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Development of the Arabian–Nubian Shield: perspectives on accretion and deformation in the northern East African Orogen and the assembly of Gondwana

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Abstract: The Arabian–Nubian Shield forms the suture between East and West Gondwana at the northern end of the East African Orogen (EAO). The older components of the shield include Archaean and Palaeoproterozoic continental crust, and Neoproterozoic (c. 870–670 Ma) continental-marginal and juvenile intraoceanic magmatic-arc terranes that accumulated in an oceanic environment referred to as the Mozambique Ocean. Subduction, starting c. 870 Ma, and initial arc–arc convergence and terrane suturing at c. 780 Ma, marked the beginning of ocean-basin closure and Gondwana assembly. Terrane amalgamation continued until c. 600 Ma, resulting in the juxtaposition of East and West Gondwana across the deformed rocks of the shield, and final assembly of Gondwana was achieved by c. 550 Ma following overlapping periods of basin formation, rifting, compression, strike-slip faulting, and the creation of gneiss domes in association with extension and/or thrusting. Most post-amalgamation basins contain molasse deposits, but those in the eastern Arabian Shield and Oman have marine to glaciomarine deposits, which indicate that seaways penetrated the orogen soon after orogeny. The varied character of the post-amalgamation events militate against any simple tectonic model of final Gondwana convergence at the northern end of the EAO, and requires that models accommodate alternating periods of Late Neoproterozoic extension and shortening, uplift and depression, deposition and erosion.

The Arabian–Nubian Shield (ANS) consists of Precambrian rocks exposed on either side of the Red Sea in western Arabia and northeastern Africa (Fig. 1), and comprises a collage of Neoproterozoic terranes and older crust caught up as the suture between East and West Gondwana (Fig. 2; Stern 1994; Kusky et al. 2003). The eastern part of the shield is now preserved in the Arabian Plate; the western part is a segment of the African Plate. In the 1970s, geologists interpreted the shield in terms of Wilson Cycle plate tectonics; terrane concepts became current in the mid-1980s and since the mid-1990s broader tectonic views have become common, which place the ANS at the northern end of the East African Orogen (EAO) and interpret its development in terms of the rifting of Rodinia, closure of the ensuing Mozambique Ocean, arc accretion, orogeny, extension and orogenic collapse (Stern 1994; Blasband et al. 2000; Greiling et al. 2000). The growth of the shield lasted c. 300 Ma, initiated at c. 870 Ma by the breakup of Rodinia and the deposition of oceanic volcanic arcs on thin juvenile crust of the Mozambique Ocean, and terminated at c. 550 Ma by the transformation of the northern EAO into a passive margin on the southern shore of palaeo-Tethys. West Gondwana is represented by Archaean and Palaeoproterozoic continental crust belonging to the ghost East Saharan Craton, which was strongly reworked by Neoproterozoic thermal and deformational events and is discontinuously exposed west of the Nile. However, representatives of East Gondwana at the northern end of the EAO are not yet identified, although general palaeogeographic considerations point to the eastern part of the Arabian Plate as a reasonable place to search. Models about supercontinent assembly and breakup at the northern end of the EAO depend, consequently, on information from the East Saharan Craton together with inferences drawn from the Neoproterozoic history of the ANS. This paper reviews the tectonic history of the ANS with the objective of adding to the body of data that constrains the history of East Gondwana. It discusses the cycle of accretion and transformation that created the shield, describes stages of terrane formation and amalgamation, and phases of post-amalgamation faulting, deposition and exhumation, and presents details of chronology and structure that constrain the timing and style of possible interactions between the shield and the converging blocks of East and West Gondwana. The emphasis is on the Arabian Shield north of Yemen, the Nubian Shield, and Precambrian rocks in the eastern part of the Arabian Plate in Oman. The paper includes brief comments about the character of the Yemen basement and its correlations with Precambrian rocks.

but the reader is referred to Windley et al. (1996, 2001) and Whitehouse et al. (1998, 2001b) for details on the Precambrian in Yemen.

**Terrane analysis**

Although mainly Neoproterozoic, the ANS includes Archaean and Palaeoproterozoic gneisses that have depositional and faulted contacts with Neoproterozoic rocks in the west (Schandelmeier et al. 1994b), Palaeoproterozoic plutonic rocks and gneisses that have intrusive contacts with Neoproterozoic rocks in central Arabia (Stoesser et al. 2001), and Archaean–Mesoproterozoic terranes structurally intercalated with Neoproterozoic terranes in Yemen, Somalia and southern Ethiopia (Kröner & Sass 1996; Windley et al. 1996; Tekley et al. 1998). Robust ages for the exposed rocks therefore span a long period of geological time (Fig. 3). Together with Pb, Nd and Sr isotope information (e.g. Fleck et al. 1980; Stuckless et al. 1984; Stoesser & Stacey 1988; Sultan et al. 1992; Reischmann & Kröner 1994; Stein & Goldstein 1996; Whitehouse et al. 2001a, b), the geochronological data divide the shield into crustal blocks that variously represent Neoproterozoic juvenile oceanic environments, are parts of old continental crust or, because of reworking or the mixing of old and juvenile material, are transitions between old crust and juvenile oceanic crust (Fig. 4). Differences in stratigraphy, structure (Fig. 5), potential field signatures and seismic wave velocities (e.g. Gettings et al. 1986; Georger et al. 1990; Johnson & Stewart 1995; Nehlig et al. 2002) add to these distinctions, and are criteria used in conjunction with the presence of ophiolite-decorated shear zones or sutures (e.g. Berhe 1990) to justify analysis of the shield in terms of tectonostratigraphic terranes (Jones et al. 1983). Despite this wealth of information, it must be noted that the varying detail and reliability of geological investigations, and the allochthonous and non-unique character of many of the ophiolite complexes that are used to identify sutures, mean that existing terrane analysis and correlations in the ANS are provisional (Church 1988; Shackleton 1988; Harris et al. 1990; Stern et al. 1990). Nonetheless, stemming from work in the 1980s (Johnson & Vranas 1984; Stoesser & Camp 1985; Vail 1985; Johnson et al. 1987) that built on earlier recognition of volcanic arcs and sutures (Al-Shanti & Mitchell 1976; Frisch & Al-Shanti 1977; Gass 1977; Camp 1984; Kröner 1985), it is widely accepted that the ANS is a collage of terranes.

The principal terranes are shown in Figure 6 and
are described in the following paragraphs. For the purpose of this review, information is given in some detail because the evolution of the terranes, especially with regard to the accumulation of juvenile material, governed the history of the Mozambique Ocean, and the accretion of juvenile arcs in the ANS paced the growth and consumption of oceanic lithosphere at spreading centres and subduction zones. Arc evolution can therefore be viewed as a measure of the Mozambique Ocean lithospheric budget, constraining the onset of convergence to the period when the oceanic budget started to decrease by a preponderance of subduction over spreading.

The Halfa Terrane comprises high-grade ortho- and paragneiss belonging to the East Saharan Craton, amphibolite-grade supracrustal rocks and ophiolitic mafic-ultramafic rocks (Denkler et al. 1984; Schandelmeier et al. 1994b; Stern et al. 1994). Because of pre-Neoproterozoic plutonic precursors, the gneisses yield $2.82-1.26$ Ga Nd model ages, Type III Pb ratios and strongly negative $e_{Nd}(t)$ values ($-27.0--1.3$), but also yield younger Rb–Sr ages, indicating extensive Neoproterozoic reworking (Harms et al. 1990, 1994; Stern et al. 1994). Type III Pb ratios are one of three groups of whole-rock and feldspar Pb isotopic compositions recognized in the Arabian Shield (Stoeser & Stacey 1988). Type I ratios characterize juvenile arcs (Fig. 4); Type II ratios reflect crust that is intermediate between oceanic and continental; and Type III ratios (e.g.
Fig. 3. Histogram of robust U-Pb, Pb-Pb, and Rb-Sr ages for rocks of the Arabian–Nubian Shield. (a) Entire age range. (b) Neoproterozoic subset. Data from references cited in text and from compilation by Johnson et al. (1997). Robust ages are analytically reliable and geologically meaningful; for other criteria see Johnson et al. (1997).

obtained from the Khida Subterrane that makes up the core of the Afif Terrane in the Arabian Shield, and from the Halfa and Aswan regions of the Nubian Shield) reflect a contribution of evolved Archaean and Palaeoproterozoic crustal material. Although now in fault contact, it is believed that the supracrustal rocks of the Halfa terrane were originally deposited against the East Saharan gneisses. They are discontinuous metasedimentary sequences, arc-related volcanic units and dismembered ophiolite complexes (Denkler et al. 1984; Schandelmeier et al. 1994b), and represent passive margin (Stern et al. 1994) and/or intraoceanic magmatic-arc rocks (Harms et al. 1994) deposited during the opening and closing of an oceanic basin or re-entrant above a NW-dipping subduction zone at the eastern margin of the East Saharan Craton (Schandelmeier et al. 1994b). Dating of ophiolitic gabbro, Type I grani-
Fig. 4. Pb and Nd isotope data that divide the Arabian–Nubian Shield into regions of oceanic, continental and intermediate settings. The Pb isotope classification, after Stoeser & Stacey (1988), is based on the ratios of $^{206}$Pb/$^{204}$Pb, $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb from whole-rock, feldspar and galena samples with respect to the orogen growth curve. Type III Pb plots above the curve and reflects the presence of evolved crustal material; Type I plots below the curve and reflects the presence of juvenile (oceanic and mantle) material. Data after Bokhari & Kramers (1981), Duyverman et al. (1982), Stacey & Stoeser (1983), Harris et al. (1984), Schandmeier et al. (1988), Stoeser & Stacey (1988 and refs cited therein), Harms et al. (1990, 1994), Sultan et al. (1990), Kröner et al. (1991), Agar et al. (1992), Stern & Kröner (1993), Reischman & Kröner (1994), Stern et al. (1994), Stein & Goldstein (1996), Teklay (1997), Moghazy et al. (1998), Stern & Abedelsalam (1998), Küster & Légeois (2001). Tectonostratigraphic terrane boundaries from Figure 6 are shown for reference.

Tectonic and metamorphic garnets suggest that rifting was active at c. 750Ma, subduction occurred between c. 760 and 650Ma, and basin closure, ophiolite obduction and metamorphism had taken place by c. 700Ma (Harms et al. 1994; Abdelsalam et al. 2003).

The Bayuda Terrane is underlain by amphibolite facies granitoid gneiss and paragneiss, and subordinate amounts of amphibolite, schist, quartzite and marble. The high-grade rocks are similar to the gneisses in the Halfa Terrane and may be a rifted part of the East Saharan rocks found there. However, Ries et al. (1985) documented a Neoproterozoic arc assemblage at the eastern edge of the terrane and
Küster & Liégeois (2001) report isotopic data suggesting that many of the high-grade gneisses in the Bayuda Desert within the bend of the Nile are derived from Neoproterozoic rocks. Amphibolite and epidote-biotite gneisses with a magmatic age of c. 806 Ma are interpreted as metamorphosed tholeiitic basalt and low- to medium-K dacite and rhyodacite extruded in a juvenile oceanic island-arc or back-arc basin environment. They yield positive $\varepsilon_{Nd}(t)$ values (+3.0 to +7.5) and Nd model ages (900–750 Ma) close to the magmatic age. Leucocratic gneisses and muscovite and garnet–biotite schist that yield negative $\varepsilon_{Nd}(t)$ values (−4.3 to −12.9) and Palaeoproterozoic model ages (2.43–2.01 Ga) are believed to be metasedimentary rocks derived from local volcanic and distant older
continental sources (Küster & Liégeois 2001). Late Neoproterozoic granitoids (585–562 Ma) represent reworked older crust [1.8–1.4 Ga Nd model ages; εNd(t) = −7.4 to −7.9] (Harms et al. 1990). Küster & Liégeois (2001) concluded that the rocks were probably metamorphosed under amphibolite facies conditions during a period of frontal convergence with an East Saharan ghost craton farther west, concurrent with the 700 Ma metamorphic-collisional event in the Hafsa Terrane and coincident with the emplacement of syntectonic granitoids that are now preserved as muscovite–biotite gneiss. A low-grade passive margin assemblage, the Bailateb Group (Stern et al. 1993), composed of thinly bedded distal to proximal turbidites, and subordinate quartzites, quartzitic schist, carbonates, carbonate conglomerates and breccia, and paragneiss (Meinhold 1979; Stern et al. 1993), crops out at the northern margin of the Bayuda Terrane.

The Barka Terrane (Fig. 7) consists of upper amphibolite to hornblende–granulite facies orthogneiss, amphibolite, marble, pelitic schist and orthoquartzite cut by a swarm of late E–W felsic dykes (De Souza Filho & Drury 1998). Although one of the
Fig. 7. Structural sketch map of the composite Asir Terrane and terranes in the southern Nubian Shield, plotted on a map showing north-trending, commonly serpentinite-decorated, shear zones and other spatial relations prior to Red Sea opening.
gneissic gneisses yields a single zircon Pb–Pb age of 700Ma (Teklay 1997), their general protolith ages are unknown and De Souza Filho & Drury (1998) consider the orthogneissies to be an exotic, far-travelled crustal mass. The terrane is overlain on the south by the Phanerzoic and has an ill-defined transition to the Haya Terrane in the north. The structural grain of the terrane is dominantly N–S but, to the north, swings eastward and becomes broadly conformable with the structural trend of the Haya Terrane. Nonetheless, De Souza Filho & Drury (1998) treat the Barka and Haya as separate terranes because of the pronounced differences in their metamorphic grades and overall structural styles.

An assemblage of middle-green schist facies mafic and felsic metavolcanic rocks, and mafic intrusive rocks including diorite, tonalite, granodiorite, gabbro and pyroxenite, make up the Hager–Tokar Terrane east of the Barka Terrane. The volcanic and mafic intrusive protolith date from 870 to 840Ma with partial resetting at c. 770–670Ma; syntectonic granite dates from 827Ma (Kröner et al. 1991; Teklay 1997). Rocks in the southern part of the terrane have a tholeiitic, mid-ocean ridge basalt (MORB)-type affinity and, in places, include chlorite schist, Fe–Mn chert, marble, pillowed metabasalt, metagabbro and serpentinite that are interpreted to be an accretionary prism deposited at or near a trench interspersed with popped-up subducted MORB-type ocean floor (De Souza Filho & Drury 1998; Woldehaimanot 2000). Nd isotopic data indicate moderate depletion in the mantle sources of the terrane protoliths \( [e_{Nd}(t) = +3.4–+6.7] \) and Mesoproterozoic Nd model ages suggest a contribution from old crust in the north (Kröner et al. 1991). The rocks are strongly deformed and, at a regional scale, appear to be an imbricated stack of east-vergent thrusts cut and displaced by strike-slip shear zones. The terrane is separated from the Nakfa Terrane on the east by a zone of sheared and thrusted, locally kyanite-bearing, metasedimentary rocks, referred to here as the Adobha Suture, and is separated from the Barka and Haya Terranes to the west by a major sinistral transpressional strike-slip system, the Barka Shear Zone (Suture). Assuming coast-to-coast closure of the Red Sea, the Hager–Tokar Terrane projects along-strike into the Al Lith area of the Asir Terrane (Fig. 7), which also contains kyanite-bearing metasedimentary units, a feature that forms a basis for correlation between the two regions (Kröner et al. 1991).

The Nakfa Terrane comprises variably deformed mafic and ultramafic cumulates, and diorites and granodiorites structurally over lain by greenschist facies metavolcanic and volcaniclastic metasedimentary rocks with calc-alkaline island-arc affinities (Woldehaimanot 1995, 2000; Teklay 1997; De Souza Filho & Drury, 1998). Although constrained by very limited data (a single zircon Pb–Pb age of c. 850Ma; Teklay 1997), the rocks may be contemporary with the Hager–Tokar Terrane, but are distinguished by a lesser degree of deformation and younger Nd model ages. Sources of the terrane protoliths were a moderately depleted mantle \( [e_{Nd}(t) = +3.7–+6.2] \) similar to those for the Hager–Tokar Terrane (Teklay 1997).

The Ghedem Terrane in eastern Eritrea is characterized by amphibolite facies, garnetiferous orthogneiss structurally overlain by garnet- staurolite- and kyanite-bearing gneisses and schists (Ghebreab 1999), which are tentatively correlated with granitoid gneisses and kyanite-bearing rocks in the Afaf Belt of the Asir Terrane (Fig. 7; cf. Beyth et al. 1997). Single zircon Pb–Pb dating of garnetiferous muscovite gneiss gives an age of 796Ma (Teklay 1997), compatible with gneiss in the Afaf Belt. The transition to the Nakfa terrane is a subhorizontal to moderately westward dipping, up to 3km wide, top-to-east shear zone (Ghebreab 1999). Farther north, the terrane disappears beneath the Red Sea coastal plain but in places is separated from the Nakfa Terrane by Tertiary extensional faults (De Souza Filho & Drury 1998).

The Haya Terrane consists of 870–850Ma plutonic and greenschist- to amphibolite-grade volcano-sedimentary rocks in the SE (Kröner et al. 1991; Reischmann et al. 1992), a 790Ma volcanic assemblage in the north (Abdelsalam & Stern 1993a, b; Stern & Abdelsalam 1998) and an intervening zone of intermediate-age 810Ma plutonic rocks (Schandelmeier et al. 1994a). The rocks in the SE comprise juvenile intraoceanic island-arc rocks and syntectonic, arc-related, calc-alkaline Type I plutons, characterized by low initial Sr ratios, moderately depleted \( e_{Nd}(t) \) values (c. +6.0) and \( T_{DM} \) model ages approaching the formation ages (1.05–1.04Ga) (Embleton et al. 1983; Klemencic 1985; Klemencic & Poole 1988; Kröner et al. 1991; Reischmann et al. 1992). The 790Ma volcanic assemblage is a tholeiitic to calc-alkaline, within-plate and/or MORB bimodal volcanic succession, interpreted to result from crustal extension and rifting at the northern margin of the Haya Terrane (Abdelsalam & Stern 1993a). The intermediate plutonic rocks are diorite and granodiorite, geochemically resembling granitoids found in modern convergent plate margins (Schandelmeier et al. 1994a; Stern & Abdelsalam 1998). The spatial relations of these constituent rock units suggest that the Haya Terrane was formed by a northward-migrating, SE-dipping subduction zone, accompanied with some rifting at the northern margin.

In early terrane models (Kröner et al. 1987), the Gebeit and Gabgaba regions were interpreted as separate terranes, but recent work shows that the intervening Hamisana structure is a post-amalgamation zone of E–W shortening and not a suture.
(Abdelsalam & Stern 1996; de Wall et al. 2001; Miller & Dixon 1992), and the Gebeit and Gabgaba rocks are treated here as a single terrane. Terrane-forming rocks are relatively sparse because younger granitoids dominate the region. Terrane protoliths in the NE comprise a c. 832 Ma assemblage of subalkaline, calc-alkaline and tholeiitic, mostly subduction-related, volcanic rocks, although some rocks with high Ti/V ratios (> 20) may reflect an incipient back-arc basin in an immature, intraoceanic island-arc environment (Gaskell 1985; Reischmann & Kröner 1994). The volcanic rocks are broadly contemporary with tonalite along the Hamisana Shear Zone (Stern & Kröner 1993), and the volcanic and plutonic rocks may represent a c. 830–810 Ma arc system. Low initial Sr ratios and Nd model ages similar to crystallization ages are strong evidence of a juvenile oceanic setting, and the strongly positive εNd(t) values (+ 6.1–+ 8.4) (Fig. 4) indicate that mantle sources were more depleted than in adjacent parts of the shield (Reischmann & Kröner 1994). The northeastern part of the terrane includes immature volcanic-arc rocks (Abdeen et al. 2000), yielding Rb–Sr whole-rock isochron and zircon-evaporation ages of 735–697 Ma (Stern & Kröner 1993), and a succession of carbonate-rich, amphibolite facies marble, quartzite and amphibolite close to and along the Keraf Suture Zone, forming an extension of the passive margin sequence found in the Bayuda Terrane (Stern et al. 1989; Abdelsalam et al. 1998). Terrane protoliths in the SE include an assemblage of c. 720 Ma basalt, basaltic–andesite, andesite, dacite, rhyolite, and interbeds of agglomerate and tuff intruded by coeval granite and mafic dykes (Klemenic 1985; Klemenic & Poole 1988).

Terrane analysis in the Eastern Desert [alternatively named the Gerf Terrane (Kröner et al. 1987), the Southeastern Desert Terrane (Abdeen et al. 2000) or the Aswan Terrane (Greiling et al. 1994)] is problematic because of disruption of early structures by later thrusting and intrusions, uncertainties about the number of subduction zones in the region, and the locations of root zones of the ophiolite nappes and mélanges that constitute a dominant feature of the area (Greiling et al. 1994; Shackleton 1994). Stratigraphically, the region is underlain by gneiss complexes interpreted by some workers to be prePan-African basement (e.g. El-Gaby & Hashad 1990). However, because they do not yield radiogenic Pb, elevated initial Sr ratios, negative εNd(t) values, or old Nd model ages (Stern et al. 1989; Sultan et al. 1992), recent compilations treat the gneisses as the metamorphic equivalents (Tier 1 infracrustal rocks) of juvenile Neoproterozoic protoliths that were probably similar to the overlying greenschist facies volcanosedimentary rocks and ophiolite nappes and mélangé (Tier 2 supracrustals) (Bennet & Mosley 1987; Greiling et al. 1988, 1994).

Exceptional negative εNd(t) values, old Nd model ages and inherited zircons in certain post-tectonic granites (Fig. 4); (Sultan et al. 1990; Hassanen & Harraz 1996) indicate local involvement of pre-PanAfrican crustal components. In the SW, the terrane includes strongly folded ortho- and paragneisses, low-grade oceanic island-arc rocks and imbricated serpentinite, tect-carbonate schist and metagabbro ophiolite nappes, together with a shelf assemblage of marble and conglomerate (Greiling et al. 1994; Abdeen et al. 2000). Adjacent to the Onib–Sol Hamed Suture, the terrane-forming rocks are steeply dipping sheared gabbro, cumulate ultramafic rock (mainly pyroxenite), serpentinite, sheeted dykes and pillow lavas belonging to a disrupted SE-facing ophiolite succession (Fitches et al. 1983; Kröner et al. 1987). An accretionary prism to island-arc volcanic–plutonic sequence is recognized at the southeastern end of the Hafafit culmination (Greiling et al. 1994). In the central parts of the region, the terrane protoliths are rootless ophiolite nappes, the largest of which make up the Gerf and Barramiya Nappes. Zircon evaporation and Sm–Nd whole-rock dating of layered gabbro and plagiogranite from these ophiolites weakly constrains ocean–crust formation to between c. 810 and 720 Ma (Kröner et al. 1992b; Stern & Kröner 1993), compatible with an Rb–Sr isochron of 768 ± 31 Ma obtained from felsic volcanic rocks in the Abu Swayel area (Stern & Hedge 1985). Intrusions in the Allaqi–Heiari–Onib–Sol Hamed Suture Zone constrain basin closure as prior to c. 735–720 Ma (Kröner et al. 1992b; Stern & Kröner 1993).

The Asir Terrane consists of juvenile oceanic rocks characterized by low initial Sr ratios and Type I Pb ratios, and εNd(t) values of + 7.5 and + 8.9 (Fig. 4), suggesting strong mantle depletion (Bokhari & Kramers 1981). The oldest rocks (c. 850–790 Ma) crop out in the west and are mainly assemblages of convergent-margin volcanic and plutonic rocks (Reischmann et al. 1984; Kröner et al. 1992a), such as the large An Nimas Arc (Fig. 7); (Stoeser & Stacey 1988). Younger volcanic and plutonic rocks form the Tarib Arc (785–720 Ma; Stoeser & Stacey 1988), and smaller areas of ocean spreading and convergent-margin assemblages (Bookstrom pers. comm.) underlie the eastern part of the terrane. Possible oceanic-plateau volcanic rocks (Reischmann et al. 1984) and kyanite–staurolite-bearing metasedimentary units in the NW of the terrane are candidates for correlation with the Hager–Tokar Terrane. The terrane is composite, created by the amalgamation of several detachment along one or more of the serpentinite–decorated shear zones that dominate the region (Fig. 7). Because of intense deformation and the reconnaissance nature of existing mapping, neither subduction polarities nor convergence trajectories of these terranes are known.
The Jiddah Terrane is dominated by northeasterly structural trends, similar to those in the Haya Terrane, and, although few robust ages are reported, is estimated to have a similar time span (c. 870–760 Ma). Structural and intrusive relations suggest that the oldest rocks, composed of diorite, tonalite and granite, are in the south. Resembling the 870–850 Ma plutonic rocks in the south of the Haya Terrane, they have tholeiitic to calc-alkaline affinities and mantle-type initial Sr ratios compatible with an active-arc setting (Fleck 1985; Moore & Al-Rehaili 1989). Younger, c. 815–810 Ma, calc-alkaline plutonic rocks (Calvez & Kemp 1982; Stoezer & Stacey 1988) in the north and NE correlate with plutonic rocks in the north of the Haya Terrane. A parautochthonous fold-thrust belt, comprising volcanosedimentary rocks and syntectonic granite gneiss, crops out along the northwestern margin of the terrane (Ramsay 1986; Johnsson 1998), and an autochthonous unit of immature volcanic-arc rocks (Maoh Group; <810–c. 780 Ma) intruded by 770–765 Ma granophyre, and 760 Ma tonalite and granodiorite occurs in the NE (Calvez & Kemp 1982; Afifi 1989). As in the Haya Terrane, the apparent temporal and spatial relations of the rocks suggest that the terrane was created by a northward-migrating SE-dipping subduction zone.

The Hijaz Terrane comprises volcanic arcs and younger volcanic sedimentary successions that may also have formed above a southerly dipping subduction system (Stoezer & Camp 1985). An older arc (<870–807 Ma) crops out in the southern part of the terrane, containing bimodal, low-K to tholeiitic greenschist facies pillow basalt, ryholite, tuffs and volcaniclastic sedimentary rocks deposited in an oceanic environment (Camp 1986). A younger arc (c. 750–710 Ma) is present in the northern part of the terrane as a sequence of basalt, ryholite and large amounts of volcaniclastic sedimentary rocks that locally may have accumulated in extensional basins above older arc complexes (Kemp 1981). Mafic-ultramafic ophiolite complexes (Bi’r Umq, Jabal Tharwah) that originated between c. 870 and 830 Ma in mixed mid-ocean-ridge and island-arc environments (Pallister et al. 1988) are thrust over the Jiddah Terrane at its southern margin (Al-Rehaili & Warden 1980; Nassief et al. 1984).

As in the Eastern Desert, the structure and distribution of protoliths in the Midyan Terrane are significantly disrupted by 725–696 Ma granitoid intrusions and post-terranne amalgamation deformation, but available information indicates that they include several assemblages of subduction-related volcanic–volcaniclastic and calc-alkaline intrusive rocks weakly constrained between >725 and >710 Ma (Hedge 1984), and ophiolite complexes dating from 780 to 740 Ma (Claesson et al. 1984; Pallister et al. 1988). In the north and west, the volcanic–volcaniclastic sedimentary rocks uniquely include a banded-iron formation, a rock type in the ANS that is confined to the Midyan and Eastern Desert Terranes (Sims & James 1984), and is regarded by the present authors as a strong correlation feature. The geochemical signature of some of the volcanic rocks indicates intraoceanic subduction environments. The ophiolitic units (Jabal Ess (Shanti 1983) and

![Fig. 8. Afif Terrane assembly showing estimated ages of amalgamation of subterraneus, and of overlap assemblages and stitching granites that constrain the minimum age of assembly.](image-url)
Jabal Wask (Bakor et al. 1976]) include discontinuous lenses and thrust sheets of serpentinite, peridotite, cumulus and non-cumulus gabbro, dyke complexes and pillow basalt. They originated in fore-arc to back-arc environments (Pallister et al. 1988) and were structurally emplaced in a subvertical shear zone at the contact between the Midyan and Hijaz Terranes, which marks the Yanbu Suture.

The Afif Terrane is a composite tectonostratigraphic unit in the northeastern part of the Arabian Shield assembled from four possible subterranes (Fig. 8; Johnson & Kattan 2001). The Khida Subterran, in the SE, includes Palaeoproterozoic biotite alkali-feldspar granite, meta-anorthosite and biotite granitic orthogneiss (c. 1860–1660Ma; Stoeser et al. 2001; Whitehouse et al. 2001a) and c. 800Ma almandine–sillimanite amphibolite facies metavolcanic rocks, paraschist, paragneiss and orthogneiss (Stoeser pers. comm.) characterized by strongly negative εNd(t) values and old Nd model ages. The Siham Subterran, in the SW, consists of 750–695Ma greenschist facies volcanosedimentary rocks, gabbro, diorite, tonalite, granodiorite and monzogranite, and is inferred to be a continental-margin arc developed above a subduction zone inclined east beneath the Khida crust (Agar 1985). The Nuqrah Subterran is a cryptic 840–820Ma tectonostratigraphic unit in the NW composed of lower greenschist to amphibolite facies island-arc-type volcanic and sedimentary rocks (Nuqrah Formation) (Delfour 1977), mafic to intermediate plutonic rocks, and tectonically disrupted ophiolite complexes (Calvez et al. 1983; Le Metour et al. 1983; Quick 1991; Pallister et al. 1988). The Suwaj Subterran, in the east, is a 745–667Ma assemblage of weakly metamorphosed but strongly cataclasized diorite, quartz diorite, tonalite, sodic granodiorite, and subordinate basalt and dacite (Cole & Hedge 1986).

The Ad Dawadimi Terrane constitutes one of the most homogeneous crustal units in the Arabian Shield and can be traced by its distinctive, subdued, aeromagnetic signature some 300km north beneath the Panerzoic (Johnson & Stewart 1995). It is commonly modelled as a thin-skinned allochthon bounded and internally sliced by listric thrusts (e.g. Al-Shanti & Mitchell 1976). Terrane protoliths include thinly layered sericite–chlorite phyllite and schist derived from fine-grained sandstone and siltstone (Delfour 1979) that were deposited in a possible accretionary prism, and c. 695Ma ocean-floor mafic-ultramafic rocks (Stacey et al. 1984) exposed as linear belts along the margins of, and within, the terrane. The terrane was thrust at c. 680Ma over the adjacent Afif Terrane (Al-Saleh et al. 1998). The tectonic setting of the terrane is uncertain and it is debated whether the oceanic-floor material represents back-arc or fore-arc oceanic crust and whether the metasedimentary rocks are related to a west- or east-dipping subduction zone (Al-Shanti & Mitchell 1976; Al-Shanti & Gass 1983; Camp et al. 1984; Quick 1991; Johnson & Stewart 1995; Al-Saleh et al. 1998; Al-Saleh & Boyle 2001a). Westward obduction of the ocean crust marked the closure of the Ad Dawadimi Basin at c. 680Ma and suggests that the basin had a relatively short life of c. 15Ma.

The Ar Rayn Terrane crops out at the eastern margin of the shield, and extends north and east beneath the Panerzoic (Johnson & Stewart 1995). It is bounded to the west by the Al Amaar Fault and to the east by a magnetically defined contact with a possible continental block in the concealed basement. The terrane includes a bimodal volcanic–volcanosedimentary assemblage transitional between the tholeiitic and calc-alkaline series (Vaslet et al. 1983) and large mafic to intermediate calc-alkaline intrusions. The rocks may have formed in a mature island arc or more likely, as suggested by their metalogographic signature, in a continental-margin environment (Doebrich et al. 2001) above an east-dipping subduction zone. The layered rocks are not directly dated but were evidently deposited prior to the emplacement of syntectonic granitoids at 667–640Ma (Stacey et al. 1984).

East of the Arabian Shield, Precambrian rocks dip beneath Panerzoic strata. They descend to structural lows as much as 12km below sea level in central Arabia (Konert et al. 2001), but rise in structural highs to within a few kilometres of the surface farther east and crop out in Oman (Fig. 9). Referred to in the literature as 'basement', crystalline Precambrian rocks in Oman consist of dolerite, granodiorite, granite, migmatite, and greenschist and amphibolite facies metasedimentary rocks and gneiss unconformable below the mainly sedimentary Late Neoproterozoic Huqf Supergroup (Al-Doukhi & Divi 2001; Allen & Hildebrand 2001; Al-Kathiri 2001; Cozzi et al. 2001; Mercoll et al. 2001). The rocks have c. 1000–730Ma crystallization ages (Gass et al. 1990; Pallister et al. 1990; Würsten et al. 1991; Mercoll et al. 2001) and are evidently Neoproterozoic, but their exact tectonic provenance is unknown. Pallister et al. (1990) reported Pb isotope data suggesting that the region contains evolved crustal material and Johnson & Stewart (1995) speculatively inferred that much of eastern Arabia is underlain by continental crust. However, the extent of continental crust, if truly present, is not established. Nor is it established whether such crust was part of a plate east of the juvenile rocks of the ANS (perhaps as a candidate for a fragment of East Gondwana) or was a detached continental terrane intercalated with Pan-African terranes similar to the terrane assemblage in Yemen (Windley et al. 1996; Whitehouse et al. 2001b). It is likewise unknown whether the plutonic rocks
are related to a west- or east-facing (Al-Shanti & Mitchell 1983; Camp et al. 1984; Stewart 1995; Al-Saleh et al. 2001a). Westward it marked the closure of c. 680Ma and suggests a short life of c. 15Ma. crops out at the eastern extends north and east in the concealed base- a bimodal volcanic–volcanic transitional between-rises (Vaslet et al. intermediate calc-alkaline rocks formed in a mature continental-margin, L. 2001) above an east-ern layered rocks are not emplaced prior to tectonic granitoids at 984).

Precambrian rocks dip and descend to structurally lower sea level in central Arabian shield, but rise in structural metres of the surface Oman (Fig. 9). Referred to as the basement, crystalline basement consist of dolerite, greenstone, and gneissic and greenish-gray sedimentary rocks and the main sedimentary succession, the Huqf Supergroup (Al- & Hildebrand 2001; et al. 2001; Mercollis et al. 2000–2003; crystallized, Pallister et al. 1990; Illi et al. 2001) and are just the exact tectonic Pallister et al. (1990) suggesting that the region is orogenic and Johnson & y inferred that much of p by continental crustal, nent, crust, if true. Nor is it established t of a plate east of the as a candidate for Shana) or was a detached with Pan-African e assemblage in Yemen house et al. 2001b). It is the plutonic rocks sampled for age dating in Oman are Neoproterozoic intrusions into older continental crust or denote an entirely Neoproterozoic basement.

Terrane amalgamation and suture zones

The contacts between terranes in the ANS include (1) sutures composed of ophiolite-decorated shear zones; (2) cryptic or unimpressive shear zones that may be original sutures, faults superimposed on original sutures or post-suturing structures; and (3) post-amalgamation fault zones that may be unrelated to original suturing events. The sections below describe shear zones believed to relate to convergence and amalgamation among the ANS terranes. They are treated as sutures or modified sutures and record an episode of terrane amalgamation. Structural information is sufficiently detailed so as to indicate the sense of convergence at some of the suture zones, which range from frontal (involving large amounts of thrusting) to transpressional (involving large amounts of strike-slip coupled with thrusting), and provides some constraints on the possible direction of convergence of the larger, flanking blocks of the east and west of the Arabian Shield.

Amalgamation events at 786–760 Ma

The oldest structures believed to reflect convergence and amalgamation in the ANS include the Tabalah-Tarj Shear Zone and the Afaf Belt of syntectonic gneiss in the central part of the Asir Terrane, and similar rocks in Eritrea. The Tabalah-Tarj Shear Zone (Figs 5 & 10) comprises phyllonite, schist and gneiss resulting from ductile mylonitic deformation and dextral shear, prior to 755Ma (Johnson et al. 2001). Syntectonic gneiss dated at 778–763Ma (Cooper et al. 1979) suggest a contemporary deformation and
Fig. 10. Terrane assembly in the Arabian–Nubian Shield, showing inferred ages and trajectories of amalgamation.
magmatic event west and SW of the Tabalah-Tarj Shear Zone. The syntectonic gneiss is one of several gneiss antiforms that make up the Arafat Gneiss Belt in the western Asir Terrane (Fig. 7). The Arafat Belt possibly extends into the Ghedum Terrane of Eritrea, where one of the gneisses is dated at 796 Ma (Teklay 1997), and the combined structure of gneisses and shear zones is inferred to mark a period of early amalgamation in the southern Arabian and Nubian Terranes.

Contemporary accretion events to the south and north are suggested by Pb loss at c. 760 Ma from zircon grains in the Al-Mahfid Gneisses in Yemen (Whitehouse et al. 1998, 2001b), and by pre-, syn- and post-tectonic rock units along the Bi’r Umq-Nakasib Suture between the Jiddah-Haya and Hijaz-Gebeit Terranes. The Bi’r Umq-Nakasib Suture is an ophiolite-decorated fold-shear zone, 5–65 km wide and >600 km long, which is characterized by a commonality of structure and gold and base-metal deposits along its length, and constitutes one of the best-documented sutures in the ANS, separating regions of contrasting stratigraphy and Nd isotope signatures (Stoesser & Camp 1985; Vail 1985; Camp 1984; Ramsay 1986; Abdelsalam & Stern 1993a; Stern & Kröner 1993; Johnson 1994; Johnson et al. 2002). The ophiolites are allochthons of mid-oceanic to supra-subduction oceanic crust (Nassief et al. 1984; Abdel Rahman 1993; Schandelmeier et al. 1994a) that formed at c. 870–830 Ma (Pallister et al. 1988) and, together with intercalated volcanosedimentary rocks, were deformed during a period of dextral transpression involving orthogonal convergence, thrusting and non-coaxial dextral shear (Abdelsalam & Stern 1993a; Wipfler 1996; Johnson 1998; Johnson et al. 2002). As constrained by the ages of terrane protoliths, suturing began some time after 810–780 Ma, was active between 780 and 760 Ma, the ages of putative syntectonic intrusions, and was complete by c. 760–750 Ma, the ages of stitching plutons and ophiolite obduction (Pallister et al. 1988; Schandelmeier et al. 1994a; Stern & Abdelsalam 1998).

Amalgamation events at 750–660 Ma

Convergence in the vicinity of the East Saharan Craton began soon after creation of the Bi’r Umq-Nakasib Structure. It was driven by N–S shortening and closure of the ocean basin in the southeastern part of the Halfa Terrane at c. 700–650 Ma (Abdelsalam et al. 2003), and caused frontal collision between the Halfa and Bayuda Terranes. The resulting Atmûr-Delgo Suture is marked by a chain of ophiolite nappes thrust SE over the Bayuda Terrane and locally back thrust over the Halfa vol-
canic supracrustal rocks (Harms et al. 1994; Schandelmeier et al. 1994b). Collision and suturing, weakly constrained between 720 and 700 Ma, was associated with progressive deformation, migmatization of granitoids and upper greenschist to amphibolite facies metamorphism, which reached a peak at c. 702 Ma (Denkler et al. 1984; Harms et al. 1994; Stern et al. 1994; Abdelsalam et al. 1998). Küster & Liégeois (2001) propose concurrent accretion between the Bayuda and the East Saharan Terranes farther west. Continued downwarping, after terrane convergence, of the (detached?) oceanic crust subducted along the suture zone is suggested by the emplacement of c. 760 Ma–650 Ma Type I calc-alkaline granitoids (Harms et al. 1994), following which there was a transition to intraplate rifting and volcanism (Stern et al. 1994).

Concurrent terrane amalgamation farther north created the Allaqi–Heiani–Sol Hamed–Onib–Yanbu Suture, which comprises nappes and slices of dismembered ophiolite in a sinuous but broadly east-trending shear zone between the Gebeit-Hijaz and Eastern Desert–Midyan Terranes. Continuity along this zone is generally accepted (e.g. Kröner et al. 1987; Stern et al. 1990; de Wall et al. 2001), although Shackleton (1996) argued that the Allaqi Ophiolites are obducted klippens with roots north of, rather than along, the Allaqi–Heiani Zone and Greiling et al. (1994) suggested that the Allaqi–Heiani–Sol Hamed–Onib sector of the suture may continue in a postulated South Hafait Suture, rather than the Yanbu Suture (Fig. 10). The presence of a metamorphic-sole complex in the Wadi Haimur–Abu Swayel area is evidence, however, that the Allaqi Nappes are close to their origin, i.e. not far travelled (Abd El-Nabî & Frisch 1999). The Allaqi–Heiani segment of the suture zone contains gneiss and discontinuous thrust duplexes of ophiolitic, metamorphosed island-arc rocks and mylonite linked by the Allaqi Shear Zone (Taylor et al. 1993; Greiling et al. 1994). Metamorphism is mainly in the greenschist facies, although higher grades are reported locally, and high-pressure/low-temperature blueschist assemblages occur in the footwall of the Allaqi Structure (Taylor et al. 1993). Regionally, the structure dips east to NE, but the nappes are folded and, in detail, dip south and north where the suture zone has an easterly trend, and east and west where the suture swings to the north (Greiling et al. 1994). The Allaqi Shear is a steeply dipping sinistral strike-slip shear with gently NW- and SE-plunging stretching lineations in east-trending sectors, and a reverse fault in more northerly trending sectors. These structural and kinematic changes imply a combination of orthogonal shortening and strike-slip and, although it is not clear whether the Allaqi Shear Zone itself is the original suture or a later structure that modified the suture, are compatible with terrane convergence.
and top-to-the-NW transport during sinistral NW-directed transpression (Taylor et al. 1993; Greiling et al. 1994). The Sol Hamed segment of the suture comprises a subvertical south-facing ophiolite. The ophiolite has a ductile flattening fabric that may reflect pre-obduction suboceanic deformation, and was tilted to the vertical and deformed by NE–SW shears prior to folding about steeply plunging axes and shearing in association with the development of steeply plunging stretching lineations (Fitches et al. 1983). The ophiolite was locally thrust SE over younger volcanic rocks and restructured as a tectonic mélangé. The Yanbu segment of the suture is a subvertical to steeply dipping shear zone containing nappes and fault-bounded lenses of mafic and ultramafic rocks (Bakor et al. 1976; Shanti 1983; Pallister et al. 1988). The segment is discontinuously exposed because of late- to post-amalgamation granitoids and Cenozoic flood basalt, and is deformed by left-lateral strike-slip faults belonging to the later Najd system and by folding around the ESE-plunging nose of a gneiss antiform. The trajectory of suturing has not been established. Stoeser & Camp (1985) suggested a southerly vergence but abundant S–C fabrics indicate a significant component of dextral strike-slip. The timing of convergence along the suture is not well constrained, but amalgamation probably occurred some time between c. 700 and 600 Ma, following 808–721 Ma ophiolite formation, and preceding emplacement of 730–690 Ma granodiorite and tonalite (Fitches et al. 1983; Claesson et al. 1984; Kröner 1985; Stern & Hedge 1985; Pallister et al. 1988; Kröner et al. 1992b; Stern & Kröner 1993). This time period is somewhat old, however, for K–Ar ages of c. 600 and 585 Ma that represent cooling of hornblende and biotite following peak metamorphism in paragneiss 20–50 km north of the suture zone in the Allaqi area (Abd El-Naby & Frisch 2002), and robust dating of the suturing event is needed.

**Amalgamation events at 680–640 Ma**

Younger suturing events in the ANS reflect the convergence of oceanic terranes and mixed continental–oceanic terranes in the central and eastern parts of the Arabian Shield. The *Hulayfah–Ad Dafinah–Ruwah Suture* between the Afīf Terrane and oceanic terranes to the south and west is a subvertical shear zone 5–30 km wide and 600 km long created during an episode of sinistral transpression (Johnson & Kattan 2001). Putative syntectonic orthogneiss dated at 683 ± 9 Ma (Stacey & Agar 1985) and overlap-assemblage rhyolite dated c. 650 Ma (Doebrich pers. comm.) suggest that convergence was in progress by 683 Ma and completed by 650–630 Ma. Later faulting and thrusting modified the suture, particularly along its SE segment, and Najd faults offset the suture along its northern segments. The *Halaban Suture* between the Afīf and Ad Dawadimi Terranes (Urd Suture of Pallister et al. 1988) is a nappe of Halaban ophiolite thrust westwards over the Suwaj Subterrane at c. 680 Ma (Al-Saleh et al. 1998). Aeromagnetic data suggest that the suture originally extended north of the shield, as shown in Figure 10 (Johnson & Stewart 1995), but it has been extensively modified by post-amalgamation Najd faulting. The *Ar Amar Suture* between the Ad Dawadimi and Ar Rayn Terranes is a high-angle fault zone (the Al Amar Fault) containing narrow lenses of ophiolites and carbonate-altered ultramafic rock (listwaenite) (Al-Shanti & Mitchell 1976; Nawab 1978), which is also inferred to extend north of the shield. Syn- and post-tectonic plutons (670–640 Ma) in the Ar Rayn Terrane (Stacey et al. 1984) weakly constrain the timing of convergence; however, the trajectory of convergence is unknown, although limited observations on shear fabrics suggest that the convergence included a component of dextral horizontal shear. The *Nabitah Fault Zone* is a north-trending, serpentinite-decorated shear zone in the eastern part of the Asir Terrane (Figs 7 & 10). It is one of the classic suture zones described in the Arabian Shield literature, giving its name to the 680–640 Ma Nabitah Orogeny and to the Nabitah Orogenic (or mobile) Belt, which is mapped as a zone of deformation and magmatism extending north across the entire shield (Stoeser & Camp 1985; Stoeser & Stacey 1988). The tectonic significance of the Nabitah Fault Zone is, however, ambiguous. It may be a suture in the northern part of the Asir Terrane, south of the Ruwah Fault Zone, where it separates greenschist facies, mainly volcanosedimentary, rocks to the west from amphibolite facies paragneiss and orthogneiss to the east, but farther south it extends as a ductile shear zone through the middle of the Tarib Batholith. Bodies of syntectonic gneiss on either side of and along the fault zone are conspicuous features of the structure, and together with late tectonic granites constrain fault movement to between c. 680 and <640 Ma (Stoeser et al. 1984; Stoeser & Stacey 1988; Johnson et al. 2001). Shear fabrics in gneiss, granite and volcanosedimentary rocks, as well as offsets of passive markers, indicate that shearing was dextral during both early ductile and later brittle phases of deformation.

**Amalgamation events at 650–600 Ma**

The youngest arc–arc amalgamation event in the ANS is represented by the *Kerf Suture*, a north-trending ophiolite-decorated belt of folded and sheared rocks at the contact between the Bayuda-Halfa and Gebeit–Gabgaba Terranes (Fig. 10;
Abdelsalam et al. 1998). Originally identified by reconnaissance ground surveys and shuttle imaging data (Almond & Ahmed 1987; Abdelsalam et al. 1995), the belt of deformed rocks was proposed to mark the arc–continent suture between Neoproterozoic rocks of the Nubian Shield and the East Saharan Craton (Stern 1994). However, if the Bayuda Terrane is largely Neoproterozoic, as inferred by Küster & Liégeois (2001) on the basis of its isotopic characteristics, the cratonic margin lies farther west. Together with the Nakasib–Bi’r Umq, Hulayfah–Ad Dafinah–Ruwah and Halaban suture zones, the Kerif Zone is one of the more clearly evidenced sutures in the ANS, extending for considerable length, separating regions of contrasted crustal compositions and metamorphic grade [amphibolite–granulite facies rocks yielding $e_{Nd}(t)$ values of $-12.9 \pm 7.5$ to the west and greenschist facies rocks yielding $e_{Nd}(t)$ values of $+6.1 \pm 8.4$ to the east; Fig. 4], and coinciding, at least in the north, with slope facies sedimentary rocks that denote a strong change in depositional environments at the time of deposition of the terrane-forming rocks. The suture zone truncates the Atmur–Delgo Suture and fold-thrust belts in the Bayuda Terrane. Whether it truncates and/or converges with the Hammama Zone in the Gebiet–Gabgaba Terrane and the Nakasib Suture is unresolved because of sand cover. Suturing was caused by c. 650–600Ma E–W shortening and NW–SE oblique collision (Abdelsalam et al. 1998, 2003), concurrent with isotopic rehomogenization and cooling and/or uplift of high-grade basement in the Bayuda and Halfa Terranes, which was coincident with the closing stage of metamorphism along the Atmur–Delgo Suture, the emplacement of high-level granitoid plutons and a period of wrench faulting that correlates with the Najd Fault System in the Arabian Shield (Bernau et al. 1987; Harms et al. 1994). Its cessation at c. 580Ma was marked by the rapid uplift and cooling of a deformed granite (Abdelsalam et al. 1998).

**Amalgamation in Yemen**

As extensively documented by Windley et al. (1996) and Whitehouse et al. (1998, 2001b), the Pre cambrian basement of Yemen is a collage of Archaean gneissic and Pan-African island-arc terranes. The boundaries between the terranes are prominent shear zones marked by structural dislocations, mylonite and, at Hajjah and NE of Sada (Fig 11), ophiolite. Correlation across the Gulf of Aden between the terrane collage in Yemen and equivalent rocks in the Horn of Africa is reasonably well established (Kröner & Sassi 1996; Whitehouse et al. 2001b; Windley et al. 2001), but correlation to the north with terranes in Saudi Arabia is less certain. Windley et al. (1996) correlate the ophiolite-decorated Shear Zone at Hajjah and Sada with the Nabiath Fault Zone, but much of the intervening ground is covered by Phanerozoic sediments and flood basalt so that the true extension of the Hajjah–Sada Shear Zone is problematic and, as shown in Figure 10, is more likely to be in the region east of Najran rather than along the Nabiath Fault Zone, unless there is a hidden intervening Najd-type fault. As noted above, the Asir Terrane is characterized by north-trending shear zones, numbers of which are decorated by slivers of mafic–ultramafic rocks (Fig. 11). In this context, the Hajjah–Sada Shear Zone is not extraordinary and is viewed by the present authors as a typical structural element of what may be a southern continuation of the Asir Terrane into Yemen. The northward extent of the Archaean Abas and Al Mahfíd Terranes are concealed by Phanerozoic sedimentary rocks and eolian sand of the Rub al Khali Basin, and are likely limited
Fig. 11. Map of the southern part of the Arabian Shield showing relations between the Precambrian exposures in Saudi Arabia and Yemen. A, Asharah Fault Zone; B, Baydah Shear Zone; I, Ibran Shear Zone; J, Junaynah Fault Zone; N, Nabitah Fault Zone; UM, Umm Farwah Shear Zone; TB, Tabalah Shear Zone; TJ, Tarj Shear Zone.

by NW-trending Najd faults (Fig. 11). Proposed correlation of these terranes with the Khida Subterrane of the Afif Terrane is therefore problematic, not least because of isotopic and geochronological differences between the crustal units, as pointed out by Whitehouse et al. (2001b). Given the available data, it is clear that the Abas, Al-Mahfid and Khida Terranes are continental in character. They are conceivable fragments rifted from an older continent, but strict correlation of the terranes, in the sense that they are pieces of the same crust, is not currently supported.

Post-amalgamation overlap assemblages

The 150–100 Ma span between the 680–640 Ma amalgamation event in the eastern Arabian Shield and the transformation of the entire northern EAO into a passive margin on the southern flank of palaeo-Tethys is one of noteworthy tectonic heterogeneity involving diachronous deformational and crust-forming events that, from the Gondwana perspective, provide a large amount of information about the closing stages of Gondwana convergence and assembly. One expression of this heterogeneity was the development of post-amalgamation volcano-sedimentary basins (Fig. 12; Johnson 2003). About 40 such basins are known, ranging in size from large basins in Oman and the Arabian Gulf (Figs. 1 & 9) (underlying an area >200 × >600 km²) to small basins in the Midyan Terrane (5 × 15 km²), and varying in age, based on the onset of deposition, from c. 723 to 580 Ma (Table 1). Using the criteria of more than/less than 500 m thickness of carbonate succession, the relative abundances of grey-green and red-purple rocks, and other sedimentary characteristics, the overlap assemblages are provisionally divided into marine and epicontinental basins.

Marine basins

The largest marine post-amalgamation basins occur in Oman and western Saudi Arabia. The Omani Assemblage, the Huqf Supergroup (c. 732–540 Ma: Brasier et al. 2000), comprises epiclastic, carbonate, subordinate volcanic rocks and thick successions of salt that crop out along the Arabian Sea coast and are intersected in drillholes or imaged on seismic pro-
files beneath Phanerozoic cover inland (Loosveld et al. 1996; Blood 2001). The supergroup is unconformable on the older crystalline basement. The upper and lower parts accumulated in fault-controlled NE-trending basins, whereas the middle part was deposited in a platform-and-slope environment (Husseini & Husseini 1990; Loosveld et al. 1996; Allen & Hildebrand 2001; Leather et al. 2001). Glaciomarine deposits in the basal 1100 m, including diamictites and dropstone-bearing laminated mudstone, are associated with cap dolomites and deepwater siltstone and argillaceous sandstone, and have Sr isotopic excursions that appear to correlate with the Surtian and Marinoan glaciations (Brasier et al. 2000; Amthor et al. 2001).

The Murdama group (<670–650 Ma; Cole & Hedge 1986) is a shallow-marine succession of deformed, but virtually unmetamorphosed, grey-green feldspathic sandstone and siltstone, polymict conglomerate, carbonate, subordinate basalt and rhyolite, and carbonate build-ups >1000 m thick that crops out in basins in the eastern Arabian Shield (Fig. 12; Johnson 2003). Conglomerate is typically massive and probably originated as fanglomerate, but the sandstone and siltstone are well bedded and represent tidal-flat to turbiditic deposits (Wallace...
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The carbonates are locally stromatolitic. The basins may have been locally controlled by faults, but on the regional scale appear to be large down sags. Smaller marine basins are represented by the Fatima and Bani Ghayy groups, weakly constrained at 688–675 Ma (Duyverman et al. 1982; Darbyshire et al. 1983) and c. 650 (Doebrich pers. comm.), respectively. Both groups contain significant amounts of terrestrial conglomerate and sandstone, and bimodal volcanic rocks consistent with their deposition in fault-controlled basins (Agar 1986), but also contain up to 750 and 800 m, respectively, of stromatolitic carbonate (Grainger 2001). The large volume of Murdama Basin sedimentary material, testified by its surface area of c. 72,000 km², including a concealed 200 km extension beneath Phanerozoic rocks SE of the Arabian Shield (Johnson & Stewart 1995), a reported thickness of >8000 m, and exhumation of amphibolite and granulite facies rocks at the sub-Murdama unconformity, imply that the sub-Murdama Afif and adjacent terranes were extensively eroded soon after orogenic uplift (Cole 1988). Furthermore, the carbonate build-ups preserved in the Murdama, Fatima and Bani Ghayy Basins suggest that even if terrane assembly caused orogenic uplift, large parts of the northeastern Arabian Shield were reduced to sea level and developed connections to the ocean flanking the emerging Gondwana supercontinent within a few million years of orogeny. Basin deposition was terminated by the onset of folding, which, in the case of the Murdama and Bani Ghayy groups, denotes significant E–W shortening in the eastern shield soon after 650 Ma (Johnson 2003).

Epicontinental terrestrial basins

Other overlap assemblages in the ANS were deposited between 680 and c. 580 Ma in shallow water and terrestrial, fault-controlled down sags, pull aparts, rifts, half-grabens and calderas (e.g. Stern et al. 1984; Willis et al. 1988; Jarrar et al. 1993; El-Kalilouby 1996; Wilde & Youssef 2000). The rocks are molassic, characterized by abundant hematite stain and cement, a coarse clastic grain size, including cobbles and boulders in many conglomerates, and bimodal volcanic chemistry. Rocks in some basins were gently folded, but in others were intensely deformed and metamorphosed to greenschist and amphibolite facies depending on basin proximity to Late Neoproterozoic shear zones (Fowler & Osman 2001). Some, but not all, of the basins contain thin successions of thin-bedded stromatolitic limestone and some contain dachnites, which together with reported dropstones might be evidence of Marinoan glaciation. The stratigraphic distribution of the basins (Fig. 13) implies that several periods of extension affected the ANS after the peak of orogeny, but deformation of the basins, conversely, indicates that extension was interrupted by compression and brittle–ductile shearing, and gives evidence that Late Neoproterozoic extension was not steady state but alternated with shortening, diastrophism and, in some cases, intense metamorphism (cf. Greiling et al. 1994). Although many basins were surrounded by local high relief, as evidenced by the coarse grain size of typical deposits, the Hammamat Basin, and possibly others, may have been fed by far-flowing river systems rather than by local transport into restricted intermontane basins (Wilde & Youssef 2001).

Late Neoproterozoic gneiss belts and domes

Another expression of diachronous post-amalgamation deformation and crust-forming events is the development of Late Neoproterozoic antiforms and domes of ortho- and para-gneisses (Figs. 12 & 14). Well known in the Eastern Desert of Egypt and Sinai as examples of metamorphic core complexes associated with late-orogenic extension (e.g. Fritz et al. 1996; Blasband et al. 1997) or imbricated stacks of antiformal thrust duplexes (e.g. Geering 1997; Fowler & Osman 2001), Late Neoproterozoic gneiss domes and antiforms also occur along Najd faults in the Arabian Shield (Fig. 14) and farther south in the Nubian Shield.

The Kirsh gneiss (Fig. 15), an example of such gneiss in the Arabian Shield, is an antiform of strongly foliated biotite monzogranite orthogneiss (Al Hawriyeh Anticlinorium) intruded into steeply dipping kyanite–quartz schist and subvertical zones of mylonite and ultramylonite at the southeastern end of the Ar Rika-Qazaz Shear System, the main trans-Arabian Shield Najd Structure (Fig. 14). The gentle plunge of the northwesterly trending mineral and elongate-pebble lineations indicates a large component of constriction, or unidirectional stretching, and in several sectors of the shear zone the deformed rocks are L-tectonites. The monzogranite is located in a zone of extension between left-stepping faults along the sinistral Ar Rika Fault Zone. On the presumption that the Al Khushaymiyah complex to the NE, which is composed of massive monzogranite, is an undeformed equivalent of the orthogneiss, activity on this section of the Ar Rika Shear Zone is dated at c. 610 Ma.

Well-documented examples of gneiss domes in the Eastern Desert and Sinai (Fig. 12) include the structures at Mea'qit (Sturchio et al. 1983; Habib et al. 1985), Gebel el-Shalul (Hamimi et al. 1994; Osman 1996), Gebel El-Sibai (Kamal El-Din et al. 1992; Bregar 1996), Um Had (Fowler & Osman 2001) and Wadi Kid (Blasband et al. 1997), as well
as the Hafaif Culmination (El-Ramly et al. 1984; Rashwan 1991; Greiling 1997), but though closely studied, their origins, ages and tectonic significances are debated. The chief common structural elements include: (1) an antiformal structure; (2) a prevailing NNW to NW orientation of stretching lineations and thrust transport direction; and (3) NW-trending sinistral transcurrent shears at the margins of and within the domes, which may be true subvertical shears or folded and steepened mylonitic thrust zones. Less common elements observed at some of the domes include: lineation and transport directions to the SW or NE, as well as to the NW; SW- or NE-vergent thrusts in addition to NW-vergent thrusts; NW-directed thrusts on their southern margin and NW-directed low-angle normal faults on their northern margin; or low-angle normal faults on both margins.

Geothermobarometric experiments on samples from the Meatiq Dome indicate high \( P-T \) core conditions (8 kbar and 750–600°C; Neumayr et al. 1996), and the juxtaposition of high-grade core rocks and greenschist facies envelopes implies domal uplift of as much as 10–13 km (Fowler & Osman 2001). Conditions at Wadi Kid were not as extreme, reaching \( P-T \) conditions of 3.4–4.2 kbar and 684–488°C during dome exhumation (Brooijmans et al. 2003). The presence of Hammamat and Dhokhan rocks in pull-apart basins on the flanks of the domes and in thrust sheets above the domes, as well as dating on pre-, syn- and post-tectonic plutonic rocks and metamorphic minerals, constrain doming, thrusting, shearing and exhumation to sometime after 615 Ma and prior to 585 Ma (Fritz et al. 1996).

Gneiss domes that may also reflect Late
Neoproterozoic extension and exhumation have been described recently from Eritrea (Fig. 7). The Aneseba Gneiss (protoliths aged 700–796 Ma; Teklay 1997) comprises well-banded high-grade garnet–hornblende gneiss, two-mica gneiss and migmatite (Woldehaimanot 2001). D1 deformation affected the hornblende gneiss only; D2 deformation affected hornblende and mica gneisses; and D3 deformation affected all rocks. An early thrust event manifested by imbricate structures showing an east-directed shearing is superimposed by a widespread extensional event marked by mesoscopic ductile shear zones with normal sense-of-slip coincident with moderately NW-plunging stretching lineations, and kinematic markers (S-type asymmetrically folded quartz/pegmatite veins, rotated, pre-shearing foliation and σ-type tailed porphyroblasts of K-feldspar) indicating top-to-NW shear. This extensional event deformed earlier synmetamorphic planar fabrics and is interpreted as a late, post-metamorphic event reflecting late-stage exhumation and extension. In the Ghedem area (Ghebreab 1999) the dominant exposures are low domes of garnetiferous orthogneiss structurally overlain by garnet-bearing staurolite–kyanite paragneisses. An anticlockwise $P$–$T$ path of metamorphism, determined from mineral assemblages included in cores, outer cores and rim matrices of garnets, indicate that the rocks of the Ghedem Terrane were subjected to heating with loading at intermediate crustal levels followed by near-isothermal loading at deep crustal levels, and were subsequently subjected to decompression and cooling associated with rapid exhumation from a minimum depth of 44 km (Ghebreab 1999). A different study of the high-grade metasedimentary rocks in the Ghedem area (Beyth et al. 1997) deduces a clockwise $P$–$T$ path of thermobarometry with peak conditions of 8–10 kbar and c. 700°C, and proposes a collision setting for the metamorphism followed by extension on the low-angle shear zones that are observed in the area (Beyth et al. 2003).

**Late Neoproterozoic shear zones**

Shears are a third class of structure that reflect tectonic heterogeneity during the Late Neoproterozoic history of the ANS. They include: (1) north-trending
subvertical brittle–ductile shear zones in the southern part of the shield; (2) the Hamisana and Oko Shear Zones in the Haya and Gebeit Terranes; and (3) NW- and NE-trending transcurrent faults of the Najd system.

North-trending shear zones

The brittle–ductile shears in the southern shield (Fig. 7) are zones of phyllonite, schist, gneiss, mylonite, and subordinate serpentinite and serpentinite schist. They crop out as belts of strongly deformed rocks, 400 km long and 1 km wide and contain an abundance of dextral and sinistral kinematic markers (S–C fabric, rotated and winged porphyroclasts). Some belts contain sheared serpentinite (Fig. 11) and certain of them, the easternmost Nabibah Fault Zone, in particular, are interpreted as sutures within the Asir Composite Terrane, but the structures are currently topics of research because recent work shows that similar shear zones in western Ethiopia and the Mozambique Belt in Madagascar formed late in the EAO orogenic cycle as a result of E–W shortening and orogen-parallel extension (Martelat et al. 2000; Braathen et al. 2001; de Wit et al. 2001). A few of the shear zones in the Asir Terrane, such as the Tabalah–Tarj Zone discussed above, appear to be old and related to an early phase of terrane amalgamation. Others underwent displacements as late as <640 Ma (Johnson et al. 2001) and may be examples of Mozambique-Belt-type shears that reworked older sutures or part of the Late Neoproterozoic deformation event that created the Najd faults (Nehlig et al. 2002).

Prominent sinuous north-trending shear zones in the Haya and Gebeit Terranes (Fig. 12) accommodated large amounts of E–W shortening. The Hamisana Shear Zone postdates and deforms the Allaqi–Heiani–Onib–Sol Hamed Suture, and was active between 665 and 610 Ma (Stern & Kröner 1993). It is a belt of strongly foliated and lineated amphibolite and amphibolite-grade paragneiss and biotite–muscovite orthogneiss (de Wall et al. 2001) that resulted from a high-T metamorphic event superimposed on the greenstitch metamorphic assemblages of the adjacent rocks. Pervasive L–S fabrics and an absence of S–C fabrics indicate that
deformation was mainly constrictional, consistent with inferred E–W regional compression (Miller & Dixon 1992; de Wall et al. 2001). Subsequent low-T retrograde metamorphism was associated with local NE-trending dextral shearing (de Wall et al. 2001). The longitudinal extent of the Hamisana Shear Zone is unknown: to the south it passes under eolian sand so that its relations to the Nakasib or Kerf Sutures are obscured; to the north, the zone projects into outcrops of ortho- and paragneiss extensively exposed between the Onib–Sol Hamed Suture and the Gebel Gerf Ophiolite Nappe, in which the Hamisana Shear Zone loses its character as a discrete narrow zone of high strain. de Wall et al. (2001) conclude that the Hamisana Shear Zone is an expression of orogenic compressional deformation unrelated to either large-scale transpression or major escape tectonics; Stern & Kröner (1993) suggest that it may be a Mozambique-type zone of shortening.

The Oko Shear Zone, superimposed on the Nakasib Suture, likewise accommodated a large amount of E–W shortening (Abdelsalam 1994; Abdelsalam & Stern 1996), but contains additional structures that denote a more complex evolution than that of the Hamisana Shear Zone. The shear zone was active between 700 and 560 Ma (Abdelsalam & Stern 1993b) and developed in stages of: (1) E–W shortening marked by the development of north-trending upright interference folds in pre-existing Nakasib Suture fabrics; (2) NW-trending sinistral and subordinate NE-trending dextral subhorizontal strike-slip faulting, which displaced the Nakasib Suture c. 10 km and rotated E–W structures into N–S trends; and (3) terminal east- and west-vergent thrusting and buckling that created a flower structure (Abdelsalam & Stern 1996).

**Najd Fault System – issues of timing and origin**

The Najd Fault System was originally defined for Late Neoproterozoic NW- and subordinate NE-trending transcurrent faults in the northern part of the Arabian Shield (Fig. 12; Brown & Jackson 1960; Brown 1972; Moore 1979). Faults correlated with the Najd System are known in the southern Arabian Shield, in the subsurface of Oman and in the Nubian Shield, and similar NW-trending sinistral faults are known in the EAO in southern Ethiopia, Tanzania and Madagascar (Fig. 2).

The faults are conventionally interpreted as having a common history and tectonic setting but, in detail, have significant differences in age of activity and structural style (Johnson & Kattan 1999). The Halaban–Zarghat Fault Zone was chiefly active between <640 and 620 Ma (Cole & Hedge 1986); the Ar Rika–Qazaz Shear Zone was strongly active c. <640–610 Ma; Najd faults that bound the Meatiq and other gneiss domes in the Eastern Desert were active between 615 and 585 Ma (Fritz et al. 1996); and the so-called Najd Rift System that operated in Oman as the putative cause of the Ara Group salt basins is inferred to have been active between 570 and 530 Ma (Al-Husseini 2000). Movement persisted on the Halaban–Zarghat Fault until the cessation of Jibalah deposition at c. 570 Ma. Final movements on the Ruwah Fault Zone, interpreted by Johnson & Kattan (2001) to be a Najd-reactivated segment of a c. 680 Ma suture zone, cataclastically deformed a 592 Ma granite (Stoesser & Stacey 1988). The Ar Rika–Qazaz Shear Zone contains gabbro and quartz syenite plugs that yield whole-rock K–Ar ages of 512 ± 17 and 487 ± 17 Ma (Brown et al. 1989), indicating significant Cambrian cooling. The NE-trending faults are dextral and the NW-trending faults are largely, but not entirely, sinistral: Agar (1987), Matsah & Kusky (2001) Kusky & Matsah (2003) argue that a dextral phase preceded the dominant sinistral phase, although the chronologic data of Cole & Hedge (1986) would question this argument in the case of the Halaban–Zarghat Fault. The two fault sets appear to form a conjugate pair. Complementary vertical movements of >10 km on parts of the Najd faults are indicated by the juxtaposition of amphibolite-grade mylonitic gneiss along the fault axes, with greenschist facies rocks at the fault margins, and by ductile–brittle transitions along the fault zones that, in typical transcurrent fault systems, would have originated at depths of 10–15 km (Davis & Reynolds 1996). Vertical uplifts of as much as 5 km across the faults are implied also by the regional gentle plunge of stretching lineations (Davies 1984).

Many workers infer that the Najd Fault System formed under general conditions of compression. Soon after the publication of work by Molnar & Tapponnier (1977) on the application of slip-line theory to faulting north of the Himalayas, proposals were made that the Najd faults resulted from indentation of the ANS by a rigid continental plate – from the east in the case of Schmidt et al. (1979) and Davies (1984), and from the west in the case of Fleck et al. (1980). Agar (1987) and Abdelsalam (1994) argued that the fault system resulted from E–W shortening and collision, whereas Burke & Sengör (1986) advocated an origin during E–W collision and simultaneous orogen-parallel extension and escape. Abdelsalam & Stern (1996) modified the collisional interpretation by postulating that the direction of maximum strain during Najd faulting (σ1) was oriented NW–SE, oblique to the axis of the EAO rather than orthogonal, as a result of regional transpression caused by the entrapment of the ANS between the East Saharan Craton on the west and an Ar Rayn microplate on the east. Stern (1994)
suggested that the NW-trending shear zones were a mechanism that allowed northward orogen-parallel extension as the ANS escaped from collision between East and West Gondwana. In Oman, Loosveld et al. (1996) concluded that Late Precambrian north-trending folds and thrust faults, as well as the NE-trending Ara Group salt basins (part of the Huqf Supergroup), resulted from E–W directed compression associated with movement on presumed extensions of the Najd Fault System into the eastern part of the Arabian Peninsula shown in magnetic and gravity data. This concept was supported by Hall et al. (2001), who favoured a compressive origin of the Ghaba Basin as a push-down basin between overriding thrusts. Fowler & Osman (2001) presented a model that joins Najd faulting, extension, bidirectional thrusting and gneiss doming in the Eastern Desert in a concept of escape tectonics engendered by general E–W compression, and alternating NNW-directed extrusion and NE–SW-directed thrusting in the EAO away from a pivotal collision zone in the Mozambique Belt.

Other interpretations situate the Najd faults in an extensional stress field. Hussein & Hussein (1990) and Al-Husseini (2000), in discussing the origin of the Omani Salt Basins, envisaged that the Najd structures were part of a 570–530 Ma transform fault system on the northern flank of the Gondwana supercontinent that linked rifts in the eastern Arabian Plate (Oman), the northwestern Indian Plate (Punjab) and the Eurasian Plate (Dibbab Rift and Derik Rift), with extensional-collapse structures in the ANS. Based on the kinematic implications of NE-trending dyke swarms and volcanic grabens in the northern Eastern Desert, the Midyan Terrane and Jordan, Stern (1985) proposed that the Najd System was a set of transform faults caused by pervasive NW–SE-directed Late Neoproterozoic extension during the period of 600–575 Ma. Mercoll et al. (2001) report two periods of dyke emplacement in the basement in Oman: one event at c. 750 Ma when the crystalline basement was uplifted and intruded by a large mass of granite dykes along a conjugate set of compressional brittle faults and a second event at c. 550 Ma marked by the emplacement of a suite of calc-alkaline basalt, andesite and rhyolite dykes. Both events are consistent with extension and potential rifting. Emplacement of the Mukeiras Dyke Swarm (c. 715–615 Ma) in the Al Bayda Terrane in Yemen is believed to reflect a similar period of post-collisional rifting in the southern part of the Arabian Shield (Whitehouse et al. 1998).

Clearly, there are conflicting issues of timing and kinematics concerning the Najd Fault System that need to be resolved by additional structural mapping (e.g. Shackleton 1994) or by redefining the system or excluding some faults that conventionally are included in the system (Johnson & Kattan 1999).

The core issue is whether the Najd faults formed under conditions of shortening and compression or extension and rifting. Structurally, the two conditions are not mutually exclusive, and temporally and spatially may grade from one to the other, but conceptually the difference between ‘push’ and ‘pull’ is major and raises the question of whether the faults relate to the final convergence of East and West Gondwana or the breakup and rifting of the newly formed Gondwana supercontinent. In the present authors’ opinion, Late Neoproterozoic transcurrent faults that meet broad criteria for inclusion in the Najd Fault System in the Najd region of the central Arabian Shield (the type area for the system), the Midyan Terrane and the Eastern Desert have structural features that strongly favour a compressive origin. These include: (1) a conjugate system geometry with common indications of sinistral horizontal movement on NW-trending faults and dextral movement on NE faults; (2) the presence, in the Murdama Group adjacent to the Ar Rika Fault, of north-trending en-echelon folds of the type expected from sinistral transcurrent movement (Johnson 2003); (3) an abundance of NW-trending stretching lineations along the fault zones; (4) gradations between amphibolite grade, ductile and greenschist facies, brittle conditions of metamorphism and deformation along the faults; and (5) the location of Jibalah Group depositional basins and orthogneiss domes at releasing bends between left-stepping sinistral faults. The orientations of the faults and folds are consistent with a broad E–W orientation of σ1, and would be consistent with the general regime envisaged by many workers of Late Neoproterozoic E–W shortening, which resolved into NW- and NE-directed shearing.

Whatever their origin, most movements on the Najd faults ceased by the Early Cambrian, at which time a thick succession of post-fault continental to shallow-marine Lower Palaeozoic siliciclastic rocks began to cover the region. The siliciclastic rocks were deposited on a regional unconformity created by the erosion and depression of what had become a stable platform, and were distributed by north- and NE-flowing fluvial systems from sources in interior Gondwana (Konert et al. 2001). Subsequent movements on the Najd faults were entirely brittle and presumably caused by adjustments of the faults, as zones of crustal weakness, to Phanerozoic plate movements. Phanerozoic movements are evidenced by post-sedimentary faulting in the Wajid Group on the line of the Ruwah Fault at the SE flank of the Arabian Shield and by displacement of Cretaceous formations on the line of the Ar Rika Shear Zone in central Arabia (André 1989).
Discussion

A large body of substantial data indicates that the ANS evolved in the Mozambique Ocean between the converging blocks of East and West Gondwana (Stern 1994). However, representative rocks of West Gondwana only crop out in poorly exposed, commonly reworked form in the East Saharan Craton and representatives of East Gondwana in Arabia are unknown. By default, therefore, the ANS is the chief source of information about convergence and supercontinental assembly at the Tethyan margin of Gondwana. The timing of terrane accretion and amalgamation (Fig. 3) helps to constrain the history of closure of the Mozambique Ocean and the initial convergence of East and West Gondwana; the post-amalgamation events constrain models about the final contact of East and West Gondwana, and Gondwana assembly.

As indicated by the weakly constrained age of the oldest (Jabal Tharwah, Bi'r Umq, Bi'r Tuluhah) and the better constrained age of the youngest (Halaban) ophiolites in the ANS, the Mozambique Ocean existed by 870Ma and continued to grow until c. 695Ma, suggesting that Rodinia, in the region of the eventual ANS, had started to rift by 870Ma. Oceanic-basin closure at subduction zones, indicated by the creation of juvenile oceanic and continental-margin arcs, commenced soon after rifting and continued until completion c. 600Ma by creation of the Keraf Suture. Because of a paucity of palaeopole data, the pre-amalgamation spatial relationships among the terranes in the ANS are unknown. Simplistically working back from their amalgamated relations, and acknowledging that terrane analysis in the ANS is still provisional, relations such as those shown in Figure 16 are envisaged. The figure depicts rifted segments of Rodinia – a large segment in the west (present-day coordinates), which eventually became West Gondwana and a smaller segment in the east, which was entrained in the Afif Terrane as part of the Khida Subterranee – and a juvenile ocean between them. Rocks that became the Halfa and Bayuda Terranes were evolving at or close to the western Rodinia fragment, acquiring both mature and juvenile isotopic signatures, and volcanic and plutonic rocks with juvenile signatures were forming within the ocean. By 780–760Ma (Fig. 16a), the earliest intracratonic arcs were in contact and amalgamation was in progress along probably intracratere sutures within the Asir Terrane and, during a period of dextral transpression, along the Bi'r Umq–Nakasib Suture. Over the next 100Ma (Fig. 16b), the Bayuda/Halfa Terranes converged with each other and with the East Saharan Craton, marking the onset of approach of West Gondwana, and composite intraoceanic crustal units composed of the Asir–Barka–Hager/Tokar–Nakfa–Ghedum Terranes, the Haya–Gebeit–Eastern Desert–Midyan Terranes and the Afif Subterranes assembled and converged. Approach of the continental-margin Ar Rayn Terrane from the east completed amalgamation of the Arabian Shield by 680–640Ma (Fig. 16c) and marked closure of the eastern part of the Mozambique Ocean. By 600Ma (Fig. 16d), the assembly of the ANS was complete and the Mozambique Ocean was fully consumed by the creation of the Keraf Suture. It is envisaged that by this stage East and West Gondwana were in contact across the amalgamated rocks of the ANS, although the absence of proven representatives of East Gondwana in the region makes its tectonic role speculative. Metallogenic and geochemical investigations in the Ar Rayn Terrane (Doebrich et al. 2001) interpretations of the Huqf Supergroup as a type of foreland basin (Cosca et al. 2001), and the magnetic character of eastern Arabia (Johnson & Stewart 1995) are permissive of continental crust east of the Arabian Shield, but its presence and provenance are yet to be established.

The amalgamated terranes exposed at the present level of erosion in the ANS constitute an orogenic belt minimally 1200km wide in a E–W direction that at the time of orogeny would have formed part of an enormous mountain chain with a large root extending deep into the lithosphere. Since then, the root has been lost, and the present-day crust of the Arabian and northeastern African Plates displays a layered structure and a uniform Moho depth of 35–45km (Gettings et al. 1986; Rodgers et al. 1999) that retains no vestige of crustal thickening. Modern geophysical measurements consequently provide little clue as to the Late Neoproterozoic fate of the mountain belt. The geological data, conversely, gives ample evidence of Late Neoproterozoic uplift, erosion, extension, depression and compression before transformation into a palaeo–Tethys passive margin.

The ANS lacks large areas of exhumed granulite facies rocks of the type found farther south in the Mozambique Belt (Stern 1994), implying there was no large-scale isostatic rebound following the relaxation of confining orogenic stress, but small exposures of granulite facies and more extensive amphibolite facies regional metamorphism are evidence that parts of the shield were, indeed, uplifted by 10–20km. Additional evidence of uplift comes from the P–T conditions of gneiss domes described along the Najd Fault System and from Ar/Ar evidence of rapid cooling in the eastern Arabian Shield (Al-Saleh et al. 1998; Al-Saleh & Boyle 2001b), Sinai (Cosca et al. 1999) and the western Nubian Shield (Abdelsalam et al. 1998). Judging by the onset of post-amalgamation basin deposition, uplift and erosion occurred by 725 Ma east of the shield (in
Fig. 16. Sketch of the assembly of the Arabian–Nubian Shield in four time slices showing progress of terrane amalgamation and eventual accretion to East and West Gondwana. A, 780–760 Ma; B, 750–660 Ma; C, 680–640 Ma; D, 650–600 Ma. Terranes and subterranes: B, Bayda; D, Ad Dawadimi; EDM, Eastern Desert–Midyan; GH, Gebeit–Hijaz; Ha, Halfa; HJ, Haya–Jiddah; K, Kirsh; N, Nuqrah; R, Ar Raayn; Si, Siham; Sw, Suwaj. Small arrows show sense of frontal convergence; half arrows show sense of transpression; large arrows show inferred convergence trajectories of terranes and composite terranes.

Oman), and by 680–670 Ma in the Arabian and Nubian Shields. The gneiss domes denote uplifts of as much as 44 km between 615 and 585 Ma, and the $^{40}$Ar/$^{39}$Ar data indicate uplifts of between 600 and 580 Ma (Fig. 13).

Post-amalgamation erosion was intense, producing enormous volumes of debris that filled the Murdama Basin and others, and, coupled with depression, effectively reduced large parts of the eastern ANS to sea level shortly after peak orogeny. The large numbers of fault-controlled basins, starting with the Fatima Group and ending with the Hammamat, Dokhan, Saramuj and Jibalah Groups (Fig. 13), are evidence that depression was linked to
extension and local rifting throughout the period of c. 680–560 Ma. On the other hand, deformation of the large Murdama and Huqf Supergroup Basins may have been driven by more complex mechanisms, such as thrusting and downflexing, as well as rifting in the case of the Huqf and thermal contraction and subduction delamination in the case of the Murdama. But evidence from the same basins demonstrates that extension was periodic and alternated with compression, which folded successively and sometimes sheared, thrusted and metamorphosed all the post-amalgamation basins (with the possible exception of the Asoteraiba), causing unconformities in places where younger basins are atop older basins. As late as 600 Ma and possibly younger in the case of the Oko zone, the southern part of the ANS underwent E–W shortening and north-directed shearing resulting in the Hamisana Zone, the transpressional Keraf Suture and later shear zones in Asir. Between 640 and 560 Ma the entire ANS underwent NW- and NE-directed Najd transcurrent faulting, and between 615 and 585 Ma parts of the ANS, particularly the NW, experienced thrusting and low-angle normal faulting during the creation of gneiss domes.

The range of post-amalgamation events and the variations in orientations of the local strain ellipsoids in time and space implyed by the varying orientations of the observed structures, militates against any simple model for the final assembly of Gondwana. Nonetheless, the far-field motion (Fig. 16d) implied by sinistral tranposition on the Keraf Suture, the dominant NW–SE direction of thrusting evident at some gneiss domes and the predominant NW strike of Najd faults, are consistent with the broad E–W to NW–SE uniform or scissor-like direction of Gondwana convergence modelled by a number of authors (e.g. Stern 1994; Greiling et al. 2000; Fowler & Osman 2001). Post-amalgamation tectonic models that envisage only extension and collapse driven by gravitational instability (e.g. Blasband et al. 2000) do not account for the periods of shortening that interrupted extension, and models of Najd faulting that rely on passive-margin rifting do not adequately account for E–W compression. Many details of the timing and kinematics of events during the final convergence of East and West Gondwana are still problematic. Nonetheless, it is clear that adequate models of Late Neoproterozoic Gondwana convergence at the northern end of the EAO will need to accommodate both E–W and NW–SE extension and shortening soon after orogenic climax, as well as orogen-parallel extension because of slip on the Najd faults, crustal depression, the creation of depositional basins and a westerly penetration of seaways into the core of the orogen.

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