall. Other authors consider the salt to be partially or entirely Serravalian in age (e.g., Bunter et al., 1998; Hughes et al., 1999). Depositional thickness of the massive salt is difficult to ascertain due to subsequent halokinesis. We assume ~300 m in Fig. 9, but there were certainly large lateral depositional variations.

The nature of Neogene strata within the Gulf of Aqaba is unknown, as no wells have been drilled there. Several significant left-stepping pull-apart basins have been imaged with marine reflection seismic data (Ben-Avraham et al., 1979; Ben-Avraham, 1985; Garfunkel and Ben-Avraham, 2001). Gravity studies indicate that a minimum of 4–5 km of sedimentary rocks are preserved in the three main lows (Ben-Avraham, 1985). Along the Egyptian Sinai coastline, pull-apart basins parallel to the Gulf of Aqaba that developed within the Neoproterozoic basement contain predominantly highly deformed Cambro-Ordovician sandstones of the pre-rift section (Abdel Khalek et al., 1993).

Following initiation of the Aqaba–Levant transform in the Middle Miocene, alkali olivine basaltic volcanism spread from Harrat ash Shama (~13 Ma; Ilani et al., 2001) southward to the other younger Harrats Uwayrid (~12 Ma), Khaybar (~11 Ma), and Rahat (~10 Ma) (Figs. 1 and 4; reviewed in Coleman et al., 1983; Coleman, 1993). Unfortunately, $^{40}Ar/^{39}Ar$ dating is not yet available to better define the details of the timing of this event. Cinder cones associated with the eruption of these Middle Miocene and younger basalts are aligned close to N–S (i.e., underlying feeder dikes are N–S), suggesting that the maximum horizontal stress over a broad region of Arabia was also oriented N–S (Burke, 1996). Reactivation of older N–S trending basement faults was also proposed as a cause for this alignment (Coleman et al., 1983).

The time span represented by deposition of the Middle to Late Miocene massive halite is highly speculative. Sometime in the Tortonian (~11.6–7.2 Ma; e.g., Wescott et al., 1997) or Messinian (~7.2–5.3 Ma; e.g., Richardson and Arthur, 1988), deposition switched to alternating beds of anhydrite, shale, sandstone, and occasional halite and limestone, representing shallow marine to sabkha environments (Hughes et al., 1999). These units are referred to the Zeit (Gulf of Suez), Ghawwas (Saudi Arabia), and Desset (Eritrea) Formations (Fig. 9). In the Gulf of Suez and southern Red Sea this sequence commonly attains ~1500 m thickness, but along the Saudi Arabian margin it is typically ~350 m.

4.3. Mid-ocean spreading and drift phase evolution

Magnetic striping indicative of an oceanic spreading center first appeared at ~5 Ma in the south-central Red Sea, at ~17°N latitude (Roeser, 1975; Searle and Ross, 1975; Cochran, 1983). This was undoubtedly preceded by several million years of less organized igneous activity in the axial region, as may be occurring in the Northern Red Sea at the present time (Bonatti et al., 1984; Bonatti, 1985). Onset of southern Red Sea spreading was quickly followed at ~3 Ma by propagation of the Gulf of Aden spreading center, westward from 'fracture zone M' (Cochran, 1981) toward the Gulf of Tadjura (Section 3.4. above). Synchronously, the rate of slip along the Gulf of Aqaba–Levant transform boundary increased significantly (Quennell, 1958; Freund et al., 1968, 1970).

The pronounced tectonic activity that marked the Miocene–Pliocene transition (~5.3 Ma) was accompanied by an end to widespread evaporite deposition and a major unconformity (top Zeit, Ghawwas, and Desset Formations; Fig. 9). This unconformity is the most readily mappable seismic event in the Red Sea, and corresponds at least partly to the Messinian unconformity of the Mediterranean. Locally, the unconformity bevels through to the crest of salt ridges, which must therefore have been sub-aerially or sub-

aqueously exposed. During the Pliocene the marine connection through Bab al Mandeb deepened and re-established open marine conditions throughout the Red Sea basin.

During the Pliocene, detritus continued to enter the Red Sea basin from the rising rift margins. Most of this sediment was trapped in mini-basins along the coasts of the main Red Sea basin, both offshore and onshore. The mini-basins formed above the flowing Miocene massive halite. This resulted in a sediment-starved axial trough, lying at several kilometers below the elevation of the coastal basins. The elevation differential drove gravitational collapse of the margins via large, listric faults that generally detach near the base of the evaporites (Fig. 10). The up-dip break-aways for these mega-slumps occur either along the present coastlines, or at re-entrants within the basement complex such as at Midyan. The Egyptian and northern Saudi margins of the Red Sea are very straight suggesting that the positions of the coastal break-aways were influenced by underlying late basement-involved normal faults (Bosworth, 1994). Slumping of the post-salt section in the southernmost Gulf of Suez also occurred, driving large slide masses toward the bathymetrically deeper northern Red Sea (Orszag-Sperber et al., 1998).

4.4. Quaternary geology and neotectonics

Rift extensional faults in the southern Gulf of Suez and juncture with the Gulf of Aqaba remained active in the Quaternary and to the present. This is evidenced by continued seismic activity (Daggett et al., 1986; Jackson et al., 1988), uplift of late Pleistocene and Holocene coral terraces (Gvirtzman and Buchbinder, 1978; Andres and Radtke, 1988; Gvirtzman et al., 1992; Strasser et al., 1992; Reyss et al., 1993; Gvirtzman, 1994; Choukri et al., 1995; Bosworth and Taviani, 1996; Plaziat et al., 1998), and development of extension gashes, joints and extensional faults in late Pleistocene sediments (Por and Tsurnamal, 1973; Gvirtzman and Buchbinder, 1978; Bosworth and Taviani, 1996). Based on analysis of borehole breakout data from exploration wells and small-scale fault kinematics, the present-day orientation of the minimum horizontal stress is N15°E, approximately parallel to the Gulf of Aqaba–Levant transform boundary, and not perpendicular to the trend of the rift (Bosworth and Taviani, 1996; Bosworth and Strecker, 1997). West of the Red Sea throughout much of the Sudan, the present-day minimum stress is aligned N–S (Bosworth et al., 1992).

Joffe and Garfunkel (1987) reassessed the plate kinematics of the entire Red Sea region, and concluded that the instantaneous slip velocity (post-5 Ma) between Sinai and Africa in the southern Gulf of Suez is ~1.6 mm/yr, oriented N29°E. Similarly, geometric analysis of the uplifted southern Gulf of Suez coral terraces and associated active faults indicated ~1.0 mm/yr of N15°E extension (~0.75 mm/yr rift-normal component) over the past ~400 ka (Bosworth and Taviani, 1996). Steckler et al. (1988) calculated similar rates (~0.8 mm/yr), but in a direction between N32°E and N90°E.

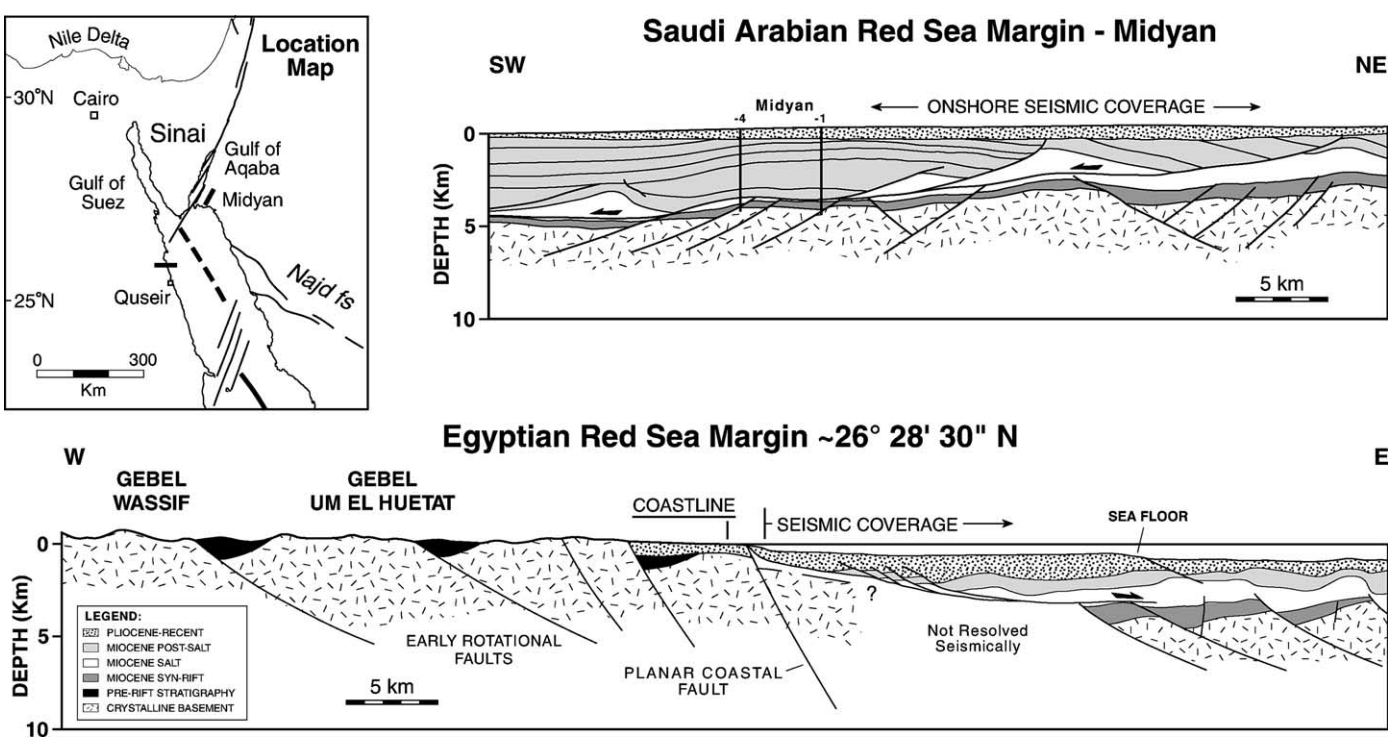


Fig. 10. Structural profiles across the northern Red Sea conjugate margins incorporating outcrop and reflection seismic data: (a) Egyptian bounding fault margin between Safaga and Quseir; with strong rotation of strata into east-dipping faults; (b) Saudi Arabian flexural margin, with gradual west dip into the basin (modified from Mougenot and Al-Shakhis, 1999). Inset map shows locations.

The topographically highest Pleistocene coral terraces occur along the Jordanian and Saudi Arabian margin of the Gulf of Aqaba, up to ~100 m above sea level (Dullo, 1990; Dullo and Montaggioni, 1998). This demonstrates a significant element of uplift along the southern side of the transform plate boundary. Major earthquakes in the offshore are predominantly sinistral strike-slip, although their associated aftershocks can be very complex (Al-Amri et al., 1991; Ambraseys et al., 1994; Hofstetter, 2003; Salamon et al., 2003). Faulted Quaternary alluvial fans along the Egyptian coastline commonly show dip-slip movement, confirming the complex local kinematics of the fault zone (Bosworth and Strecker, 1997; see also discussion in Garfunkel and Ben-Avraham, 2001).

Significant earthquake activity is not recorded in the northern Red Sea south of the Gulf of Suez (Fig. 2), despite continued extension along the rift axis. Both the Saudi Arabian and Egyptian margins are rimmed by several levels of Pleistocene and Holocene terraces, but they do not show significant uplift and generally suggest relative vertical stability (Plaziat et al., 1998).

South of latitude 21°N, significant seismicity is associated with the axial spreading center of the southern Red Sea (Fig. 2; Ambraseys et al., 1994). Offshore from Eritrea in the Dahlak archipelago (Fig. 1), the Pleistocene Dahlak Reef Limestone is highly disrupted by active normal faults (Carbone et al., 1998). Most of this deformation is attributed to the linking of the active Danakil rift with the southern Red Sea spreading center.

Large areas of the Red Sea continental margin alkali basalt province in Saudi Arabia remain volcanically active. Post-Neolithic and historical eruptions are recorded for Harrats Kishb, Khaybar, and Rahat, and Late Quaternary flows are present at these and Harrats al Birk (Jabal al Haylah), Ishara, Kura, and Ithnayn (Fig. 4; Camp et al., 1987; Camp and Roobol, 1989; Camp et al., 1991; Coleman, 1993).

5. Plate scale considerations

Since the pioneering work of McKenzie et al. (1970), Chase (1978), and Le Pichon and Francheteau (1978), the recent kinematics of the Afar triple junction have been refined (e.g., Joffe and Garfunkel, 1987; Jestin et al., 1994), but not fundamentally modified. Relative plate motions can be deduced from analysis of the Gulf of Aqaba–Levant transform (~14 Ma to present), fracture zones in the Gulf of Aden (starting at ~19–18 Ma in the east; ~10 Ma in the west), and from real-time geodetic measurements. Geologic estimates of crustal extension and outcrop fault kinematic data can be used for the early phases of rifting.

At the southwestern corner of the Arabian Peninsula (12.5°N, 43.5°E), global circuit closure (Chu and Gordon, 1998) and GPS measurements (McClusky et al., 2003) indicate that the present-day motion of Arabia with respect to Africa (Nubia) is 1.7–2.0 cm/yr toward azimuth N48°E. Motion of Arabia with respect to Somalia is 1.7 cm/yr toward N37°E (Jestin et al., 1994), indicating slight divergence between Africa and Somalia. The global plate motion model NUVEL-1 predicts convergence between Arabia and Eurasia along the Zagros-Bitlis main fault at 35°N, 47°E of ~2.8 cm/yr toward N11°W (DeMets et al., 1990). Moving east to about 29°N, 55°E this increases to ~3.2 cm/yr toward due north. GPS measurements record relative motion in approximately the same directions, but with only about 70% of the velocity (McClusky et al., 2003).

In the eastern Mediterranean, NUVEL-1 predicts convergence between Africa and Eurasia at the subduction zone at Crete of 1.0 cm/yr toward N7°W. The rate of convergence increases to about 1.4 cm/yr due north further east near the Levant margin (Jackson, 1993). The difference there between Africa–Eurasia and Arabia–Eurasia convergence is ~1 cm/yr, which manifests itself as sinistral movement on the Aqaba–Levant transform boundary. This is a very similar average rate to the geologic estimate by Quennell (1951, 1958) that ~45 km of slip has occurred during the Pliocene to Recent (~5.3–0 Ma). GPS measurements give a much more complicated picture for the eastern Mediterranean (McClusky et al., 2003), largely due to deformation around the Anatolian plate that NUVEL-1 does not resolve.

Other constraints are available concerning the opening history of the Red Sea–Gulf of Aden rift system. Approximately 35 km of total extension has occurred across the southern Gulf of Suez (Section 4.2.3). Unlike other segments of the rift system, this estimate is constrained by numerous exploration wells that frequently reach crystalline basement and by refraction and reflection seismic data. Total offset along the Aqaba–Levant transform boundary is constrained to be ~107 km (62 km Miocene, 45 km Pliocene to Recent; Section 4.2.3). Cochran (1981) calculated post-anomaly five spreading rates of 2.8 cm/yr for the eastern Sheba Ridge, 2.1 cm/yr for the eastern Gulf of Aden, and 2.0 cm/yr for the central Gulf of Aden. During the past 10 My, this has resulted in ~250 km of oceanic crust being accreted to the central Sheba Ridge, and during the 8–9 My prior to this, ~50 km of oceanic crust was added to the ridge east of the Alula-Fartaq fracture zone (Fig. 6b; Section 3.3). In the south-central Red Sea, ~85 km of oceanic crust was formed over the past 5 My (Roeser, 1975; Searle and Ross, 1975; Cochran, 1983).

The tectono-stratigraphic sequences of the Red Sea and Gulf of Aden can also be correlated to important events at other boundaries of the Arabian lithospheric plate (e.g., Hempton, 1987; Joffe and Garfunkel, 1987; McQuarrie et al., 2003; Burke, 1996). This provides a geodynamic context in which to view extension and magmatism within the rift system, and offers insight into why events occurred when they did. For example, our interpreted age of initiation of the Aqaba–Levant transform boundary (~14 Ma) corresponds to the collision of Arabia with Eurasia (Hempton, 1985, 1987; Dercourt et al., 1986; Savostin et al.,

Table 1
Restoration parameters used in Fig. 11

Map age (Ma)	Time interval (My)	Central Zagros convergence [km; (cm/yr)]	Eastern Mediterranean convergence [km; (cm/yr)]	Gulf of Suez extension [km; (cm/yr)]	Aqaba-Levant sinistral slip [km; (cm/yr)]	Sheba Ridge spreading [km; (cm/yr)]	S-Central Red Sea spreading [km; (cm/yr)]
31							
27	4	120 (3)	56 (1.4)				
24	3	90 (3)	42 (1.4)				
14	10	300 (3)	140 (1.4)	21 (0.2)		110 (2.2)[a]	
10	4	120 (3)	56 (1.4)	4 (0.1)	28 (0.7)	120 (3)	
5	5	150 (3)	70 (1.4)	5 (0.1)	34 (0.7)	125 (2.5)	
0	5	150 (3)	70 (1.4)	5 (0.1)	45 (0.9)	125 (2.5)	85 (1.7)[b]
+10	10	300 (3)	140 (1.4)	0	90 (0.9)	250 (2.5)	170 (1.7)

[a] Spreading at the Sheba Ridge is assumed to start at 19 Ma. Spreading rates are full rates. See text for explanation.
[b] At the north end of the Red Sea, the interpreted extension rate reduces to ~1.0 cm/yr.

1986), which would have resulted in a major change in the boundary forces operating around the Arabian plate.

Utilizing this information about present and past movements around the Arabian plate (summarized in Table 1) and regional tectonic events, we have reconstructed the opening history of the Red Sea–Gulf of Aden rift system since the late Early Oligocene (Fig. 11). The reconstruction starts at 31 Ma, the onset of plume volcanism in the greater Afar region (Fig. 11a; Sections 2.1.3 and 2.2.3). Rifting began in the Gulf of Aden by ~29.9–28.7 Ma, and in the southernmost Red Sea by ~27.5–23.8 Ma (Sections 3.2.1 and 4.2.1). The map chosen to depict this phase is 27 Ma (Fig. 11b). Rifting then spread rapidly throughout the Red Sea at the Oligocene–Miocene transition (24 Ma; Fig. 11c). The orientation of the Red Sea was essentially parallel to the distant Urumieh-Doktar arc, perpendicular to extensional stresses generated by the subducting northeastern Africa–Arabia plate (Davison et al., 1994; Burke, 1996). In the early Middle Miocene, the Gulf of Aqaba–Levant transform boundary initiated (14 Ma; Fig. 11d; Section 4.2.3). By this time, some oceanic spreading was occurring at the eastern Sheba Ridge, in particular east of the Alula-Fartaq fracture zone. Spreading propagated through most of the Gulf of Aden in the early Late Miocene (10 Ma; Fig. 11e). At the Miocene–Pliocene transition, spreading initiated in the south-central Red Sea (5 Ma; Fig. 11f). Fig. 11g is the present plate configuration, at the same projection as Fig. 1. A future situation, in which the Arabian Gulf has disappeared and an organized oceanic spreading center has connected the Gulf of Aden through the Gulf of Tadjoura, Danakil Depression, and Gulf of Zula is the final map (+10 Ma).

6. Discussion

The coincidence between hotspots, continental flood basalts, and rifts has long been recognized (Morgan, 1971; Burke and Dewey, 1973). The precise role of hot plume material impinging the continental lithosphere in the development of rift triple junctions has therefore received considerable attention. Active rifting processes, in which lithospheric extension is driven by the energy of a convecting plume, were initially favored (e.g., Burke and Şengör, 1978; Bott, 1992). Other analyses suggested a scenario in which both active and passive processes are involved: hotspots focus the break-up but require additional, coeval extensional plate-boundary generated forces (Hill, 1991; Courtillot et al., 1999). The strongest plate-boundary forces are generally thought to be generated by slab pull in subduction zones and ridge push from oceanic spreading centers, but the ultimate controls of plate motion are still controversial (reviewed by Wilson, 1993). The geometry of break-up may also be influenced at a variety of length scales by pre-existing heterogeneities in the continental lithosphere, as discussed above.

In the Red Sea–Gulf of Aden rift system, we have the opportunity to assess a rifting history that includes both a well-defined plume, subsequently split by extensional structures, and segments of the same rifts that are far removed from significant coeval volcanism. The timing of the various phases of this plume-rift system is now relatively well known, particularly in comparison to other ancient rift systems.

The pre-plume setting of Afar was generally low-relief, and at or near sea level (Cochran, 1981; Coleman, 1993). Mid-Oligocene (late Rupelian) marine strata are preserved in the central Gulf of Aden (Hughes et al., 1991). There therefore was not any advance signal of the imminent arrival of the Afar plume in the form of regional doming, at least in terms of the presently known stratigraphic constraints. Eruption of the Ethiopian, Sudan Derudeb, and Yemeni basaltic trap series began at ~31 Ma (Baker et al., 1996; Hoffmann et al., 1997; Chernet et al., 1998; Kenea et al., 2001; Ukstins et al., 2002). The bulk of the basaltic rocks was erupted within ~1 My, and subsequently accompanied by rhyolitic volcanism. Most workers report that very little extension accompanied the plume-related volcanism.

The central Gulf of Aden offshore Hami-1 and Bandar Harshau-1 wells encountered syn-rift sediments dated between ~29.9 and ~28.7 Ma (late Rupelian; Hughes et al., 1991). These are the oldest, tightly age-constrained

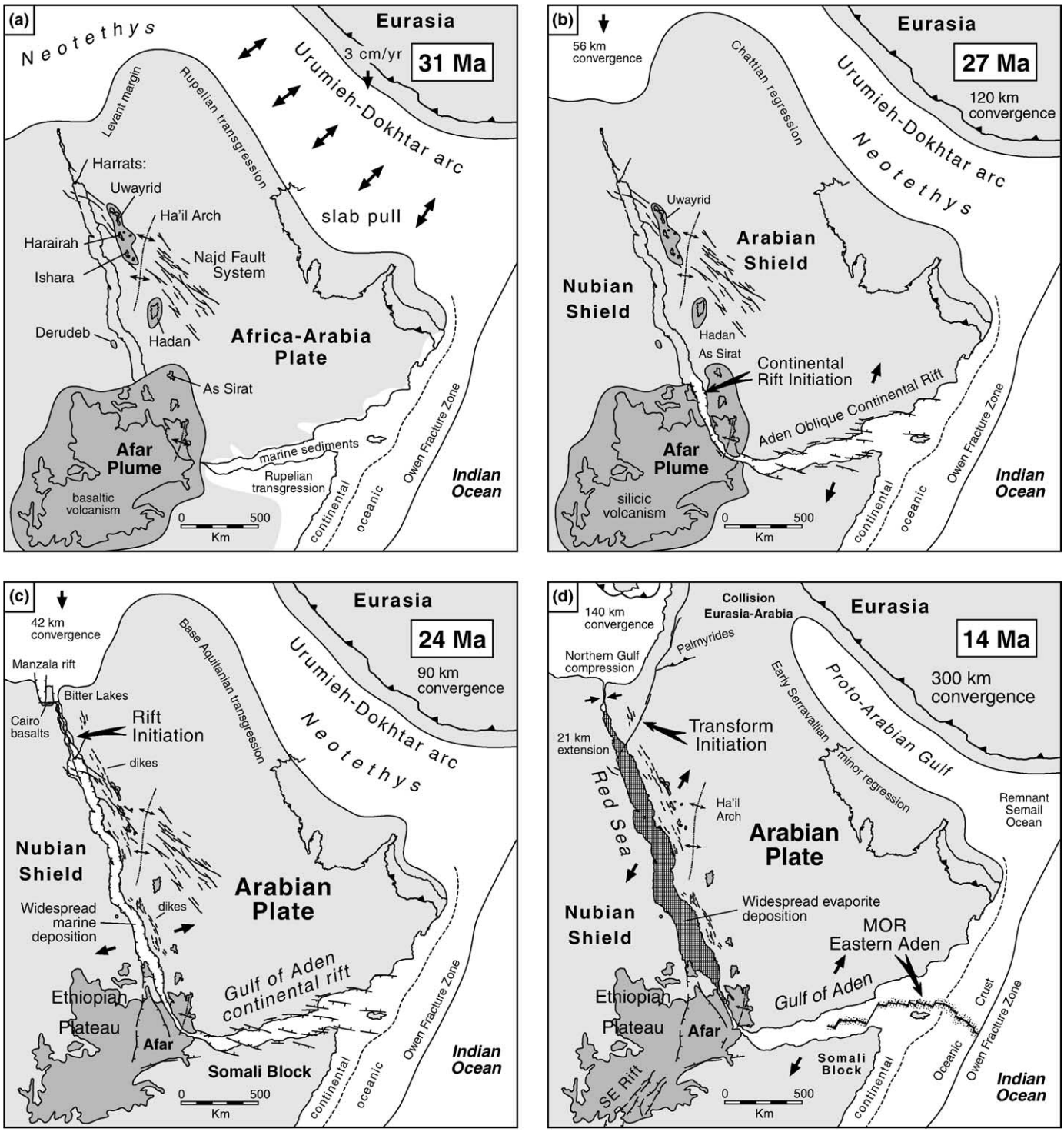


Fig. 11. Palinspastic restoration of the Red Sea–Gulf of Aden rift system: (a) 31 Ma (Early Oligocene) Afar plume initiation and formation of the Older Harrats; (b) 27 Ma (Late Oligocene) initiation of Gulf of Aden and Eritrean Red Sea continental rift; (c) 24 Ma (Oligocene–Miocene transition) rapid spread of extension throughout Red Sea; (d) 14 Ma (Middle Miocene) onset of Aqaba–Levant transform boundary and by-passing of Gulf of Suez basin; oceanic spreading started on eastern Sheba Ridge by ~19 Ma; (e) 10 Ma (early Late Miocene) rift–drift transition in central Gulf of Aden; (f) 5 Ma (Miocene–Pliocene transition) rift–drift transition in south-central Red Sea, jump of spreading west of 45°E in Gulf of Aden (perhaps ~3 Ma), and acceleration of Aqaba–Levant movement; (g) Present-day situation; (h) +10 My future geometry of the Arabian plate. Assumptions for timing and plate kinematics are outlined in the text. A summary of the kinematic restoration parameters used between each step are listed in Table 1. Dark gray onshore = volcanic rocks; stipple = Afar oceanic crust >10 Ma; medium gray offshore = oceanic crust ⩽10 Ma; cross-hatching = evaporite deposition.

syn-tectonic strata in the central or western Gulf of Aden. A similar sequence of marine syn-rift strata is present offshore Eritrea, deposited between ~27.5 and 23.8 Ma (Chattian; Hughes et al., 1991). Rifting is interpreted to

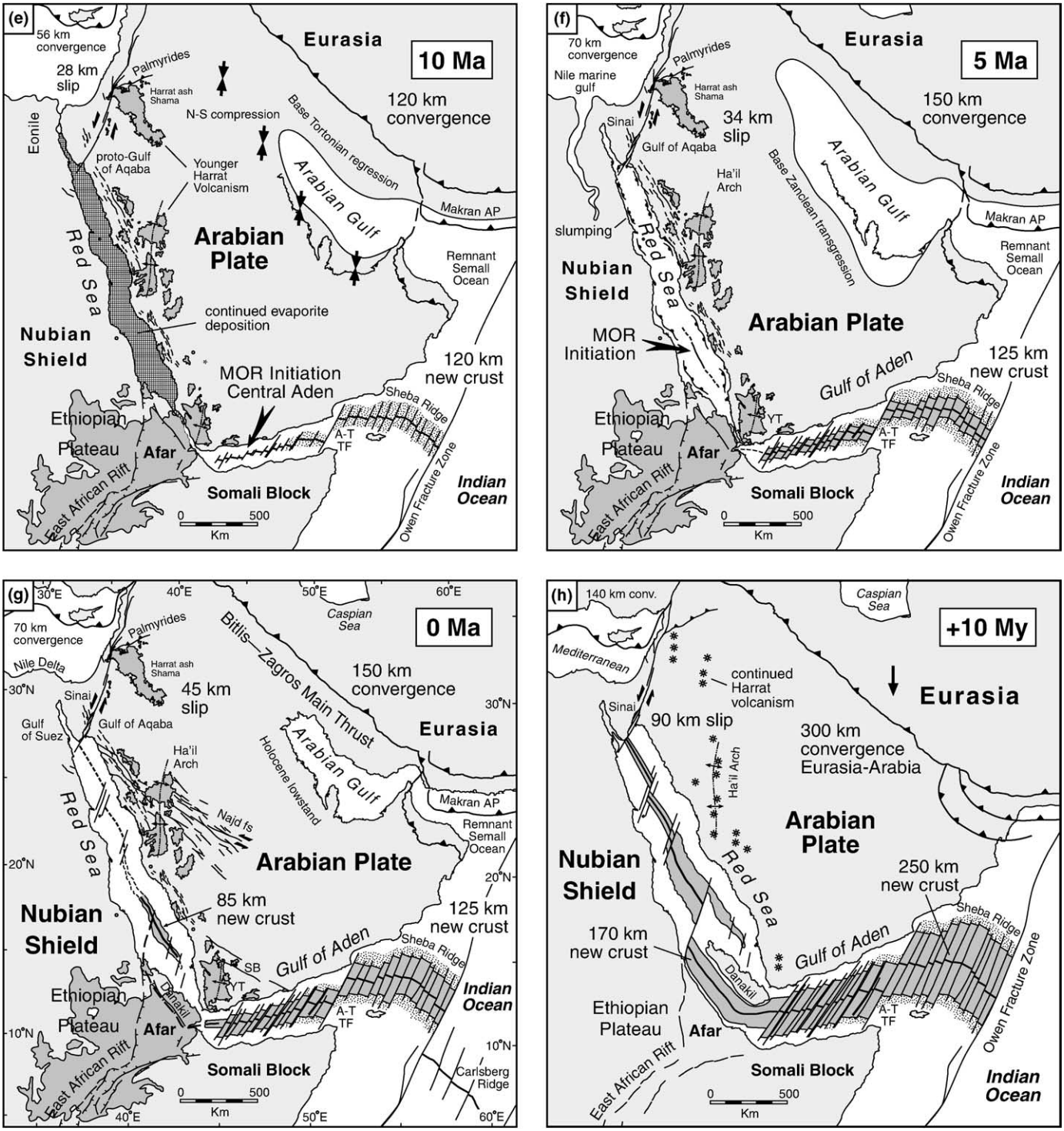


Fig. 11 *(continued)*

have started in the Afar Depression after ~25 Ma, based on volcano-tectonic relationships in rocks constrained by K–Ar age determinations (Barberi et al., 1972b, 1975; Zanettin et al., 1978). Afar extension may actually have started somewhat earlier, as the early syn-rift section is largely covered by younger volcanics (Kenea et al., 2001). Emergence of the Horn of Africa in the Late Oligocene (Azzaroli, 1968) also suggests that the southern Gulf of Aden rift margin was beginning a phase of tectonically driven denudation and uplift. These data indicate that faulting and significant subsidence occurred in the Gulf of Aden and southernmost Red Sea within a few million years of the onset of plume volcanism, but not prior to the plume reaching the surface. In the Afar region of both Yemen and Ethiopia, the sequence as presently constrained was onset of magmatism–onset of extension (with local uplift

and continued magmatism)–major uplift (with continued extension and major subsidence). As emphasized by Menzies et al. (1992), progressions of this type do not fit traditional models of either active (uplift–magmatism–rifting) or passive (rifting–uplift–magmatism) rifting. For most of the Red Sea, the sequence was synchronous extension and magmatism followed by uplift, but the volumes of volcanic rock were small.

Did the Afar plume cause Gulf of Aden and Red Sea rifting to occur? There is some evidence that the Gulf of Aden continental rift initiated first in Oman in the Early Oligocene (Roger et al., 1989; Platel and Roger, 1989), and then propagated to the west at the Rupelian–Chattian transition (see also Huchon and Khanbari, 2003). It is possible, therefore, that arrival of the plume and first rupturing of the lithosphere were coeval, but that the rift did not reach Afar itself until ±1 My into the eruption sequence, at about the onset of rhyolitic eruptions. Prior to the impingement of the Afar plume, NE Africa–Arabia was experiencing NNE–SSW directed extensional forces due to slab pull in the subduction zone beneath the Urumieh-Dokhtar volcanic arc on the facing margin of Neotethys (Fig. 11a; e.g., Davison et al., 1994; Burke, 1996; McQuarrie et al., 2003). The Indian plate may also have been dragging Arabia northward along a 'sticky' Owen fracture zone (Cochran, 1981), though this was undoubtedly a subsidiary force in comparison to slab pull. The Afar plume may have provided the stress concentration or local weakening of the continental lithosphere (e.g. Malkin and Shemenda, 1991; Bellahsen et al., 2004) that triggered a rifting event that was already ready to occur.

In this scenario, the Gulf of Aden continental rift started in the western Indian Ocean, perhaps controlled by the paleo-position of the Carlsberg Ridge (Manighetti, 1993; Manighetti et al., 1997; Hubert-Ferrari et al., 2003). It propagated toward Afar as an oblique rift, with NE to NNE directed extension. For a few million years, the rift stopped in the vicinity of Afar and offshore Eritrea, allowing regional plate stresses to build. At ~24 Ma, the rift abruptly propagated north to the Mediterranean Sea via the Red Sea, Gulf of Suez, Bitter Lakes region, and Manzala rift. The Red Sea opened perpendicular to its axis, and except for irregularities (~100 km and smaller length scales) controlled by reactivation of basement structures, its orientation was dictated by far-field, plate boundary generated stresses (e.g., Dixon et al., 1987). Red Sea dikes associated with the rift propagation were caused by rupturing of the lithosphere, and confirm the early extension direction (N60°E). Various models that invoke an early phase of oblique Red Sea extension or pull-apart basin formation (e.g., Jarrige et al., 1986; Hempton, 1987; Makris and Rihm, 1991; Rihm and Henke, 1998) are not supported by the dike orientations or the preponderance of outcrop data (reviewed in Bosworth et al., 1998; Bosworth and McClay, 2001).

In summary, the Red Sea–Gulf of Aden Rift System appears to be the product of a complex interplay between active and passive rift processes, as envisioned by Hill (1991) and Courtillot et al. (1999). Recent geologic and geophysical studies have helped clarify this model, which will undoubtedly continue to evolve in the future.

7. Conclusions

- The pre-rift basement structure of the Red Sea, Afar, and western Gulf of Aden was predominantly established during the Neoproterozoic Pan-African orogenies, with lesser modifications during the Late Cretaceous (early Alpine-Tethyan) and Eocene (late Alpine) in the north and Jurassic and Cretaceous (Indian Ocean rifting) in the south. Basement structures were demonstrably reactivated at the scale of kilometers to tens of kilometers, and terrane boundaries and regional faults had significant influence at larger scales.
- The composite pre-rift stratigraphic section in the Gulf of Suez is ~2600 m, thins to ~1000 m in the northern Red Sea, and expands again to over 3000 m in the southern Red Sea. In the Danakil Alps, the pre-rift section exceeds 4000 m. This distribution of pre-rift strata reflects the effects of proximity to the subsiding Paleo- and Neotethyan margins in the north (Early Paleozoic to Mesozoic), and the development of the Indian Ocean margin in the south (Mesozoic).
- During the Paleogene, most of the area of the later Red Sea–Gulf of Aden rift system was at or near sea level. Marine incursions reached the central Red Sea region during the Early Cretaceous and Early Eocene. Early Oligocene marginal marine pre-rift strata are present in Dhofar (Oman), and offshore Somalia and Yemen. Stratigraphic evidence, although not conclusive, suggests that regional doming did not occur prior to the eruption of the Afar flood basalts and subsequent rifting, or that if it did occur, it was limited to the immediate region of central Afar.
- Plume related basaltic trap volcanism initiated at Afar, Derudeb in NE Sudan, and SW Yemen at ~31 Ma. Rhyolitic volcanism commenced at ~30 Ma. Volcanism thereafter spread northward to Harrats Sirat, Hadan, Ishara-Khirsat, and Ar Rahat in western Saudi Arabia. This early magmatism occurred without significant regional extension, and continued to ~25 Ma.
- Starting after ~30 Ma, marine syn-tectonic sediments were deposited on rifted continental crust in the eastern and central Gulf of Aden, and by ~27 Ma a small rift basin was forming in the Eritrean Red Sea. Significant extension commenced in Afar by ~25 Ma, but the constraints on this are limited.
- The Gulf of Aden rift connected the Owen fracture zone in the east (oceanic lithosphere) with the Afar plume in the west (continental lithosphere). Extension direction apparently changed through time but was always highly oblique to the general trend of the rift.

- By ~24 Ma (slightly older in southernmost Yemen), a new phase of volcanism, principally basaltic dikes but also layered gabbro and granophyre, appeared nearly synchronously throughout the entire Red Sea, from Afar and Yemen to northern Egypt. This second phase of magmatism accompanied strong rift-normal Red Sea extension and deposition of syn-tectonic sediments, mostly of marine and marginal marine affinity. This marks the formation of the greater Red Sea–Gulf of Aden rift system. Localized rift shoulder uplift commenced at this time in the Red Sea, and greatly accelerated a few My later at ~20 Ma.
- The site of Red Sea rifting was controlled by the position of the Afar plume, but its N30°W orientation was determined by regional plate stresses generated by slab-pull in the subduction zone beneath the Urumieh-Doktar volcanic arc on the north side of Neotethys (present-day Bitlis-Zagros thrust zone). The position of the northern Red Sea may also reflect the guiding effects of a stress concentration at the corner between the Egyptian and Levant continental margins.
- Mesozoic rift faults were reactivated during initiation of the Gulf of Aden and helped control the geometry of the new passive continental margin. The overall Aden rift system orientation, however, cut through these older basins. The geometry of the early Red Sea rift was strongly influenced by pre-existing, predominantly Neoproterozoic, basement structures, and as a consequence followed a complex path from Afar to Suez. Pre-rift structures also played an important role in localizing extension in Afar.
- Pre-rift and early syn-rift structures can be restored to their original contiguous geometries along both the Red Sea and Gulf of Aden conjugate margins. The initial rift basins were generally asymmetric half-graben and 60–80 km in width.
- Stretching factors of the continental crust are estimated to be $\beta = \sim 1.15$ in the onshore Yemen central Gulf of Aden margin, 1.6–1.8 in the onshore Yemen Red Sea margin, ~1.6 in the southern Gulf of Suez regionally and ~2 in the basin axis, ~2.4 in the offshore Yemen Red Sea margin, and up to 2.5 in the Danakil Alps. The syn-rift phase lasted from ~30 to ~19 Ma in the Gulf of Aden, ~27 to ~5 Ma in the southern Red Sea, and ~24 to ~14 Ma in the Gulf of Suez. In the northern Red Sea, continental extension is continuing at the present in the offshore domain.
- By ~19–18 Ma, oceanic spreading initiated on the Sheba Ridge between the Owen and Alula-Fartaq fracture zones. Spreading appears to have propagated a few 100 kilometers west of the Alula-Fartaq fracture zone by ~16 Ma.
- At ~14 Ma, a transform boundary cut through Sinai and the Levant continental margin, linking the northern Red Sea with the Bitlis-Zagros convergence zone. This corresponded with the collision of Eurasia with Arabia, and resulted in a new plate geometry with different boundary forces. Slab pull did not terminate but its strength may have decreased. Red Sea extension changed from rift normal (N60°E) to highly oblique and parallel to the Aqaba–Levant transform (N15°E). Gulf of Suez extension slowed dramatically, but did not cease altogether.
- Coincident with onset of the Aqaba–Levant transform, the area north of Suez was gently uplifted, perhaps due to minor compression of the Sinai sub-plate. Influx of Mediterranean Sea water was greatly reduced, although not terminated. Flow through Bab al Mandeb was also restricted, and Red Sea sedimentation changed from predominantly open marine to evaporitic. A third phase of magmatism commenced, again predominantly in western Saudi Arabia but extending north to Harrat Ash Shama and Jebel Druse in Jordan, Lebanon, and Syria.
- Salt diapirism began in the Red Sea soon after the widespread deposition of massive halite in the Middle to early Late Miocene. Salt domes reached the surface in many parts of the basin by the end of the Miocene, and most were subsequently buried. A few domes are presently at or very near the surface along the Egyptian, Saudi Arabian, and Yemeni margins.
- At ~10 Ma, oceanic spreading in the Gulf of Aden rapidly propagated over 400 km west to the Shukra al Sheik discontinuity at ~45°E longitude.
- Oceanic spreading followed in the southern Red Sea at ~5 Ma. This accompanied increases in the rate of movement on the Aqaba–Levant transform (from ~6.5 km/My to ~9 km/My) and the rate of separation between Arabia and Africa. A corresponding increase in the rate of separation across the Sheba Ridge is difficult to demonstrate in published magnetic data.
- The onset of Red Sea oceanic spreading was accompanied by a significant unconformity throughout the Red Sea basin, major development of pull-apart basins along the Aqaba–Levant transform, and increased influx of marine waters through Bab al Mandeb. Red Sea sedimentation returned to predominantly open marine conditions. The marine connection between the Gulf of Suez and the Mediterranean Sea was lost except during very high sea levels.
- Elevation differences between the Red Sea rift margins and its axial trough drove gravity sliding of large detached blocks toward the rift axis. Decollement generally occurred at the base of the Middle/Late Miocene evaporite section.
- At ~3 Ma and perhaps coincident with Red Sea spreading, oceanic spreading initiated west of 45°E in the Gulf of Aden. During the past 1 My, spreading has continued to propagate west into the Gulf of Tadjura and into Afar via the sub-aerially exposed Asal rift. The active plate boundary continues along the west side of the Danakil Block, and links to the Red Sea at the Gulf of Zula.
- The axial trough in the Gulf of Aden west of 45°E shows a pronounced WNW–ESE bathymetric fabric, rather than NE–SW trending fracture zones. This may be

caused by extensional faulting formed under NNE–SSW extension, with faults arranged en echelon because they occupy a zone of weakness controlled by sub-crustal magmatism that trends oblique to the imposed extension direction. This region is still experiencing the sub-crustal influence of the Afar plume.

- Presently, the Red Sea spreading center appears to be propagating toward the northern Red Sea to link with the Aqaba–Levant transform. North of the Zabargad fracture zone, the Red Sea axial trough is not parallel to the coastal margins, but rotated slightly counter-clockwise to be nearly perpendicular to the Aqaba–Levant transform.
- The present-day stress fields of Arabia and Africa, now separate plates, are completely decoupled. The maximum horizontal stress is oriented N–S in Arabia, whereas it is E–W in Egypt and in the Sudan west of the Red Sea.

8. Future research

The Red Sea–Gulf of Aden rift system has provided an amazing perspective of how rifting can be initiated in continental lithosphere and how this may lead to the formation of a new ocean basin. There will continue to be unlimited needs and avenues for future research in this region, spanning all disciplines of our science. We would like to highlight several key problems that can be addressed with existing geological and geophysical tools, and that will result in a greatly improved understanding of these basins:

- the three-dimensional crustal structure of the Afar triple junction;
- the detailed geometry and timing of movement within the Gulf of Aden fracture zones;
- the detailed chronologic history of the yet undated Harrats of Saudi Arabia, the basal trap 'Ashangi' series of Ethiopia, and the extensive basaltic dikes that extend from Yemen to Sinai. In particular, U–Pb studies of baddeleyite and zircon single crystals should be undertaken;
- the detailed timing and spatial variation of exhumation/uplift along the margins of the Red Sea, Gulf of Aqaba, and Gulf of Aden;
- the age of the oldest basinal syn-rift strata along the length of the rift system;
- the age of the sedimentary fill of the Gulf of Aqaba offshore pull-apart basins;
- the age and detailed correlations of the Middle to Late Miocene evaporite sequences;
- the detailed crustal structure margin-to-margin for each of the segments of the rift system;
- the cause and history of evolution of the present large-scale asymmetry of the Red Sea basin and its continental margins;
- the full extent of oceanic crust within each segment of the rift system, and the nature of the crust beneath the axis of the northern Red Sea.

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