Annu. Rev. Earth Planet. Sci. 1994. 22:319–51

ARC ASSEMBLY AND CONTINENTAL COLLISION IN THE NEOPROTEROZOIC EAST AFRICAN OROGEN: Implications for the Consolidation of Gondwanaland

Robert J. Stern

Center for Lithospheric Studies, University of Texas at Dallas, Richardson, Texas 75083-0688

KEY WORDS: tectonics, crustal evolution

INTRODUCTION

Some of the most important, rapid, and enigmatic changes in our Earth's environment and biota occurred during the Neoproterozoic Era (1000–540 million years ago; Ma). Paramount among these changes are the rapid evolution of eukaryotes and appearance of metazoa (Knoll 1992, Conway Morris 1993), major episodes of continental glaciation that may have extended to low latitudes (Hambrey & Harland 1985), marked increases in the oxygen concentration of the atmosphere and hydrosphere (Derry et al 1992), the reappearance of sedimentary banded iron formations (BIF; James 1983), and striking temporal variations in the isotopic composition of C and Sr (Asmerom et al 1991, Derry et al 1992). Understanding the causes of and relationships between these changes is a challenging focus of interdisciplinary research, and there are compelling indications that the most important causes were tectonic (Des Marais et al 1992, Veevers 1990). For example, development of ocean basins may have been accompanied by the development of seafloor hydrothermal systems, which lowered the $^{87}Sr/^{86}Sr$ of seawater, led to the development of BIF, and formed anoxic basins where organic carbon could be buried, thus leading to an increase in O_2. Continental collision and formation of a supercontinent may have led to continental glaciation and an increase in the $^{87}Sr/^{86}Sr$ of seawater,

0084–6597/94/0515–0319$05.00

and could also have formed sites (high sedimentation rate submarine fans) where organic carbon was buried. Understanding the style and sequence of Neoproterozoic tectonic events is fundamental if we are to establish the cause-and-effect relationship between tectonics and Neoproterozoic global change.

The Neoproterozoic Era encompasses a protracted orogenic cycle referred to here as the "Pan-African" orogenic cycle. Kennedy (1964) coined the term "Pan-African Thermo-Tectonic Episode" to characterize the structural differentiation of Africa into cratons and mobile belts during the latest Precambrian and earliest Paleozoic, but the Pan-African has been redefined (Kröner 1984) as involving a protracted orogenic cycle from 950 to 450 Ma. This tectonism was not limited to Africa—diastrophic events of similar age and style are common throughout Gondwanaland (Figure 1) and in many parts of Laurasia. The time span of 500 million years is much longer than any Phanerozoic orogeny, so reference to a Pan-African "orogenic cycle" must suffice until the timing and regional extent of discrete tectonic events is better constrained.

Not only will questions about the evolving Neoproterozoic environment be answered when we better understand Pan-African tectonics, but the Pan-African orogenic cycle is interesting in its own right. The nature of the episode supports arguments that modern plate tectonic systems had begun by about 1000 Ma (Davies 1992). In contrast to crust produced between 1.0 and 1.8 Ga, Pan-African juvenile (i.e. mantle-derived) crust preserves most of the hallmarks of modern plate tectonic regimes, including abundant ophiolites, calc-alkaline batholiths and volcanic sequences, and immature clastic sediments. Horizontal tectonic displacements are proved by ophiolitic nappes which traveled at least several tens and perhaps a few hundreds of kilometers. In many instances, Pan-African metamorphism resulted in granulite formation, suggesting crustal overthickening due to continent-continent collision (Burke & Dewey 1972).

McWilliams (1981) argued that Gondwanaland was assembled during the Neoproterozoic from two fragments—East and West Gondwanaland—along the Pan-African Mozambique Belt of East Africa (Figure 1). This suggestion is supported by the lithologic associations characteristic of the Mozambique Belt and its extension to the north, the Arabian-Nubian Shield (ANS). Following the arguments of Berhe (1990) that ophiolite-decorated meridional sutures can be traced in E. Africa, I accept these entities as along-strike correlatives, and refer to the larger structure as the East African Orogen (EAO; Figure 2). The ophiolites, granulites, and structures of the EAO are fossil fragments of a Neoproterozoic Wilson cycle, representing the opening and closing of an ocean basin that lay between the older crustal blocks of East and West Gondwanaland. The

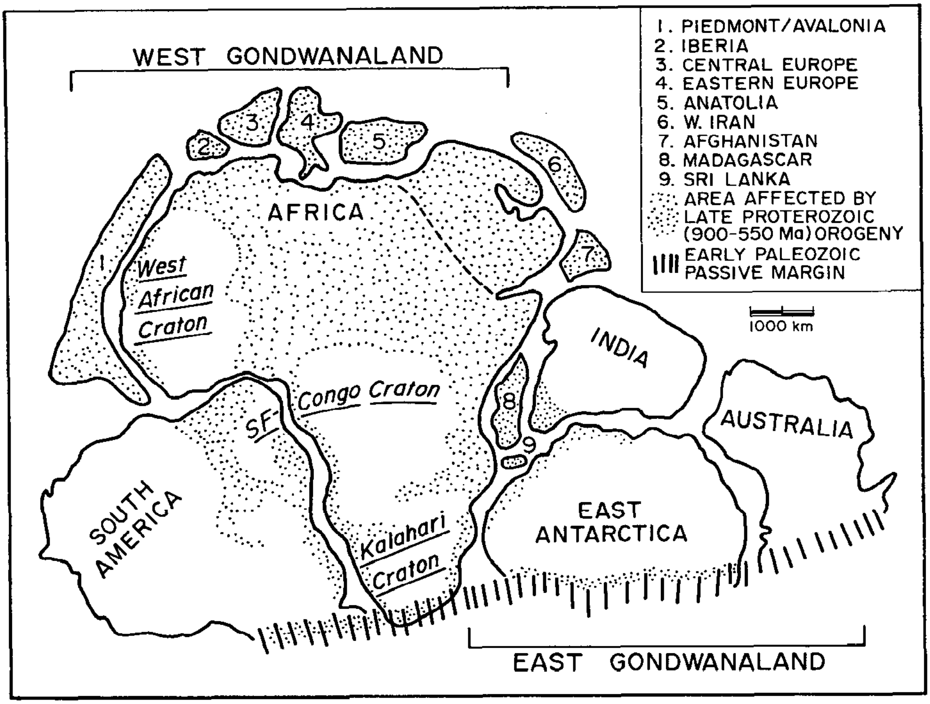
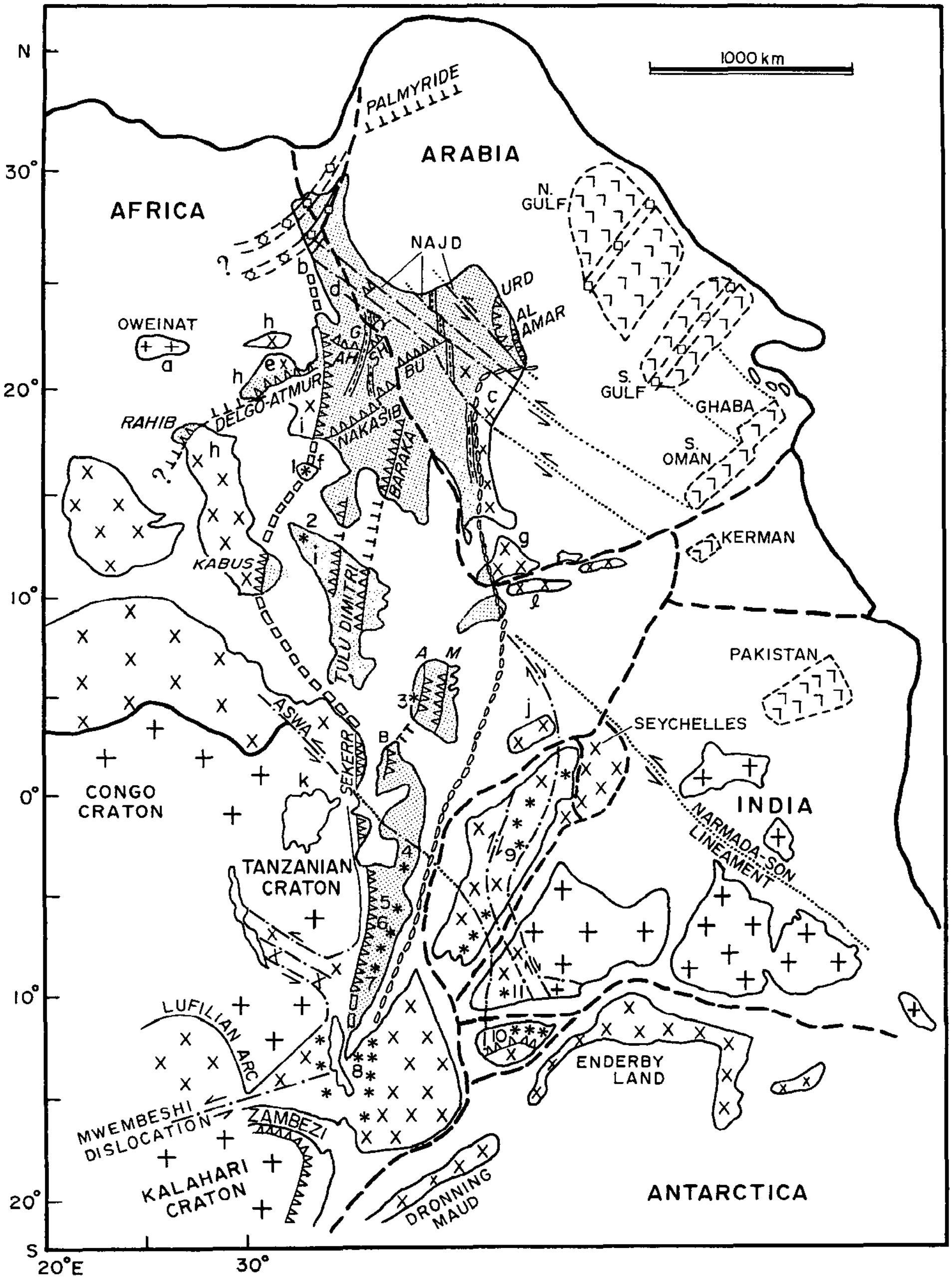


Figure 1 Map of a part of Gondwanaland, showing the position of the continents and smaller continental fragments in the early Mesozoic (De Wit et al 1988). This is the surviving nucleus of the supercontinent of Greater Gondwanaland which at the end of the Precambrian was about 25% larger in area, mostly on the Australian, Indian, Arabian, and NE African margins. Stippled area shows the region affected by the 950–450 Ma Pan-African event. SF = Sao Francisco

length and breadth of the EAO are similar to those of more recent orogenic belts, such as the North America cordillera (Oldow et al 1989) and of the mountain belt that stretches from western Europe to eastern Asia—which Şengör (1987) calls the "Tethysides orogenic collage" (Figure 3). The time span involved in all these orogens is similar—several hundreds of millions of years. Other Wilson cycles of similar or younger age affected West Gondwanaland, but the abundance of ophiolites and juvenile arc assemblages indicates that the EAO formed during a protracted period of juvenile crust formation in intra-oceanic arcs culminating in continental collision. The Neoproterozoic assembly of Gondwanaland was more complicated than simply bringing two halves together, and West Gondwanaland in particular was assembled from several pieces (see Figure 2 of


N
30°
20°
10°
0°
10°
20°
S
20°E
30°
1000 km
PALMYRIDE
ARABIA
AFRICA
N. GULF
NAJD
URD
AL AMAR
OWEINAT
S. GULF
GHABA
S. OMAN
RAHIB
DELGO-ATMUR
NAKASIB
BARAKA
KABUS
TULU DIMITRI
KERMAN
PAKISTAN
ASWA
SEKERR
SEYCHELLES
INDIA
CONGO CRATON
TANZANIAN CRATON
NARMADA-SON LINEAMENT
LUFILIAN ARC
MWEMBESHI DISLOCATION
ZAMBEZI
ENDERBY LAND
KALAHARI CRATON
DRONNING MAUD
ANTARCTICA

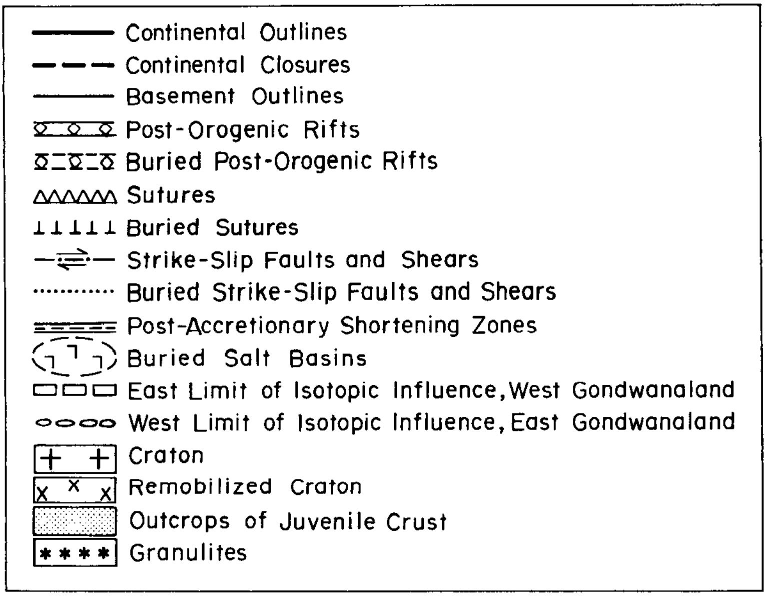


Figure 2 Tectonic map of the East African Orogen. Continental fragments are shown as configured at the end of the Precambrian.

Small letters (a–l) are places where pre–Pan African crustal components can be documented: a = Oweinat (2.5 Ga basement; Klerkx & Deutsch 1977); b = Nakhil (1.6 Ga xenocrystic zircons in 580 Ma granite; Sultan et al 1990); c = Afif terrane (1.6–1.8 Ga crust; Stacey & Hedge 1984, Stoeser et al 1991); d = E. Desert of Egypt (1.1–2.3 Ga cobbles in Atud conglomerate, Dixon 1981; 1.5–2.7 Ga detrital zircons in wackes, Wust et al 1987); e = Wadi Halfa (1.2–2.5 Ga basement; Stern et al 1993); f = Sabaloka (1.0–2.6 Ga detrital zircons in metasedimentary granulites; Kröner et al 1987); g = Eastern Yemen (<3.0 Ga T_{DM} Nd model ages; Stoeser et al 1991, Whitehouse et al 1993); h = W. Egypt and Sudan (1.5–2.7 Ga T_{DM} Nd model ages, Harms et al 1990; 1.9–2.6 Ga zircon ages, Sultan et al 1992b); i = central Sudan (1.5 Ga T_{DM} Nd model age; Harris et al 1984); j = S. Somalia (1.8–2.5 Ga inherited zircons in 540 Ma granites; Küster et al 1990); k = Uganda (2.6–2.9 Ga basement; Leggo 1974); l = N. Somalia (1.8 Ga xenocrystic zircons in 715–820 Ma granites; Kröner et al 1989).

Numbers 1–11 denote Neoproterozoic granulites: 1 = Sabaloka (710 Ma zircon age; Kröner et al 1987); 2 = Jebel Moya (740 Ma zircon age; Stern & Dawoud 1991); 3 = Bergudda Complex (545 Ma zircon age; Ayalew & Gichile 1990); 4 = Pare Mountains (645 ± 10 Ma U-Pb zircon age; Muhongo & Lenoir 1993) and Wami River (715 Ma zircon age; Maboko et al 1985); 5 = Uluguru (695 ± 4 Ma U-Pb zircon age, Muhongo & Lenoir 1993; 630 Ma hornblende Ar closure age, Maboko et al 1989); 6 = Furua (650 Ma zircon age, Coolen et al 1982; 615–665 hornblende Ar closure age, Andriessen et al 1985); 7 = Songea; 8 = Malawi and Mozambique (800–1100 Ma Rb-Sr whole rock; Andreoli 1991); 9 = Madagascar (860 Ma; Cahen et al 1984); 10 = Sri Lanka (520–550 Ma; Burton & O'Nions 1990, Kröner 1991); 11 = S. India (Choudhary et al 1992).

Capital letters refer to ophiolites and/or sutures: G = Gerf ophiolite; AH = Allaqi-Heiani suture; SH = Sol Hamed suture; BU = Bir Umq suture; A = Adola ophiolite; M = Moyale ophiolite; B = Baragoi ophiolite.

Dalziel 1992), but the timing and sequence of events preserved in the EAO are representative of the consolidation of Gondwanaland.

The Pan-African orogenic cycle figures prominently in recent hypotheses regarding the breakup and collisional rearrangement of a Neoproterozoic supercontinent (Moores 1991, Hoffman 1991, Dalziel 1991). Specifically, East and West Gondwanaland must have collided after supercontinent breakup, (Storey 1993). so the evolution of the EAO provides important constraints on these hypotheses.

The purpose of this paper is to review the sequence of tectonic events responsible for forming the EAO and thereby to demonstrate when Gondwanaland was born. Our knowledge of the EAO lags behind most other orogens of similar size and importance, and this is partly due to the fact that the orogen cuts across many developing African countries, where it

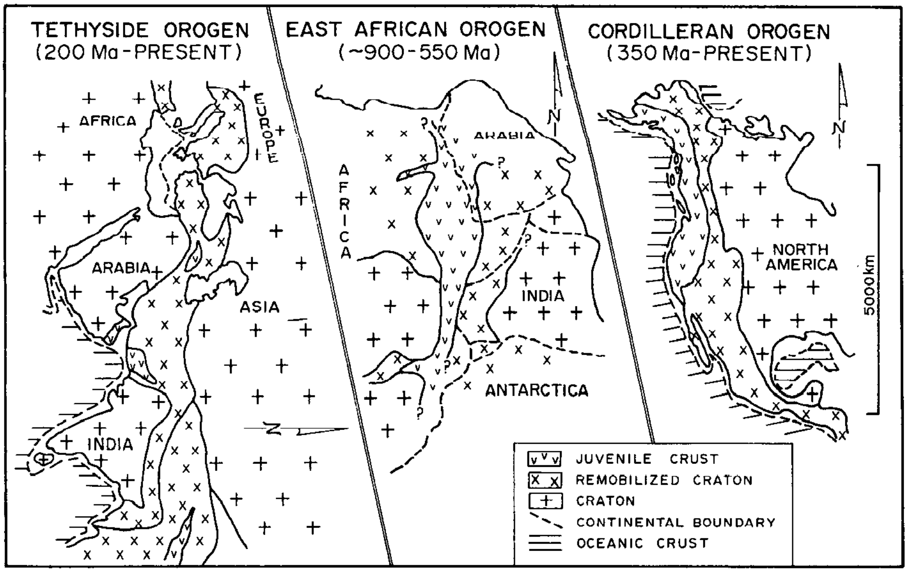


Figure 3 Comparison at the same scale of the size and style of the Neoproterozoic East African Orogen with the Phanerozoic Cordilleran (Oldow et al 1989) and Tethyside (Şengör 1987) orogens. For the East African Orogen, subdivision of juvenile, cratonic, and reworked cratonic crust is based on geochronologic and radiogenic isotopic data. For Phanerozoic orogens, distinction of craton from reworked craton is based on topography, while subdivision of reworked craton from juvenile crust is based on isotopic data ("0.706 line") in the Cordilleran and interpretation of regional geologic maps (Şengör 1987). Note that the East African Orogen is like the Cordilleran Orogen in the abundance of juvenile crust and like the Tethyside Orogen in manifesting the effects of continent-continent collsion.

may be difficult to work, or where only limited resources can be spent by national geologic surveys on academic pursuits, no matter how important. This situation is unlikely to improve in the near future. Nonetheless, how and when Gondwanaland was assembled has important implications for the evolution of the Neoproterozoic environment and Precambrian tectonics, and a summary at this time serves to emphasize problems that could reward further work with fundamental insights. To enliven the tale, this review is heavily spiced with interpretation, but my intent is to present the EAO so that both the constraints that we have and the problems that persist are apparent. Because so many of the fundamental questions begin with "When?," the story is told from beginning to end of the EAO orogenic cycle: the beginning, arc evolution and consolidation, continental collision, and escape tectonics. Towards the end, two of the more pressing unresolved problems are outlined: Where does the EAO continue to the south (or does it)?; and what happened to the EAO forelands in the north? Finally, the implications of EAO evolution for Neoproterozoic continental reconstructions are explored.

THE BEGINNING

The main question is, "Does the EAO manifest a Wilson cycle orogeny?"; that is, did it begin with rifting of an earlier continental assembly? Direct evidence for rifting may be preserved in sedimentary successions in the EAO that have been interpreted as passive margin deposits in Kenya (Vearncombe 1983, Key et al 1989, Mosley 1993) and in Sudan (Kröner et al 1987). These sequences are difficult to date. Metasedimentary rocks from Sudan were metamorphosed to granulite facies at about 720 Ma, so deposition must be earlier. A thick succession of multiply-deformed slope- and basinal-facies carbonates is found in northern Sudan along the line of the Keraf zone, and these are most readily interpreted as remnants of a passive margin sequence (Schandelmeier et al 1993). These rocks are also undated, but regional considerations indicate deposition prior to 750 Ma. Arguments for the existence of a passive margin are clearest for Kenya. Key et al (1989) argued that the Kenyan sediments were deposited during a period of crustal extension sometime between 840 and 770 Ma. These sediments have been isoclinally folded, thrust, and metamorphosed to amphibolite facies, so that their stratigraphic relationships and original dimensions are poorly known. Nevertheless, the metasediments appear to become finer as one moves eastward, and characteristic lithologies include quartzites, semipelitic gneiss, and carbonates. These rocks often preserve primary sedimentary structures, including ripple marks and cross-bedding (Mosley 1993). Along strike to the south in Tanzania, Prochaska & Pohl

(1983) interpreted Neoproterozoic amphibolite-facies mafic and ultramafic rocks as having formed during early rifting, an interpretation that also applies to ca. 790 Ma mafic-ultramafic complexes with negative initial $\varepsilon Nd(t)$ in Madagascar (Guerrot et al 1993). [Note: $\varepsilon Nd(t)$ refers to the difference in the $^{143}Nd/^{144}Nd$ of a rock from that of the bulk earth, in parts per 10,000, at the time of formation t. For further discussion, see DePaolo (1988).] Vearncombe (1983) recognized facies in western Kenya indicating eastward transitions from continental margin to shelf and finally deep sea conditions. Complications due to metamorphism and orogeny and incomplete exposure along strike make it difficult to decide whether all of the metasediments preserved in the Kenyan sector of the EAO were once laterally equivalent or whether they have been juxtaposed along an intervening suture. Shackleton (1986) argued that the Kenyan ophiolites are rooted on the east side of the EAO. If this is true, then the telescoped Kenyan metasedimentary basin is about 400 km wide, consistent with it being interpreted as a passive margin.

Two possible aulacogen-like structures (rifts extending inland from the continental margins) are found along and west of the EAO. In south-central Africa, the E-W trending Zambezi belt (Figure 2) has been interpreted as an intensely deformed aulacogen exposed at mid-crustal levels (Hanson et al 1989). The Zambezi belt contains extensive tracts of remobilized sialic basement structurally overlain by thick supracrustal rocks. Bimodal volcanics (U-Pb zircon age of 879 ± 19 Ma; Wilson et al 1993, Hansen et al 1993) inferred to record initial extension-related magmatism are structurally overlain by clastic sediments and then shallow marine carbonate rocks. This sequence is interpreted by Hanson et al (1989, 1993) to represent extension and thermal subsidence. It was metamorphosed to amphibolite facies and penetratively deformed at about 820 Ma (Hanson et al 1988b). Crust and structural trends of the Mesoproterozoic Irumide belt are correlated across the Zambezi belt, apparently precluding a large Pan-African ocean basin (Hanson et al 1988a), and supporting the interpretation of an aulacogen [although many workers disagree with this conclusion, such as Burke & Dewey (1972)]. If so, the age of 880 Ma for rift-related volcanism implies a similar age of rifting in the EAO. In Sudan, the recently discovered Atmur-Delgo suture (Figure 2; Schandelmeier et al 1993) appears to represent the third arm of a RRR triple junction, floored by oceanic crust. The Sm-Nd age of about 750 Ma obtained from the ophiolite (Denkler et al 1993) is a minimum age for the rifting event in northeast Africa.

Finally, consideration of the oldest Pan-African igneous rocks in the Arabian-Nubian Shield suggests that rift-related igneous activity began about 880 Ma in the northern part of the EAO. There, the oldest Pan-

African igneous rocks are a bimodal sequence of moderately high-titanium tholeiitic basalts and subordinate rhyodacites of the Baish Group, interpreted by Reischmann et al (1984) to have been erupted in an intra-oceanic arc setting. These, and similar volcanics in Sudan, yield single zircon evaporation ages of 840–870 Ma (Kröner et al 1991, 1992a) and can be interpreted as manifesting early stages of rifting in the EAO, for the following reasons: 1. They are compositionally similar to other rift-related volcanic sequences in the Arabian-Nubian Shield, such as the Arbaat Group (Abdelsalam & Stern 1993a); 2. volcanic sequences that form during continental breakup show many of the chemical characteristics of arc lavas (e.g. depletion in Nb) and typically plot in the field of arc lavas on discriminant diagrams (Wang & Glover 1992); and 3. these volcanics occur in the Haya terrane in Sudan and in the Asir terrane in Arabia (Figure 4), adjacent to older crust of the Afif terrane and interbedded with pelitic and quartzitic rocks. The T_{DM} age (crust formation Nd model ages; Nelson & DePaolo 1985) for one rhyolite from a similar sequence in Sudan is 1.3 Ga, indicating a significant contribution from older crust, plausibly incorporated as a result of contamination during the initial stages of rifting. If this interpretation is correct, then rifting in the northern part of the EAO occurred about 840–870 Ma, and the Asir-Haya terranes traveled as the conjugate margin of the eastern continental fragment.

In summary, an argument can be made that the EAO began by rifting of a continent, beginning about 870 Ma. Rifting led to the development of a passive margin and perhaps two aulacogen-like structures, one of which evolved into a narrow ocean basin. The passive margin is best preserved in the west, where it nonetheless has been multiply deformed and metamorphosed, and is poorly developed in the east. This may be due to poor exposure and more intense deformation and metamorphism, or it may be due to the fact that the eastern margin of the EAO developed later as an Andean-type convergent margin.

The hypothesis that the EAO began by rifting, with the implication that a larger continental block was broken up into West Gondwanaland and, perhaps, East Gondwanaland about 870 Ma, is consistent with global considerations. Hoffman (1989, 1991) marshaled evidence that a supercontinent (Rodinia) formed by about 1.1 Ga. Application of the model of supercontinent cyclicity (Gurnis 1988, Veevers 1990) suggests that Rodinia should begin to break up after about 150–200 Ma. This inference is supported from the subsidence histories of some Neoproterozoic basins in Australia, which preserve evidence for a regional rifting event at 900 Ma (Lindsay et al 1987). Evidence that this was followed by a major episode of sea-floor spreading is contained in the EAO ophiolite record (discussed later) and supported by the global record of $^{87}Sr/^{86}Sr$ in carbonates, which

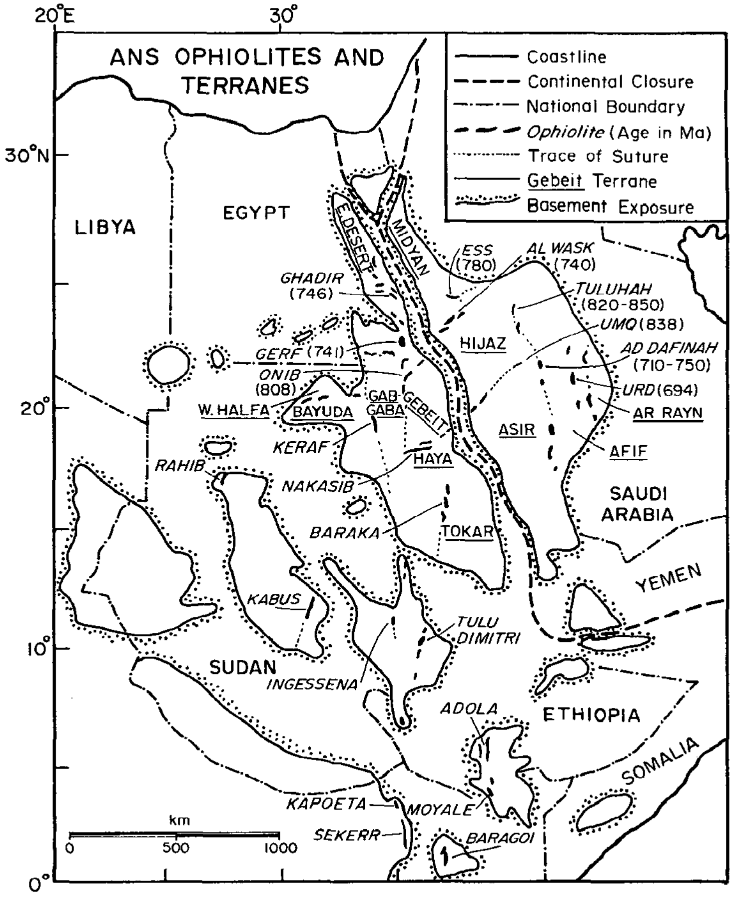


Figure 4 Distribution of ophiolites and terranes in the Arabian-Nubian Shield, northern EAO. Not all of the ophiolites are shown, but the ones that are shown are drawn as close to scale as possible. Ages (in Ma) are given in parentheses. Dotted lines show the inferred extensions of sutures. Terrane names are underlined.

reaches a minimum—indicating that the seafloor hydrothermal flux dominated over continental runoff—sometime around 800 and 900 Ma (Figure 5).

OPHIOLITES, ARCS, AND THE SIZE OF THE MOZAMBIQUE OCEAN

While older ophiolites are known from elsewhere on Earth, the earliest record of abundant ophiolites is preserved in the Arabian-Nubian Shield

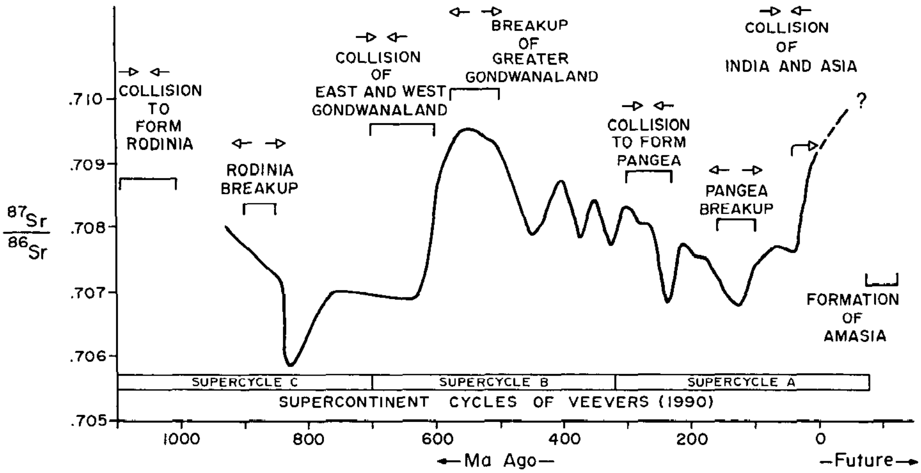


Figure 5 Generalized variation in the isotopic composition of Sr in the oceans through time, based on studies of $^{87}Sr/^{86}Sr$ in carbonate rocks (Burke et al 1982, Veizer et al 1983, Asmerom et al 1991). Temporal and compositional resolution decrease with increasing age. Nonradiogenic Sr reflects high inputs of seafloor hydrothermal Sr, whereas radiogenic Sr reflects high riverine runoff from old continental crust. Supercontinent cycles of Veevers (1990) and approximate times of formation and dispersal of supercontinents are shown; note that the breakup of "Greater Gondwanaland" continued throughout the Paleozoic. "Amasia" refers to the supercontinent (America plus Asia) expected to form from closure of the Pacific. A high flux from seafloor hydrothermal activity is indicated at about 800 Ma. The rise in seawater $^{87}Sr/^{86}Sr$ did not begin until about 600 Ma.

(Figure 4). These occur as variably dismembered nappes associated with suturing either between arc terranes or between the juvenile crust of the ANS and the palimpsest older crust to the west. Where dated, these range in age from about 880 to 690 Ma (Figure 6). Unequivocal ophiolites have not been reported from south of Kenya (Price 1984, Ries et al 1992). Shanti & Roobol (1979) first recognized ophiolitic nappes in the ANS, and since that time nearly all significant ophiolitic occurences have been interpreted as nappes (Church 1988, Abdelsalam & Stern 1993b). ANS ophiolites demonstrate two important aspects of EAO evolution: (*a*) that oceanic crust formed and was partly preserved; and (*b*) that ophiolitic and other nappes traveled as much as 150 km or more (Stern et al 1990), indicating that horizontal transport was an important aspect of crustal shortening.

ANS ophiolites are similar to many Phanerozoic ophiolites in having a strong "supra-subduction zone" (SSZ; Pearce et al 1984) chemical signature. These characteristics—along with the abundance of immature and volcaniclastic sediments deposited on the ophiolites—indicate ANS ophi-

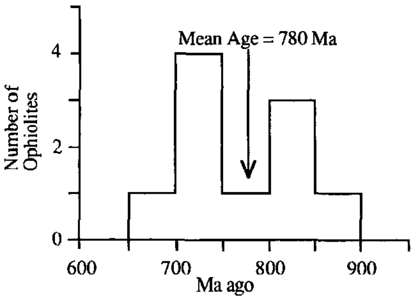


Figure 6 Histogram of robust ages for 10 ophiolites from Egypt and Sudan (Kröner et al 1992b) and Saudi Arabia (Claesson et al 1984, Pallister et al 1988).

olites for the most part formed in arc, back-arc, or fore-arc settings (Price 1984, Pallister et al 1988, Berhe 1990, Abdel Rahman 1993). Judging from the chemistry of the ophiolites and the associated sediments, we have no ophiolites that preserve the seafloor of the "Mozambique Ocean" (Dalziel 1991)—the greater ocean that may have existed between East and West Gondwanaland, perhaps 700–850 Ma. This is not surprising, because Phanerozoic ophiolites usually come from fringing SSZ environments and only rarely from the main ocean basin (Pearce et al 1984). This means that we have no *direct* information about the size, age, structure, or even the existence of the Mozambique Ocean. However, two lines of evidence indicate that the Mozambique Ocean must have been a large basin: The first comes from consideration of the implications of extensional stress regimes at convergent margins; the second concerns the significance of crustal growth rates.

Studies of stress regimes associated with modern convergent margins indicate that arcs under extension—those most likely to spawn back-arc basins—are typically those that are subducting old seafloor. This is because an important driving force for plate motion is the excess density of subducted lithosphere (Davies & Richards 1992). Lithospheric age and density correlate strongly; consequently the stresses in the over-riding plate vary from strong compression to strong extension with age of the subducting plate. Molnar & Atwater (1978) found that all modern Pacific arcs that rifted to form back-arc basins subduct seafloor older than 100 Ma. This result can be used to infer from the abundance of back-arc basin ophiolites in the ANS that the Mozambique Ocean was likely to have contained seafloor that was on the order of 100 My or older when it was subducted. Alternatively, some ANS SSZ ophiolites may have formed in fore-arcs, an interpretation which is supported by the local occurrence of boninites (Al-Shanti & Gass 1983, Reischmann 1986, Berhe 1990, Abdel Rahman 1993), in which case these ophiolites are best interpreted as forming during subduction zone initiation (Stern & Bloomer 1992). This also requires dense, old lithosphere within the plate to be digested. In either case, the inference from abundant SSZ ophiolites indicates strong extension in the associated convergent margin, suggesting that old seafloor was subducted. Old seafloor requires ocean basins that are at least as old. If seafloor spreading is continuous, age implies size. For example, given a moderate full spreading rate of 10 cm/yr, 100 Ma seafloor implies an ocean 10,000 km wide. That the Mozambique Ocean was a large ocean basin is the logical conclusion of this line of reasoning.

Calc-alkaline and tholeiitic plutons, lavas, pyroclastics, volcaniclastics, and associated immature sediments make up a major portion of the upper crust of the ANS. These are interpreted to comprise arc terranes, separated one from the other by ophiolite-decorated suture zones which mark the positions of fossil subduction zones (Figure 4; Vail 1985, Stoeser & Camp 1985). It has proven difficult to identify all of the components of individual arcs such as their accretionary prisms, fore-arc basins, or frontal arcs. The only well-documented accretionary prism is in the Al Amar suture of the eastern Arabian shield (Figure 2; Al-Shanti & Gass 1983). Blueschist occurences have not been documented, nor have any paired metamorphic belts been convincingly demonstrated. Nevertheless, the chemical and isotopic compositions, largely submarine nature of the igneous and immature sedimentary rocks of the ANS, and their occurrence in ophiolite-bounded terranes are most consistent with their origin in immature arcs [but see alternative explanations of Stern et al (1991) for one volcano-sedimentary belt].

Because the rock record indicates that processes of Neoproterozoic crustal growth were similar to modern plate tectonics, the rate of growth of Neoproterozoic continental crust may have been similar as well. Phanerozoic continental crust formation occured at convergent margins for the most part and is estimated to be about 1.1 km^3/yr (Reymer & Schubert 1984). Several efforts have been made to estimate growth rates over an estimated 300 Ma history for the ANS. These estimates range from about 20% to nearly 80% of the global Phanerozoic crustal growth rates (Figure 7). This means that, if the assumption of similar Neoproterozoic and Phanerozoic growth rates is correct, between 20% and 80% of all Neoproterozoic crustal growth occurred in the ANS. These estimates apply only to the Arabian-Nubian Shield, and the range largely reflects the uncertainties about how much of this crust is juvenile. A similar estimate can be made for the area mapped as juvenile crust in Figure 2, i.e. the area between the limits of isotopic influence of East and West

Gondwanaland. It is not generally recognized that much juvenile crust exists south of Sudan and Ethiopia, but initial $^{87}Sr/^{86}Sr$ ratios for rocks from the EAO in Kenya and Tanzania are moderately low, with a mean of about 0.7041 (Key et al 1989, Maboko et al 1985). The proportion of juvenile contributions is certainly higher in the north than in the southern EAO, but a mean of 80% juvenile contribution over the part of the EAO shown in the stippled pattern in Figure 2 is probably conservative. These considerations lead to an estimate of about 30% of the Phanerozoic growth rate (*I* in Figure 7).

The regions mapped as remobilized craton in Figure 2 contain a significant juvenile component. This is demonstrated by the discovery of an ophiolite belt striking WSW into the interior of Africa in northern Sudan (Schandelmeier et al 1993); by the presence of ca. 560 Ma batholiths with moderately low initial $^{87}Sr/^{86}Sr$ (0.705–0.706; Fullagar 1978) in the Tibesti massif, Libya, to the west; and by the presence of 700–850 Ma igneous rocks to the east, with low initial $^{87}Sr/^{86}Sr$ (0.7025–0.7041; Gass et al 1990). In many places as much as two thirds of the crust mapped as remobilized craton may be Neoproterozoic juvenile contributions; nevertheless, the crustal growth rate used here is based on a conservative 30% juvenile addition. This estimate, plus that previously calculated, suggests that about

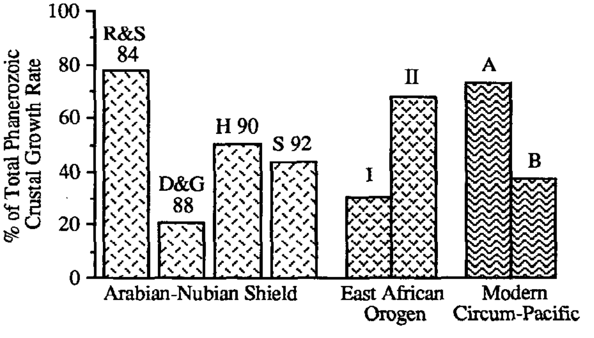


Figure 7 Estimates for crustal growth rates in the Arabian-Nubian Shield indicating that a large proportion of Neoproterozoic crutal growth occured in the northern East African Orogen [I = 80% of volume beneath "Juvenile Crust" in Figure 2 (35 km thick crust − 6 km thick oceanic crust + 6 km erosion over 300 My); II = I plus 30% of volume beneath area marked "Remobilized Craton" in Figure 2], and Modern Circum-Pacific [A = fraction of world's present convergent margins (Forsyth & Uyeda 1975); B = one half of A, estimated mean for an opening and then closing Pacific-type ocean]. The present growth rate of the continental crust is estimated to be 1.1 km^3/yr (Reymer & Schubert 1984). R&S 84 = Reymer & Schubert 1984, D&G 88 = Dixon & Golombek 1988, H 90 = Harris et al 1990, S 92 = Sultan et al 1992a.

65% of the Phanerozoic growth rate was responsible for the formation of the region affected by the Pan African in Figure 3 (*II* in Figure 7). This estimate is similar to the estimated growth rate of the modern circum-Pacific (*A* in Figure 7), and is considerably greater than the mean crustal growth expected for opening (no crustal growth) and closing of a Pacific-sized ocean basin (*B* in Figure 7). Insofar as total crustal growth is a function of the volume of seafloor subducted, the implication is that the Mozambique Ocean was large, perhaps comparable in size to that of the modern Pacific. These results also indicate that a disproportionately large fraction of Neoproterozoic juvenile crust formation occurred in the EAO, similar to the present situation in the western Pacific.

Recent crustal refraction studies indicate that ANS crust is composed of mafic lower and felsic upper crust, separated by a mid-crustal Conrad discontinuity (Gettings et al 1986, El-Isa et al 1987, Rihm et al 1991). Geochemical studies of xenoliths confirm that the lower crust is mafic (Table 1) and records equilibration pressures up to 12 kbar and temperatures of 830° to 1200°C (McGuire & Stern 1993), but have not resolved whether the mafic crust is residual or cumulate. McGuire & Stern (1993) interpret the isotopic and chemical data to indicate that the mafic lower crust formed during the Neoproterozoic, an argument that is supported by isotopic studies of spinel peridotites indicating that the mantle lithosphere beneath Arabia formed about 700 Ma (Henjes-Kunst et al 1990). The ANS lower crust preserves little evidence for the involvement of pre-existing crustal materials, and the absence of metasedimentary protoliths is noteworthy; it must be juvenile. It appears that mafic lower crust is a characteristic of the ANS and contrasts with the deeper crust recorded to the south and west, where pre-Pan African crust was involved and metasedimentary protoliths are found. In at least two of these localities [Sabaloka ("f" in Figure 2; Kröner et al 1987), and northeast Tanzania (Shackleton 1993)], however, there is evidence of mafic intrusion into the lower crust during the Neoproterozoic, and similar inferences are made from refraction studies in Kenya (KRISP Working Party 1991).

Estimates of the composition of the ANS upper, lower, and bulk continental crust are listed in Table 1. The ANS upper crust has a composition similar to that of the upper continental crust estimated by Taylor & McClennan (1985). The "Bulk ANS Crust" approximates a Mg-enriched andesite and is very similar to the estimate of the "Bulk Continental Crust" of Taylor & McClennan (1985).

Formation and coalescence of ANS arc terranes must have predated continental collision. Terrane accretion was completed by about 700 Ma (Ayalew et al 1990, Pallister et al 1988, Kröner et al 1992b, Stern & Kröner 1993), except along the Urd and Al Amar sutures in the eastern part of

Table 1 Composition of the ANS crust

Oxide	Mean ANS UC[a]	WAALC[b]	Bulk crust ANS[c]	Bulk continental crust[d]
SiO_2	66.4	50.6	58.2	57.3
TiO_2	0.63	0.90	0.77	0.90
Al_2O_3	14.9	16.4	15.6	15.9
FeO*	4.39	9.87	7.23	9.1
MgO	3.31	8.04	5.76	5.3
CaO	4.20	10.77	7.60	7.4
Na_2O	4.21	3.03	3.60	3.1
K_2O	1.86	0.32	1.06	1.1
P_2O_5	0.15	0.13	0.14	—

[a] Mean Upper Crust of the Arabian-Nubian Shield, estimated using 32,634 points of mapped basement lithologies in the Eastern Desert of Egypt (Stern 1979) and mean compositions for various lithologies reported in Stern (1979, 1981), Abdel-Rahman (1993), Greenberg (1981), and Neary et al (1976).
[b] Weighted Average Arabian Lower Crust (McGuire & Stern 1993).
[c] 48.2% mean ANS UC plus 51.8% WAALC, relative masses estimated from crustal structure and densities of Gettings et al (1986).
[d] Bulk continental crust composition (Taylor & McClennan 1985).

the Arabian shield (Figure 2) where suturing occurred between 640 and 680 Ma (Stacey et al 1984). Also, there is field evidence that the Keraf structure in northern Sudan (Figure 4) is younger than the NE-SW sutures of the ANS (Schandelmeier et al 1993). Thus, suturing of the terranes within the ANS from 700 to 760 Ma was followed by suturing of the ANS composite terrane between East and West Gondwanaland, about 640 to 680 Ma.

THE COLLISION OF EAST AND WEST GONDWANALAND

The EAO in Kenya and Tanzania has long been recognized as manifesting an episode of Tibetan-style continental collision and crustal thickening (Burke & Dewey 1972). The most direct evidence for the timing of collision between the continental blocks of East and West Gondwanaland comes from the age of granulites exposed in the EAO. Insofar as their exposure reveals 15–45 km of uplift and erosion, EAO granulites show where the greatest thickening occurred and lead directly to inferences about where continent-continent collision was most intense. Granulites are not found north of central Sudan and southern Ethiopia but are common in southern Kenya, Tanzania, Malawi, and Mozambique.

The lower crust of the southern EAO contrasts with that to the north in two fundamental ways. First, whereas that to the north is largely intact and lies preserved 20 or more km deep in the present-day crust, that of the southern EAO crops out as tectonic slices. We do not know what composes the present lower crust of the southern EAO, although in Kenya the crust east of the Tanzanian craton and the East African rift is about 35 km thick, has a well-defined Conrad discontinuity, and contains seismic suggestions of mafic underplating (KRISP Working Party 1991). Second, whereas the lower crust of the northern EAO is mafic igneous rock, southern EAO granulites include a variety of igneous and metasedimentary rocks, including marbles and quartzites. These differences indicate the different histories of the northern EAO, dominated by igneous processes and relatively mild terrane accretion events, and that of the southern EAO, the focus of continental collision between East and West Gondwanaland.

Two granulite occurrences in Sudan have been dated with U-Pb zircon techniques. At Sabaloka (locality 1, Figure 2) meta-igneous and metasedimentary protoliths were metamorphosed to granulite facies at about 710 Ma (Kröner et al 1987). The metasedimentary granulites retain isotopic evidence of an Archean to mid-Proterozoic provenance, while the metamafic granulites may manifest igneous underplating similar to that of the ANS. Igneous bodies of charnockite, enderbite, and granite were intruded at 740 Ma at Jebel Moya (Stern & Dawoud 1991; locality 2, Figure 2). The "Samburuan" granulite-facies thrusting and metamorphism of Kenya is interpreted as manifesting plate collision across the N-S orogenic strike and is given an age of ~820 Ma based on an Rb-Sr whole-rock errorchron, but this is probably partially reset (Key et al 1989). (Note: An errorchron is an alignment of data points on an isochron diagram where the correspondence of the data to a least-squares fit is worse than that expected from analytical uncertainty alone.) The best known EAO granulites occur in the Eastern Granulite Complexes of Tanzania, a 900 km long discontinuous belt of metasedimentary and meta-igneous granulites, eclogites, metapyroxenites, meta-anorthosites, and ultramafics. These complexes display a complicated structural history, but the youngest foliation is subhorizontal and is interpreted as resulting from the development of huge, deep crustal nappes accompanying crustal thickening due to collision (Malisa & Muhongo 1990, Shackleton 1993). U-Pb zircon ages from the eastern granulites of Tanzania are interpreted to approximate the time of granulite-facies metamorphism and range from 650 to 715 Ma (Maboko et al 1985, Coolen et al 1982). This metamorphism occurred at 8 to 13 kbar and 700 to 900°C (Coolen 1982, Maboko et al 1989, Appel et al 1993). It is possible that metamorphism was related to the intrusion of basalts deep in the crust and that some of the older granulites formed prior to collision (Appel et al 1993). Uplift of the granulite terranes may better bracket the time of collision. This was accompanied by hydration and retrogression under amphibolite-facies conditions (ca. 475°–500°C), dated using hornblende K/Ar and $^{40}Ar/^{39}Ar$ techniques at about 630 Ma (Andriessen et al 1985, Maboko et al 1989). The only thermochronometric study of the EAO, for the Uluguru complex, Tanzania, has been presented by Maboko et al (1989). Their study indicates uplift (ca. 0.16 mm/a) from granulite formation until closure of hornblende to Ar diffusion (ca. 475°C) at 630 ± 10 Ma. Uplift then slowed to 0.05 to 0.08 mm/yr until muscovite closure (ca. 300°C) at 490 ± 10 Ma, with continued slow cooling and uplift until K-feldspar closure (ca. 170–200°C), at 420–450 Ma. These results indicate that, after a tectonic crustal thickening event and granulite-facies metamorphism, cooling followed an exponentially decaying path that is consistent with 20 to 40 km of isostatically-driven unroofing. There are two other granulite belts in Tanzania: the central and western granulite complexes. These have not been dated using U-Pb zircon techniques, and interpretation of Rb-Sr whole-rock ages for these bodies is problematic.

Younger granulites are found in the eastern part of the EAO. In southern Ethiopia (locality 3 in Figure 2), Gichile (1992) describes mafic granulites which indicate temperatures of 670–900°C and pressures of 9 ± 1.5 kbar. Similar charnockites from the nearby Bergudda complex yield a U-Pb zircon age of 545 Ma (Ayalew & Gichile 1990). The granulites of southernmost India and Sri Lanka (localities 10 and 11, Figure 2) were once thought to be Archean and Paleoproterozoic, but have since been shown to be Neoproterozoic, forming at 660 to 550 Ma (Burton and O'Nions 1990, Baur et al 1991, Choudhary et al 1992). Zircon evaporation ages for granulites from southeast Madagascar indicate granulite-facies metamorphism at 570–580 Ma (Paquette et al 1993). These younger ages may either indicate that a second, younger continent-continent collision occurred in the eastern part of the EAO or that crustal thickening progressed from west to east across the collision zone.

Arguments for a younger age of collision between East and West Gondwanaland have been advanced. For example, Kröner (1993) argued that the aforementioned granulites in southernmost India, Sri Lanka, and Madagascar indicated collision at the end of the Neoproterozoic. Powell et al (1993) concluded from paleomagnetic data that Gondwanaland formed no earlier than the end of the Neoproterozoic and perhaps as late as the mid-Cambrian. These arguments are also attractive insofoar as they are consistent with a rapid rise in seawater $^{87}Sr/^{86}Sr$ at the end of the Precambrian (Figure 5), but they are not consistent with the sequence of tectonic events observed along the EAO. Specifically, the consolidation of juvenile arcs was completed by about 700 Ma, and this should predate terminal collision by at most a few tens of millions of years. Also, it will be shown below that tectonic escape was underway by about 640 Ma, and this should not have commenced until terminal collision was underway.

WHERE DOES THE EAO CONTINUE TO THE SOUTH—OR DOES IT?

To the south, there are abundant granulitic rocks in Mozambique, Malawi, and Madagascar (Andreoli 1984). Rb-Sr geochronologic data indicate that metamorphism and deformation in the Malawi-Mozambique region occurred about 850 to 1100 Ma (Andreoli 1981, Costa et al 1992, Pinna et al 1993). Andreoli (1981) also reported U-Pb zircon ages of 840 and 920 Ma, but the common occurrence of cores indicates that inheritance is a likely problem and these ages may provide only a maximum age constraint. K-Ar results indicate that this region experienced thermal resetting ~450 to 650 Ma (Cahen et al 1984).

If the basement of Mozambique and Malawi last experienced major deformation and metamorphism about 1 Ga, and the interpretation that the EAO manifests closure of a major ocean basin is correct, then the EAO must continue either to the southwest through the Zambezi belt or southeast through Madagascar and India. Some (e.g. Hoffman 1991, Unrug 1993) argue that the Zambezi and Damaran structures comprise a suture between East and West Gondwanaland, but I question this hypothesis for 3 reasons: 1. The Irumide orogen can be traced through and across the Zambezi belt, indicating that continental separation either did not occur or was minor (Hanson et al 1988b); 2. isotopic data for the Damaran belt indicate that little juvenile crust was produced, suggesting that it represents a collapsed intracontinental basin (Hawkesworth et al 1986); and 3. tectonism was diachronous, with rifting beginning at 870 Ma in the Zambezi belt (Wilson et al 1993, Hanson et al 1993) and at about 750 Ma in the Damaran (Miller & Burger 1983), and thrusting beginning at about 820 Ma in the Zambezi belt (Hanson et al 1988b), continuing until about 550 Ma in the Damaran belt (Hawkesworth et al 1986).

The EAO could extend to the southeast, south of the multiply-deformed rocks of Madagascar, but no likely candidate for a suture has been identified. Similarities in basement geology between South Africa and Antarctica led Groenewald et al (1991) to conclude that these were a single entity by 1 Ga. Hoffman (1991) made an intriguing suggestion: The Mozambique Ocean closed about a pivot in what is now South Africa. This would result in the creation and subduction of large areas of oceanic crust in the north but vanishingly little in the south. This model is supported by the apparent absence of ophiolites south of the equator, where the preserved mafic-ultramafic complexes manifest rifts which did not open to the dimensions of large ocean basins (Prochaska & Pohl 1983, Guerrot et al 1993).

The problem of the southern extension of the EAO is a first-order unresolved question. For now, I prefer to interpret the EAO as continuing south from Tanzania into Mozambique and then south along the present Antarctica-Africa margin. The suture in Mozambique thus may be cryptic, with its trace plausibly defined by the N-S trending Cobué-Geci-Nkalapa-Luatize thrusts and shear zones (Pinna et al 1993). The following arguments support this view:

1. Granulites from Tanzania indicate that a major suture exists, and the strike of this suture appears to continue either southwest into the Zambezi belt or to the south into Mozambique. The continuation into the Zambezi belt was rejected above, leaving only the latter hypothesis.
2. Structural trends can be traced from the southern part of the Zambezi belt southeast and then south into the western deformation front of Mozambique (Vail 1965), suggesting contemporaneity of these two deformations. Deformation in the Zambezi belt is dated at 820 Ma, thus that in the western deformation front should be similar in age.
3. Structures at the deformation front, between the Kalahari Craton and the Mozambique belt south of the Zambezi belt, are remarkably similar to those described along the deformation front in Tanzania and Kenya (compare Johnson 1968 with Hepworth & Kennerly 1970 and Sanders 1965).

WHAT HAPPENED TO THE FORELANDS OF THE ARABIAN-NUBIAN SHIELD?

Another important unresolved set of questions concerns the fate of the forelands to the ANS, both to the east in Arabia and to the west in North Africa. In both instances we have a poor understanding of the tectonic setting in which extensive crustal remobilization occurred, and thick Phanerozoic cover complicates the problem for Arabia. In North Africa, there is geochronologic and isotopic evidence that extensive tracts of continental crust existed in Mesoproterozoic and earlier times (Klerkx & Deutsch 1977, Harms et al 1990, Sultan et al 1992b). This tract has been subjected to intense remobilization, including widespread intrusion and anatexis, deformation, metamorphism, uplift, and erosion, to the extent that Black & Liegois (1993) call this the "Central Saharan Ghost Craton." It now extends from the eastern Hoggar to the Nile and from the Mediterranean to the Congo craton, covering nearly half of Africa; it is simultaneously an excellent example of cratonic reactivation and one of the most poorly understood among all tracts of continental crust. The situation for Arabia is grossly similar; a large tract of remobilized Mesoproterozoic and older crust exists in the eastern Arabian shield and Yemen (Figure 2), but juvenile Neoproterozoic crust occurs in several localities to the east in Oman (Gass et al 1990).

The timing and cause of cratonic remobilization are presently unresolved. Geologic studies in the region of interest in North Africa are politically and logistically difficult, and this is compounded by the fact that large amounts of inheritance make it difficult to interpret what geochronologic data do exist. For example, we do not know if this ghost craton collided with (Pin & Poidevin 1987), or is a remobilized part of (Sultan et al 1992b), the Congo craton. There are three general classes of explanations for the remobilization: 1. convergent margin or collisional processes; 2. extensional processes; and 3. lithospheric delamination. According to the first class of explanations, westward-directed thrusting from the ANS led to crustal thickening and cratonic reworking (Schandelmeier et al 1988). A variant of this argues that at least one ocean basin and related arc batholiths formed in the interior of North Africa (Fullagar 1978). The second class of models has not been generally considered, but crustal extension can remobilize very large crustal tracts, such as the Basin-and-Range province of the western United States. Finally, regional Neoproterozoic lithospheric delamination has been advocated (Ashwal & Burke 1989, Black & Liegeois 1993). Inasmuch as crustal extension and lithospheric delamination happen most readily where the lithosphere and/or crust are warm and thickened, all three models are consistent with early thickening of the foreland. The presence of ophiolite belts that trend westward into the interior of North Africa (Stern et al 1990, Schandelmeier et al 1993) requires early extension and sea-floor spreading followed by compressional emplacement of ophiolitic nappes. This demonstrates that both crustal extension and collision occured in the foreland, although the timing remains contentious. These models need to be assessed; our understanding of EAO evolution will remain incomplete until we have a better understanding of the cause and timing of this cratonic remobilization.

RIGID INDENTORS AND ESCAPE TECTONICS

Collision along the EAO led to the development of strike-slip shear zones and faults and related extensional basins, similar to those of Cenozoic Asia. For the collision of India with Asia, Harrison et al (1992) estimate that one third of the convergence was accomodated by underthrusting and a slightly smaller amount was taken up by the lateral extrusion of Southeast Asia. If similar processes accompanied formation of the EAO (and an oceanic "free-face" existed), then the 20 to 40 km of uplift recorded in

EAO granulites must have been accompanied by a significant episode of "escape tectonics." Several variants of this have been proposed for the EAO. The first was proposed by Schmidt et al (1979) to explain the Najd fault system of Arabia (Figure 2) as the result of a continental "rigid indentor" to the east being subducted beneath the ANS to the west. The specific predictions of this model fail, as pointed out by Stern (1985). Burke & Şengör (1986) developed the concept of tectonic escape and applied it to the entire EAO, implying that much of the along-strike dichotomy resulted from the extrusion of trapped juvenile terranes from in front of the "hard" collision between East and West Gondwanaland towards an oceanic free-face in the north. Berhe (1990) noted that the entire EAO was affected by NW-SE trending strike-slip faults and suggested that these resulted from an oblique collision. Bonavia & Chorowicz (1992) argued that the Tanzania craton acted as the rigid indentor, consistent with an interpretation of an east-dipping subduction zone beneath East Gondwanaland (Shackleton 1986).

The timing of EAO terminal collision is constrained by when tectonic escape began and how long it continued. These ages also yield insights into the kinematics of collision (for example, how long convergence continued between East and West Gondwanaland). Faulting and related deformation along the Najd fault system of Saudi Arabia began about 630 Ma (Stacey & Agar 1985) and continued until about 530 Ma (Fleck et al 1976), indicating that collision was well underway prior to 630 Ma, although it is not clear how much earlier than this collision began. This uncertainty is partly due to our ignorance about whether or not (and if so, how long) there is a lag time between collision and escape tectonics. We are also uncertain when, in the escape history of the EAO, the Najd and other NW-trending sinistral shear systems formed. For example, in Kenya and Tanzania there are N-S trending "straightening zones" (Hepworth 1967) which fold the older nappes along isoclinal, vertical axes. This fabric constitutes the most obvious N-S fabric of the EAO south of Sudan. The straightening zones clearly post-date emplacement of the ophiolitic nappes and preceed development of the NW-SE shear zones, but it is not clear whether straightening zones demonstrate continued shortening of the nappe stack or reveal the first stages in tectonic escape. In Ethiopia and Sudan, these zones are interpreted to manifest E-W shortening with little strike-slip shearing (Beraki et al 1989, Miller & Dixon 1992). In Sudan, Abdelsalam (1994) documents the evolution of a N-S zone of pure strain into a NW-SE oriented strike-slip fault and interprets this as the result of progressive E-W shortening. In Kenya and in Arabia, similar structures are interpreted as sinistral shear zones (Key et al 1989, Duncan et al 1990, Mosley 1993).

Regardless of whether the straightening zones indicate pure or simple shear, their ages provide an independant younger limit to the time of collision. In Sudan, deformation along the Hamisana shear zone has been bracketed between 610 and 660 Ma (Stern & Kröner 1993). In Kenya, two phases of post-accretionary deformation are found that predate development of the NW-trending strike-slip faults: the ca. 620 Ma Baragoian and the ca. 580 Ma Barsaloian events. Both Baragoian and Barsaloian structures are characterized by upright tight folds, but the former trend NW-SE to NNW-SSE and record sinistral displacements whereas the latter trend N-S and record dextral displacements. Both are interpreted as post-collisional ductile deformation parallel to the orogen (Key et al 1989).

Tectonic escape resulted in the formation of a large area of ca. 600 Ma old rifts (Figure 2), in NE Egypt (Stern 1985), in Oman (Wright et al 1990), and in the Persian Gulf (Husseini & Husseini 1990). This episode of collision-related extension led in turn to the rifting away of some fragments of greater Gondwanaland and the formation of a passive margin in North Africa and Arabia during latest Neoproterozoic time (Bond et al 1984). The supercontinent that resulted from collision between East and West Gondwanaland must have been significantly larger than the Gondwanaland that persisted throughout the Paleozoic, although we do not know how much larger or where the rifted fragments are now. Thus the discussion by Burke & Şengör (1986) regarding whether the Persian Gulf salt basins are related to oceanic closing or opening is moot: They are related to both.

A better understanding of the timing and style of EAO escape tectonics should be a goal of future research in the East African Orogen, but for the present synthesis the following points are accepted:

1. Collision between East and West Gondwanaland involved the Tanzanian craton as the rigid indentor and the western flank of East Gondwanaland as the region of crustal thickening and plastic deformation.
2. Escape was predominantly to the north.
3. Escape began sometime before 610 Ma and after 660 Ma and continued until about 530 Ma, implying that convergence between East and West Gondwanaland continued for 120 to 170 My after initial collision.
4. Tectonic escape led to the formation of rifted basins in northeast Africa and Arabia which, in turn, led to continental separation and the formation of a passive margin on the north flank of Gondwanaland at the very end of the Precambrian (Husseini 1989, Brookfield 1993). The size and location of the rifted fragments are unknown but may be found in central Europe, Turkey, and environs.

These relationships are shown in Figure 8, which compares the inferred tectonic relationships around the EAO with that of a mirror image of the modern India-Asia collision.

SIGNIFICANCE OF THE EAO FOR NEOPROTEROZOIC PALEOGEOGRAPHIC RECONSTRUCTIONS

Several models for the evolution and disposition of the continents during the Neoproterozoic have been advanced in the past few years (see review by Storey 1993). The above review of the EAO provides the following constraints for these models:

1. The formation of the Mozambique Ocean about 800–850 Ma reflects breakup of the Mesoproterozoic supercontinent Rodinia.
2. The abundance of ophiolites and volume of juvenile continental crust preserved in the northern EAO indicate that this rift evolved into a Pacific-sized "Mozambique Ocean" basin. The formation of this and other Neoproterozoic ocean basins was responsible for the excursion of the seawater Sr isotope curve to nonradiogenic values about 800 Ma.
3. The concentration of juvenile crust and ophiolites in the Arabian-Nubian Shield are consistent with Hoffman's (1991) model that the Mozambique Ocean closed like a fan, with a hinge in South Africa.
4. The deformation sequence—arc accretion → terminal collision → tectonic escape—indicates that the Mozambique Ocean closed to form Greater Gondwanaland around 640–700 Ma. The significance of the younger (ca. 550 Ma) granulite-facies metamorphic event of regional extent is not understood, but the absence of associated antecedent juvenile crust and ophiolites or subsequent escape tectonics implies that it does not date terminal collision along the EAO between East and West Gondwanaland.
5. Rifting at the Neoproterozoic-Cambrian boundary followed collision along the EAO and is related to initial disruption of Greater Gondwanaland. As Murphy & Nance (1991) emphasize, there are two episodes of supercontinent breakup during the Neoproterozoic, but both are protracted. The first is related to disaggregation of Rodinia; the formation of the Mozambique Ocean is related to this. Rifting of Laurentia away from Antarctica about 750 Ma (Storey 1993) may be the last stage of this episode. The second stage of rifting begins a few tens of millions of years after the formation of Greater Gondwanaland, beginning at the very end of the Neoproterozoic and continuing throughout the Paleozoic. This is the rifting episode identified by Bond et al (1984).

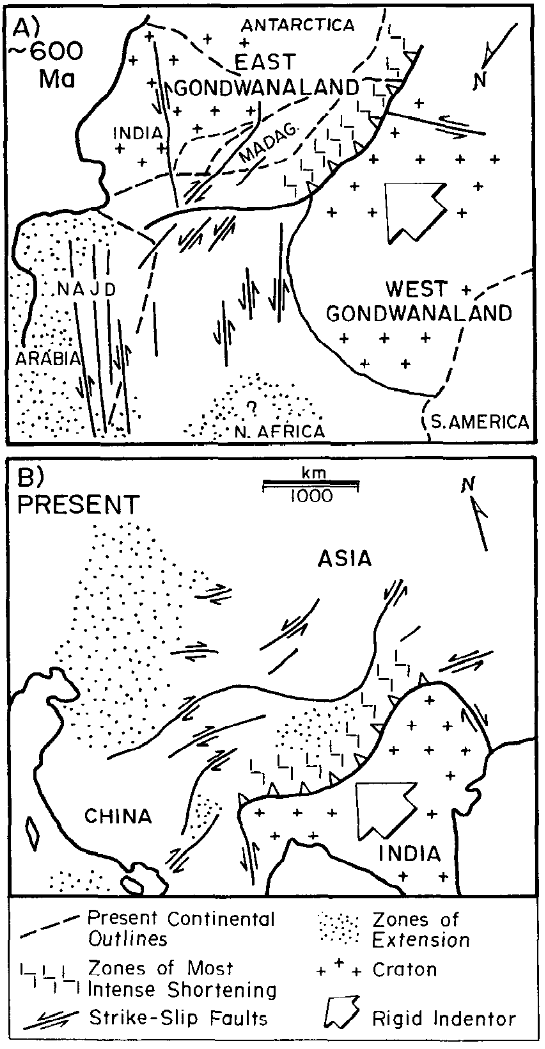


Figure 8 Comparison of continental collisions at the same scale. Both are oriented so that the rigid indentor is moving towards the upper left corner: (*A*) East African Orogen, ca 600 Ma ago. Areas without ornamentation are juvenile or were remobilized during the Neoproterozoic. (*B*) Modern India-Asia collision, shown as mirror-image so that the free face and principal zone of tectonic escape are on the same side of the rigid indentor as is the case for the EAO.

SUMMARY

Our understanding of the style and timing of collision between East and West Gondwanaland to form the East African Orogen is incomplete, but the general outline of this important event in Earth history is slowly emerging. A crude model (Figure 9) embodies present understanding of this evolution. Present reconstructions of the earliest stages in the EAO orogenic cycle may be more speculation than understanding, but initiation by rifting of a continent is simple, logical, and consistent with the data at hand. It is attractive to identify this rifting as part of the breakup of Rodinia, and we can place it in time about 850–900 Ma. In contrast, the sequence of events that begins with sea-floor spreading and formation of arcs and back-arc basins and continues with the accretion of these tectonic cells into juvenile crust is coming into sharper focus, at least for the

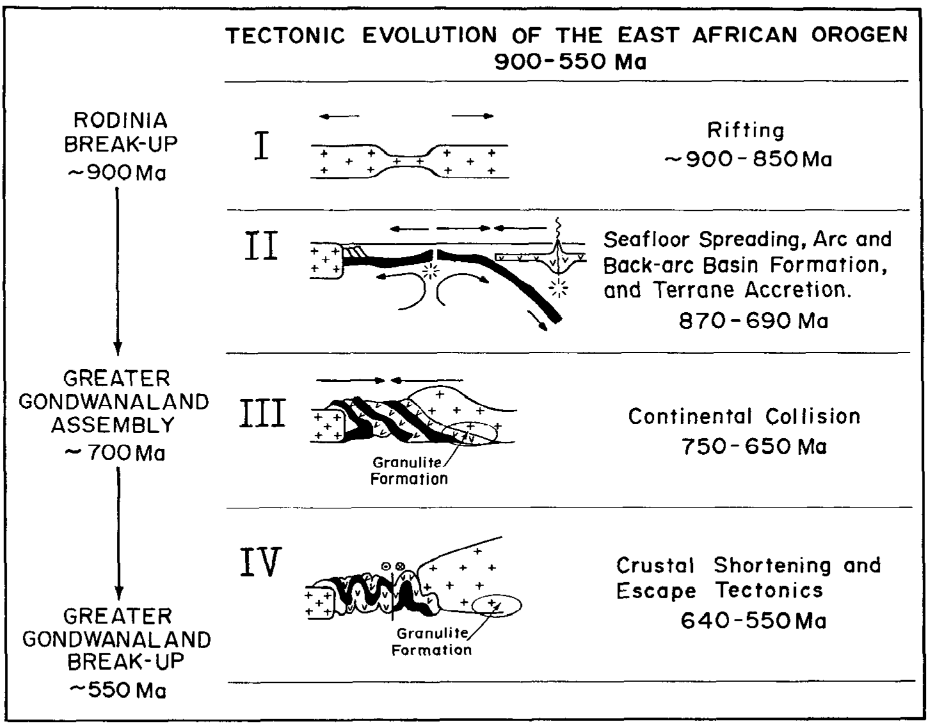


Figure 9 Two-dimensional summary of the tectonic evolution of the East African Orogen, as described in the text.

Arabian-Nubian Shield. There is good evidence that these processes were underway as early as 870 Ma and continued until at least 690 Ma. Assuming that modern crustal production rates are approximately valid for the Neoproterozoic as well, the volume of juvenile crust that formed during this interval suggests that a Pacific-sized ocean was opened and closed. Accretion of numerous juvenile arcs and a few older continental fragments continued during this interval. Continental collision first led to crustal thickening and uplift, beginning perhaps as early as 750 Ma but certainly by 700 Ma, and continued with orogenic collapse and escape tectonics until the end of the Precambrian. At least as far as the EAO is concerned, this dates the consolidation of East and West Gondwanaland, although several oceanic basins of unknown dimensions persisted within West Gondwanaland. Crustal thickening and uplift propagated with time eastward, now marked by zones of $\sim$550 Ma granulites in southernmost India, Sri Lanka, and Madagascar. Tectonic escape led to the development of major rift basins in the northern EAO and environs which led directly to sea-floor spreading and formation of a passive margin on the remnants of Gondwanaland and the formation of an ocean basin to the north, at about 550 Ma ago.

Acknowledgments

This manuscript benefits from the constructive criticisms of Kevin Burke, Tim Dixon, Richard Hanson, Alfred Kröner, Sospeter Muhongo, John Rogers, Heinz Schandelmeier, Robert Shackleton, Mohamed Sultan, Raphael Unrug, and John Vail. I especially appreciate Al Engel's early encouragement and the opportunity that he provided to begin studying the basement of northeast Africa. My work in northeast Africa is supported by NASA. This is UTD Programs in Geosciences contribution #751.

Literature Cited

Abdel Rahman ESM. 1993. Geochemical and geotectonic controls of the metallogenic evolution of selected ophiolite complexes from the Sudan. *Berliner Geowiss. Abh.* A: 145. 175 pp.

Abdelsalam MG. 1994. Structural evolution of the Late Proterozoic (Pan-African) Oko Shear Zone, Red Sea Hills, Sudan: implications for the evolution of post-accretionary structures in the Arabian-Nubian Shield. *J. Geol. Soc. London* Submitted

Abdelsalam MG, Stern RJ. 1993a. Tectonic evolution of the Nakasib suture, Red Sea Hills, Sudan: evidence for a Wilson Cycle in the Late Precambrian of the Arabian-Nubian Shield. *J. Geol. Soc. London* 150: 393–404

Abdelsalam MG, Stern RJ. 1993b. Structural styles of late Proterozoic sutures in the Arabian-Nubian Shield: the Nakasib suture, Red Sea Hills, Sudan. *J. Geol. Soc. London* In press

Al-Shanti AM, Gass IG. 1983. The Upper Proterozoic ophiolite mélange zones of the easternmost Arabian shield. *J. Geol. Soc. London* 140: 867–76

Andreoli MA. 1981. *The amphibolite and the granulite facies rocks of southern Malawi.* PhD thesis. Univ. Witwatersrand, S. Afr. 351 pp.

Andreoli MA. 1984. Petrochemistry, tectonic evolution and metasomatic mineralisations of Mozambique belt granulites from S. Malawi and Tete (Mozambique). *Precambrian Res.* 25: 161–86

Andreoli MA. 1991. Petrological markers in terrane analysis: massif anorthosites, high pressure granulites and scapolitized rocks from the Mozambique belt, southern Malawi. *UNESCO, Geol. Econ. Dev., Newslett.* 8: 43–55

Andriessen PAM, Coolen JJMMM, Hebeda EH. 1985. K-Ar hornblende dating of late Pan-African metamorphism in the Furua granulite complex of southern Tanzania. *Precambrian Res.* 30: 351–60

Appel P, Möller A, Schenk V, Muhongo S. 1993. Granulite facies metamorphism and P-T evolution in the Mozambique Belt of Tanzania. See Thorweihe & Schandelmeier 1993, pp. 171–75

Ashwal LD, Burke K. 1989. African lithospheric structure, volcanism, and topography. *Earth Planet. Sci. Lett.* 96: 8–14

Asmerom Y, Jacobsen SB, Knoll AH, Butterfield NJ, Swett K. 1991. Strontium isotopic variations of Neoproterozoic seawater: implications for crustal evolution. *Geochim. Cosmochim. Acta* 55: 2883–94

Ayalew T, Bell K, Moore JM, Parrish RR. 1990. U-Pb and Rb-Sr geochronology of the Western Ethiopian Shield. *Geol. Soc. Am. Bull.* 102: 1309–16

Ayalew T, Gichile S. 1990. Preliminary U-Pb ages from southern Ethiopia. In *Recent Data in African Earth Sciences*, ed. G Rocci, M Deschamps, pp. 127–30. *CIFEG Occ. Publ.* 22

Baur N, Kröner A, Liew TC, Todt W. Williams IS, Hofmann AW. 1991. U-Pb isotopic systematics of zircons from prograde and retrograde transition zones in high-grade orthogneisses, Sri Lanka. *J. Geol.* 99: 527–45

Beraki WH, Bonavia FF, Getachew T, Schmerold R, Tarekegn T. 1989. The Adola Fold and Thrust Belt, southern Ethiopia: a re-examination with implications for Pan-African evolution. *Geol. Mag.* 126: 647–57

Berhe SM. 1990. Ophiolites in Northeast and East Africa: implications for Proterozoic crustal growth. *J. Geol. Soc. London* 147: 41–57

Black B, Liegeois J-P. 1993. Cratons, mobile belts, alkaline rocks and continental lithospheric mantle: the Pan-African testimony. *J. Geol. Soc. London* 150: 89–98

Bonavia FF, Chorowicz J. 1992. Northward expulsion of the Pan-African of northeast Africa guided by a reentrant zone of the Tanzania craton. *Geology* 20: 1023–26

Bond GC, Nickeson PA, Kominz MA. 1984. Breakup of a supercontinent between 625 Ma and 555 Ma: new evidence and implications for continental histories. *Earth Planet. Sci. Lett.* 70: 325–45

Brookfield ME. 1993. The Himalayan passive margin from Precambrian to Cretaceous times. *Sediment. Geol.* 84: 1–35

Burke K, Dewey JF 1972. Orogeny in Africa. In *Orogeny in Africa*, ed. TFJ Dessauvagie, AJ Whiteman, pp. 583–608. Ibadan, Nigeria: Geol. Dept., Univ. Ibadan

Burke K, Sengör C. 1986. Tectonic escape in the evolution of the continental crust. In *Reflection Seismology: The Continental Crust*, ed. M Barazangi, L Brown, pp. 41–53. *Am. Geophys. Union, Geodyn. Ser.* 14 Washington, DC: Am. Geophys. Union

Burke WH, Denison RE, Hetherington EA, Koepnick RB, Nelson HF, Otto JB. 1982. Variation of seawater $^{87}Sr/^{86}Sr$ throughout Phanerozoic time. *Geology* 10: 516–19

Burton KW, O'Nions RK. 1990. Fe-Ti oxide chronometry: with application to granulite formation. *Geochim. Cosmochim. Acta* 54: 2593–602

Cahen L, Snelling NJ, Delhal J, Vail JR, Bonhomme M, Ledent D. 1984. *The Geochronology and Evolution of Africa.* Oxford: Clarendon. 512 pp.

Choudhary AK, Harris NBW, Van Calstern P, Hawkesworth CJ. 1992. Pan-African charnockite formation in Kerala, south India. *Geol. Mag.* 129: 257–67

Church WR. 1988. Ophiolites, sutures, and micro-plates of the Arabian-Nubian Shield. In *The Pan-African Belt of Northeast Africa and Adjacent Areas*, ed. S El-Gaby, RO Greiling, pp. 289–316. Braunschweig: Vieweg

Claesson S, Pallister JS, Tatsumoto M. 1984. Samarium-neodymium data on two late Proterozoic ophiolites of Saudi Arabia and implications for crustal and mantle evolution. *Contrib. Mineral. Petrol.* 85: 244–52

Conway Morris S. 1993. The fossil record and the early evolution of the Metazoa. *Nature* 361: 219–25

Coolen JJMMM. 1982. Carbonic fluid inclusions in granulites from Tanzania—a comparison of geobarometric methods based on fluid density and mineral chemistry. *Chem. Geol.* 37: 59–77

Coolen JJMMM, Priem HNA, Verdurmen EAT, Verschure RH. 1982. Possible zircon U-Pb evidence of Pan-African granulite facies metamorphism in the Mozambique belt of southern Tanzania. *Precambrian Res.* 17: 31–40

Costa M, Ferrara G, Sacchi R, Tonarini S. 1992. Rb/Sr dating of the Upper Proterozoic basement of Zambesia, Mozambique. *Geol. Rundsch.* 81: 487–500

Dalziel IWD. 1991. Pacific margins of Laurentia and East Antarctica–Australia as a conjugate rift pair: evidence and implication for an Eocambrian supercontinent. *Geology* 19: 598–601

Dalziel IWD. 1992. Antarctica: a tale of two supercontinents? *Annu. Rev. Earth Planet. Sci.* 20: 501–26

Davies GF. 1992. On the emergence of plate tectonics. *Geology* 20: 963–66

Davies GF, Richards MA. 1992. Mantle convection. *J. Geol.* 100: 151–206

Denkler T, Harms U, Franz G, Darbyshire DPF, Pilot J, Schandelmeier H. 1993. Evolution of the south-western part of the Late Proterozoic Atmur-Delgo suture zone (Northern Province, Sudan). See Thorweihe & Schandelmeier 1993, pp. 87–89

DePaolo DJ. 1988. *Neodymium Isotope Geochemistry*. Berlin: Springer-Verlag. 187 pp.

Derry LA, Kaufman AJ, Jacobsen, SB. 1992. Sedimentary cycling and environmental change in the Late Proterozoic: evidence from stable and radiogenic isotopes. *Geochim. Cosmochim. Acta* 56: 1317–29

Des Marias DJ, Strauss H, Summons RE, Hayes JM.1992. Carbon isotope evidence for the stepwise oxidation of the Proterozoic environment. *Nature* 359: 605–9

De Wit M, Jeffery M, Bergh H, Nicolaysen L 1988. *Geological map of sectors of Gondwana reconstructed to their disposition* ~150 *Ma* (1: 10,000,000).Tulsa: Am. Assoc. Petrol. Geol.

Dixon TH. 1981. Age and chemical characteristics of some pre-Pan African rocks in the Egyptian shield. *Precambrian Res.* 14: 119–33

Dixon TH, Golombek MP 1988. Late Precambrian crustal accretion rates in northeast Africa and Arabia. *Geology* 16: 991–94

Duncan IJ, Rivard B, Arvidson RE, Sultan M. 1990. Structural interpretation and tectonic evolution of a part of the Najd Shear Zone (Saudi Arabia) using Landsat thematic mapper data. *Tectonophysics* 178: 309–35

El-Isa Z, Mechie J, Prodehl C, Makris J, Rihm R. 1987. A crustal structure study of Jordan derived from seismic refraction data. *Tectonophysics* 138: 235–53

Fleck RJ, Coleman RG, Cornwall HR, Greenwood WR, Hadley DG, et al. 1976. Geochronology of the Arabian Shield, western Saudi Arabia: K-Ar results. *Geol. Soc. Am. Bull.* 87: 9–21

Forsyth D, Uyeda S. 1975. On the relative importance of the driving forces of plate tectonics. *Geophys. J. R. Astron. Soc.* 43: 163–200

Fullagar, P.D. 1978. Pan-African age granites of Northeastern Africa: new or reworked sialic materials? In *Geology of Libya*, Vol. 3, ed. MJ Salem, MT Busrewil, pp. 1051–58. New York: Academic

Gass IG, Ries AC, Shackleton RM, Smewing JD. 1990. Tectonics, geochronology and geochemistry of the Precambrian rocks of Oman. In *The Geology and Tectonics of the Oman Region*, ed. AHF Robertson, MP Searle, AC Ries, pp. 585–99. *Geol. Soc. London Spec. Publ.* 49

Gettings ME, Blank HR, Mooney WD, Healey JH. 1986. Crustal structure of southwestern Saudi Arabia. *J. Geophys. Res.* 91: 6491–512

Gichile S. 1992. Granulites in the Precambrian basement of southern Ethiopia: geochemistry, P-T conditions of metamorphism and tectonic setting. *J. Afr. Earth Sci.* 15: 251–63

Greenberg JK. 1981. Characteristics and origin of Egyptian Younger Granites: summary. *Geol. Soc. Am. Bull.* 92: 224–32

Groenewald PB, Grantham GH, Watkeys MK. 1991. Geological evidence for a Proterozoic to Mesozoic link between southeastern Africa and Dronning Maud Land, Antarctica. *J. Geol. Soc. London* 148: 1115–23

Guerrot C, Cocherie A, Ohnenstetter M. 1993. Origin and evolution of the West Andriamena Pan African mafic-ultramafic complexes in Madagascar as shown by U-Pb, Nd isotopes and trace element constraints. *Terra Abstr.* 5: 387

Gurnis M. 1988. Large-scale mantle convection and the aggregation and dispersal of supercontinents. *Nature* 332: 695–99

Hambrey MJ, Harland WB. 1985. The Late Proterozoic glacial era. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 51: 255–72

Hanson RE, Donovan RN, Wilson TJ. 1989. The late Proterozoic Zambezi belt in southern Africa: a model for the deeper levels of the southern Oklahoma aulacogen. *Geol. Soc. Am., Abstr. with Programs* 21: 13

Hanson RE, Wilson TJ, Brueckner HK, Onstott TC, Wardlaw MS, et al. 1988a. Reconnaissance geochronology, tectono-

thermal evolution, and regional significance of the middle Proterozoic Choma-Kalomo block, southern Zambia. *Precambrian Res.* 42: 39–61
Hanson RE, Wilson TJ, Munyanyiwa H. 1993. Geologic evolution of the Late Proterozoic Zambezi Belt in Zambia. *J. Afr. Earth Sci.* In press
Hanson RE, Wilson TJ, Wardlaw MS. 1988b. Deformed batholiths in the Pan-African Zambezi belt, Zambia: age and implications for regional Proterozoic tectonics. *Geology* 16: 1134–37
Harms U, Schandelmeier H, Darbyshire DPF. 1990. Pan-African reworked early/middle Proterozoic crust in NE Africa west of the Nile: Sr and Nd isotope evidence. *J. Geol. Soc. London* 147: 859–72
Harris NBW, Gass IG, Hawkesworth CJ. 1990. A geochemical approach to allochthonous terranes: a Pan-African case study. *Phil. Trans. R. Soc. London Ser. A* 331: 533–48
Harris NBW, Hawkesworth CJ, Ries AC. 1984. Crustal evolution in northeast and east Africa from model Nd ages. *Nature* 309: 773–76
Harrison TM, Copeland P, Kidd WSF, Yin A. 1992. Raising Tibet. *Science* 255: 1663–70
Hawkesworth CJ, Menzies MA, van Calsteren P. 1986. Geochemical and tectonic evolution of the Damara belt, Namibia. In *Collision Tectonics*, ed. MP Coward, AC Ries, pp. 305–19. London: Geol. Soc.
Henjes-Kunst F, Altherr R, Baumann A. 1990. Evolution and composition of the lithospheric mantle underneath the western Arabian peninsula: constraints from Sr-Nd isotope systematics of mantle xenoliths. *Contrib. Mineral. Petrol.* 195: 460–72
Hepworth JV. 1967. The photogeological recognition of ancient orogenic belts in Africa. *J. Geol. Soc. London* 123: 253–92
Hepworth JV, Kennerly JB. 1970. Photogeology and structure of the Mozambique orogenic front near Kolo, north-east Tanzania. *J. Geol. Soc. London* 125: 447–79
Hoffman PF. 1989. Speculations on Laurentia's first gigayear (2.0 to 1.0 Ga). *Geology* 17: 135–38
Hoffman PF. 1991. Did the Breakout of Laurentia turn Gondwanaland inside-out? *Science* 252: 1409–12
Husseini MI. 1989. Tectonic and deposition model of late Precambrian–Cambrian Arabian and adjoining plates. *Am. Assoc. Petrol. Geol. Bull.* 73: 1117–31
Husseini MI, Husseini SI. 1990. Origin of the Infracambrian Salt basins of the Middle East. In *Classic Petroleum Provinces*, ed. J Brooks, pp. 279–92. *Geol. Soc. London Spec. Publ.* 50
James HL. 1983. Distribution of banded iron formation in space and time. In *Iron Formations: Facts and Problems*, ed. AF Trendall, RC Morris, pp. 471–90. Amsterdam: Elsevier
Johnson RL. 1968. Structural history of the western front of the Mozambique Belt in northeast Southern Rhodesia. *Geol. Soc. Am. Bull.* 79: 513–26
Kennedy WQ. 1964. The structural differentiation of Africa in the Pan-African (±500 m.y.) tectonic episode. *Leeds Univ. Res. Inst. Afr. Geol. Annu. Rep.* 8: 48–49
Key RM, Charsley TJ, Hackman BD, Wilkinson AF, Rundle CC. 1989. Superimposed Upper Proterozoic collision-controlled orogenies in the Mozambique orogenic belt of Kenya. *Precambrian Res.* 44: 197–225
Klerkx J, Deutsch S. 1977. Resultats preliminaires obtenus par la methode Rb/Sr sur l'age des formations Precambriennes de la region d'Uweinat (Libye). *Mus. R. Afr. Centrale, Tervuren (Belg.) Dept. Geol. Min. Rapp. Annu.* 1976, pp. 83–94
Knoll A. 1992. The early evolution of Eukaryotes: a geological perspective. *Science* 256: 622–27
KRISP Working Party. 1991. Large-scale variation in lithospheric structure along and across the Kenya rift. *Nature* 354: 223–27
Kröner A. 1984. Late Precambrian plate tectonics and orogeny: a need to redefine the term Pan-African. In *African Geology*, ed. J Klerkx, J Michot, pp. 23–28. Tervuren: Musée. R. l'Afrique Centrale
Kröner A. 1991. African linkage of Precambrian Sri Lanka. *Geol. Rundsch.* 80: 429–40
Kröner A. 1993. The Pan African belt of northeastern and Eastern Africa, Madagascar, southern India, Sri Lanka and East Antarctica: terrane amalgamation during the formation of the Gondwana supercontinent. See Thorweihe & Schandelmeier 1993, pp. 3–9
Kröner A, Linnebacher P, Stern RJ, Reischmann T, Manton W, Hussein IM. 1991. Evolution of Pan-African island arc assemblages in the southern Red Sea Hills, Sudan, and in southwestern Arabia as exemplified by geochemistry and geochronology. *Precambrian Res.* 53: 99–118
Kröner A, Pallister JS, Fleck RJ. 1992a. Age of initial oceanic magmatism in the Late Proterozoic Arabian Shield. *Geology* 20: 803–6
Kröner A, Reischmann T, Todt W, Zimmer M, Stern RJ, et al. 1989. Timing, mechanism, and geochemical patterns of arc

accretion in Arabian-Nubian Shield and its extensions into Israel and Somalia. *Abstr. 28th Int. Geol. Congr.* 2: 231

Kröner A, Stern RJ, Dawoud AS, Compston W, Reischmann T. 1987. The Pan-African continental margin in northeastern Africa: evidence from a geochronological study of granulites at Sabaloka, Sudan. *Earth Planet. Sci. Lett.* 85: 91–104

Kröner A, Todt W, Hussein IM, Mansour M, Rashwan AA. 1992b. Dating of late Proterozoic ophiolites in Egypt and the Sudan using the single grain zircon evaporation technique. *Precambrian Res.* 59: 15–32

Küster D, Utke A, Leupolt L, Lenoir JL, Haider A. 1990. Pan-African granitoid magmatism in northeastern and southern Somalia. *Berliner Geowiss. Abh.* 120: 519–36

Leggo PJ. 1974. A geochronological study of the basement complex of Uganda. *J. Geol. Soc. London* 130: 263–77

Lindsay JF, Korsch RJ, Wilford JR. 1987. Timing the breakup of a Proterozoic supercontinent: evidence from Australian intracratonic basins. *Geology* 15: 1061–64

Maboko MAH, Boelrijk NAIM, Priem HNA, Verdurmen EAT. 1985. Zircon U-Pb and biotite Rb-Sr dating of the Wami River granulites, Eastern Granulites, Tanzania: evidence for approximately 715 Ma old granulite-facies metamorphism and final Pan-African cooling approximately 475 Ma ago. *Precambrian Res.* 30: 361–78

Maboko MAH, McDougall I, Zeitler PK. 1989. Dating late Pan-African cooling in the Uluguru granulite complex of Eastern Tanzania using the ^{40}Ar-^{39}Ar technique. *J. Afr. Earth Sci.* 9: 159–67

Malisa E, Muhongo S. 1990. Tectonic setting of gemstone mineralization in the Proterozoic metamorphic terrane of the Mozambique belt in Tanzania. *Precambrian Res.* 46: 167–76

McGuire AV, Stern RJ. 1993. Granulite xenoliths from western Saudi Arabia: the Lower Crust of the Late Precambrian Arabian-Nubian Shield. *Contrib. Mineral. Petrol.* 114: 395–408

McWilliams MO. 1981. Palaeomagnetism and Precambrian tectonic evolution of Gondwana. In *Precambrian Plate Tectonics*, ed. A Kröner, pp. 649–87. Amsterdam: Elsevier

Miller MM, Dixon TH. 1992. Late Proterozoic evolution of the northern part of the Hamisana zone, northeast Sudan: constraints on Pan-African accretionary tectonics. *J. Geol. Soc. London* 149: 743–50

Miller RMcG, Burger AJ. 1983. U-Pb zircon age of the early Damaran Naauwpoort formation. In *Evolution of the Damara Orogen of Southwest Africa/Namibia*, ed. RMcG Miller, pp. 267–72. *Spec. Publ. Geol. Soc. S. Afr.* 11

Molnar P, Atwater T. 1978. Interarc spreading and Cordilleran tectonics as alternatives related to the age of subducted oceanic lithosphere. *Earth Planet. Sci. Lett.* 41: 330–40

Moores EM. 1991. The southwest U.S.—East Antarctica (SWEAT) connection: a hypothesis. *Geology* 19: 425–28

Mosley PN. 1993. Geological evolution of the Late Proterozoic "Mozambique Belt" of Kenya. *Tectonophysics* 221: 223–50

Muhongo S, Lenoir J-L. 1993. Pan-African granulite-facies metamorphism in the Mozambique Belt of Tanzania: evidence from U-Pb on zircon geochronology. *J. Geol. Soc. London* In press

Murphy JB, Nance RD. 1991. Supercontinent model for the contrasting character of Late Proterozoic orogenic belts. *Geology* 19: 469–72

Neary CR, Gass IG, Cavanagh BJ. 1976. Granitic association of northeastern Sudan. *Geol. Soc. Am. Bull.* 87: 1501–12

Nelson BK, DePaolo DJ. 1985. Rapid production of continental crust 1.7 to 1.9 b.y. ago: Nd isotopic evidence from the basement of the North American mid-continent. *Geol. Soc. Am. Bull.* 96: 746–54

Oldow JS, Bally AW, Avé Lallemant HG, Leeman WP. 1989. Phanerozoic evolution of the North Ameican Cordillera; United States and Canada. In *The Geology of North America—An overview*, ed. AW Bally, AR Palmer, pp. 139–232. Boulder: Geol. Soc. Am.

Pallister JS, Stacey JS, Fischer LB, Premo WR. 1988. Precambrian ophiolites of Arabia: geologic settings, U-Pb geochronology, Pb-isotope characteristics, and implications for continental accretion. *Precambrian Res.* 38: 1–54

Paquette J-L, Nédélec A, Moine B, Rakotondrazafy M. 1993. U-Pb, single zircon Pb-evaporation and Sm-Nd dating of the S.E. Madagascar domain: a prevailing Panafrican event. *Terra Abstr.* 5: 393

Pearce JA, Lippard SJ, Roberts S. 1984. Characteristics and tectonic significance of supra-subduction zone ophiolites. In *Marginal Basin Geology*, ed. BP Kokelaar, MF Howell, pp. 77–94. *Geol. Soc. London Spec. Publ.* 16

Pin C, Poidevin JL. 1987. U-Pb zircon evidence for a Pan-African granulite facies metamorphism in the Central African Republic. A new interpretation of the high-grade series of the northern border of the Congo Craton. *Precambrian Res.* 36: 303–12

Pinna P, Jourde G, Calvez JY, Mroz JP,

Marques JM. 1993. The Mozambique Belt in northern Mozambique: Neoproterozoic (1100–850 Ma) crustal growth and tectogenesis, and superimposed Pan-African (800–550 Ma) tectonism. *Precambrian Res.* 62: 1–59

Powell C, Li ZX, McElhinny M, Meert J, Park J. 1993. Paleomagnetic constraints on timing of the Neoproterozoic breakup of Rodinia and the Cambrian formation of Gondwanaland. *Geology* 21: 889–92

Price RC. 1984. *Late Precambrian mafic-ultramafic complexes in northeast Africa.* PhD dissertation. Open Univ., Milton Keynes, UK. 326 pp.

Prochaska W, Pohl W. 1983. Petrochemistry of some mafic and ultramafic rocks from the Mozambique Belt, northern Tanzania. *J. Afr. Earth Sci.* 1: 183–93

Reischmann T. 1986. *Geologie und Genese spätproterozoischer Vulkanite der Red Sea Hills, Sudan.* PhD dissertation. Johannes Gutenberg Univ., Mainz

Reischmann T, Kröner A, Basahel A.1984. Petrography, geochemistry, and tectonic setting of metavolcanic sequences from the Al Lith area, southwestern Arabian Shield. In *IGCP Proj.* 164, *Proc., Vol.* 6, *Pan-African crustal evolution in the Arabian-Nubian Shield*, pp. 365–79. Jeddah: Faculty of Sci., King Abdulaziz Univ.

Reymer A, Schubert G. 1984. Phanerozoic addition rates to the continental crust and crustal growth. *Tectonics* 3: 63–77

Ries AC, Vearncombe JR, Price RC, Shackleton RM. 1992. Geochronology and geochemistry of the rocks associated with a Late Proterozoic ophiolite in West Pokot, NW Kenya. *J. Afr. Earth Sci.* 14: 25–36

Rihm R, Makris J, Möller L. 1991. Seismic surveys in the northern Red Sea: asymmetric crustal structure. *Tectonophysics* 198: 279–95

Sacchi R, Marques J, Costa M, Casati C. 1984. Kibaran events in the southernmost Mozambique Belt. *Precambrian Res.* 25: 141–59

Sanders LD. 1965. Geology of the Contact between the Nyanza Shield and the Mozambique Belt in western Kenya. *Geol. Surv. Kenya Bull.* 7 45 pp.

Schandelmeier H, Darbyshire DPF, Harms U, Richter A. 1988. The East Saharan Craton: evidence for pre-Pan African crust in NE Africa west of the Nile. In *The Pan-African Belt of Northeast Africa and Adjacent Areas*, ed. S El-Gaby, RO Greiling, pp.69–94. Braunschweig: Vieweg

Schandelmeier H, Küster D, Wipfler E, Abdel Rahman EM, Stern RJ, et al 1993. Evidence for a new Late Proterozoic suture in Northern Sudan. See Thorweihe & Schandelmeier 1993, pp. 83–86

Schmidt DL, Hadley DG, Stoeser DB. 1979. Late Proterozoic crustal history of the Arabian Shield, southern Najd Province, Kingdom of Saudi Arabia. In *Evolution and Mineralization of the Arabian-Nubian Shield*, Vol. 2, ed. SA Tahoun, pp. 41–58. New York: Pergamon

Şengör CA. 1987. Tectonics of the Tethysides: orogenic collage development in a collisional setting. *Annu. Rev. Earth Planet. Sci.* 15: 213–44

Shackleton RM. 1986. Precambrian collision tectonics in Africa. In *Collision Tectonics*, ed. MP Coward, AC Ries, pp. 329–49. London: Geol. Soc.

Shackleton RM. 1993. Tectonics of the lower crust: a view from the Usambara Mountains, NE Tanzania. *J. Structural Geol.* 15: 663–71

Shanti M, Roobol MJ. 1979. A late Proterozoic ophiolite complex at Jabal Ess in northern Saudi Arabia. *Nature* 279: 488–91

Stacey JS, Agar RA. 1985. U-Pb isotopic evidence for the accretion of a continental microplate in the Zalm region of the Saudi Arabian Shield. *J. Geol. Soc. London* 142: 1189–204

Stacey JS, Hedge CE. 1984. Geochronologic and isotopic evidence for Early Proterozoic crust in the Eastern Arabian Shield. *Geology* 12: 310–13

Stacey JS, Stoeser DB, Greenwood WR, Fischer LB. 1984. U-Pb geochronology and geological evolution of the Halaban-Al Amar region of the Eastern Arabian Shield, Kingdom of Saudi Arabia. *J. Geol. Soc. London* 141: 1043–55

Stern RJ. 1979. *Late Precambrian ensimatic volcanism in the Central Eastern Desert of Egypt.* PhD dissertation. Univ. Calif., San Diego, 210 pp.

Stern RJ. 1981. Petrogenesis and tectonic setting of Late Precambrian ensimatic volcanic rocks, Central Eastern Desert of Egypt. *Precambrian Res.* 16: 195–230

Stern RJ. 1985. The Najd Fault System, Saudi Arabia and Egypt: a Late Precambrian rift-related transform system? *Tectonics* 4: 497–511

Stern RJ, Bloomer SH. 1992. Subduction zone infancy: examples from the Eocene Izu-Bonin-Mariana and Jurassic California arcs. *Geol. Soc. Am. Bull.* 104: 1621–36

Stern RJ, Dawoud AS. 1991. Late Precambrian (740 Ma) charnockite, enderbite, and granite from Jebel Moya, Sudan: a link between the Mozambique Belt and the Arabian-Nubian Shield? *J. Geol.* 99: 648–59

Stern RJ, Kröner A. 1993. Geochronologic and isotopic constraints on late Precambrian crustal evolution in NE Sudan. *J. Geol.* 101: 555–74
Stern RJ, Kröner A, Bender R, Dawoud AS. 1993. Juvenile Pan-African and remobilized ancient crustal components preserved in the basement around Wadi Halfa, N. Sudan: insights into the evolution of the foreland to the Arabian-Nubian Shield. See Thorweihe & Schandelmeier 1993, pp. 79–81.
Stern RJ, Kröner A, Rashwan AA. 1991. A late Precambrian (~710 Ma) high volcanicity rift in the southern Eastern Desert of Egypt. *Geol. Rundsch.* 80: 155–70
Stern RJ, Nielsen KC, Best E, Sultan M, Arvidson RE, Kröner A. 1990. Orientation of late Precambrian sutures in the Arabian-Nubian shield. *Geology* 18: 1103–6
Stoeser DB, Camp VE. 1985. Pan African microplate accretion of the Arabian Shield. *Geol. Soc. Am. Bull.* 96: 817–26
Stoeser DB, Whitehouse MJ, Agar RA, Stacey JS. 1991. Pan-African accretion and continental terranes of the Arabian Shield, Saudi Arabia and Yemen. *Eos, Trans. Am. Geophys. Union* 72: 299 (Abstr.)
Storey BC. 1993. The changing face of late Precambrian and early Palaeozoic reconstructions. *J. Geol. Soc. London* 150: 665–68
Sultan M, Bickford ME, El Kaliouby B, Arvidson RE. 1992a. Common Pb systematics of Precambrian granitic rocks of the Nubian Shield (Egypt) and tectonic implications. *Geol. Soc. Am. Bull.* 104: 456–70
Sultan M, Chamberlin KR, Bowring SA, Arvidson RE, Abuzied H, El Kaliouby B. 1990 Geochronologic and isotopic evidence for involvement of pre-Pan-African crust in the Nubian shield, Egypt. *Geology* 18: 761–64
Sultan M, Tucker RD, Gharbawi RI, Ragab AI, El Alfy Z. 1992b. On the extension of the Congo craton into the Western Desert of Egypt. *Geol. Soc. Am., Abstr. with Programs* 24: 52 (Abstr.)
Taylor SR, McClennan SM. 1985. *The Continental Crust: Its Composition and Evolution*. Oxford: Blackwell Sci. 312 pp.
Thorweihe U, Schandelmeier H, eds. 1993. *Geoscientific Research in Northeast Africa (abstracts)* Rotterdam: Balkema
Unrug R. 1993. Neoproterozoic assembly of the Gondwana supercontinent: definition of component cratons and the keystone position of the Salvador-Congo craton. *Precambrian Res.* In press
Vail JR. 1965. Zones of progressive regional metamorphism across the western margin of the Mozambique Belt in Rhodesia and Mozambique. *Geol. Mag.* 103: 231–39
Vail JR. 1985. Pan-African (Late Precambrian) tectonic terrains and the reconstruction of the Arabian-Nubian Shield. *Geology* 13: 839–42
Vearncombe JR. 1983. A proposed continental margin in the Precambrian of western Kenya. *Geol. Rundsch.* 72: 663–70
Veevers JJ. 1990. Tectonic-climatic supercycles in the billion-year plate-tectonic eon: Permian Pangean icehouse alternates with Cretaceous dispersed-continents greenhouse. *Sediment. Geol.* 68: 1–16
Veizer J, Compston W, Clauer N, Schidlowski M. 1983. $^{87}Sr/^{86}Sr$ in Late Proterozoic carbonates: evidence for a "mantle" event at ~900 Ma ago. *Geochim. Cosmochim. Acta* 47: 295–302
Wang P, Glover L III. 1992. A tectonics test of the most commonly used geochemical discriminant diagrams and patterns. *Earth Sci. Rev.* 33: 111–31
Whitehouse MJ, Windley BF, Ba-Bitat MA. 1993. Tectonic framework and geochronology of the southern Arabian Shield, Yemen. *Eos, Trans. Am. Geophys. Union* 74: 300–1 (Abstr.)
Wilson TJ, Hanson RE, Wardlaw MS. 1993. Late Proterozoic evolution of the Zambezi belt, Zambia: implications for regional Pan-African tectonics and shear displacements in Gondwana. In *Gondwana 8—Assembly, Evolution, and Dispersal*, ed. RH Findlay, R Unrug, MR Banks, JJ Veevers, pp. 69–82. Rotterdam: Balkema
Wright VP, Ries AC, Munn SG. 1990. Intraplatformal basin-fill deposits from the Infracambrian Huqf Group, east Central Oman. In *The Geology and Tectonics of the Oman Region*, ed. AHF Robertson, MP Searle, AC Ries, pp. 601–16. *Geol. Soc. London Spec. Publ.* 49
Wust HJ, Todt W, Kröner A. 1987. Conventional and single grain zircon ages for metasediments and granite clasts from the Eastern Desert of Egypt: evidence for active continental margin evolution in Pan-African times. *Terra Cognita* 7: 165 (Abstr.)

Annual Review of Earth and Planetary Sciences
Volume 22, 1994

CONTENTS