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A synopsis of events related to the assembly of eastern Gondwana

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Abstract

The assembly of the eastern part of Gondwana (eastern Africa, Arabian–Nubian shield (ANS), Seychelles, India, Madagascar, Sri Lanka, East Antarctica and Australia) resulted from a complex series of orogenic events spanning the interval from ~750 to ~530 Ma. Although the assembly of Gondwana is generally discussed in terms of the suturing of east and west Gondwana, such a view oversimplifies the true nature of this spectacular event. A detailed examination of the geochronologic database from key cratonic elements in eastern Gondwana suggests a multiphase assembly. The model outlined in this paper precludes the notion of a united east Gondwana and strongly suggests that its assembly paralleled the final assembly of greater Gondwana. It is possible to identify at least two main periods of orogenesis within eastern Gondwana. The older orogen resulted from the amalgamation of arc terranes in the Arabian–Nubian shield region and oblique continent–continent collision between eastern Africa (Kenya–Tanzania and points northward) with an, as of yet, ill-defined collage of continental blocks including parts of Madagascar, Sri Lanka, Seychelles, India and East Antarctica during the interval from ~750 to 620 Ma. This is referred to as the East Africa Orogen (EAO) in keeping with both the terminology and the focus of the paper by Stern [Annu. Rev. Earth Planet. Sci. 22 (1994) 319]. The second major episode of orogenesis took place between 570 and 530 Ma and resulted from the oblique collision between Australia plus an unknown portion of East Antarctica with the elements previously assembled during the East African Orogen. This episode is referred to as the Kuunga Orogeny following the suggestion of Meert et al. [Precambrian Res. 74 (1995) 225]. Paleomagnetic data are currently too few to provide a rigorous test of this proposal, but the extant data do not conflict with the notion of a polyphase assembly of eastern Gondwana. The major conclusion of this paper is that east Gondwana did not exist until its Cambrian assembly.

Keywords: East Africa Orogen; Kuunga Orogen; Pan-African; Gondwana; Mozambique belt

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1. Introduction

The Mozambique belt lies along the eastern margin of the African continent and is generally thought to represent a zone of continent–continent collision on the scale of the modern Alpine–Himalayan orogen although others have favored a largely ensialic origin (Holmes, 1951; Dewey and Burke, 1973; Burke et al., 1977; Stern, 1994; Piper, 2000). For the most part, the formation of the Mozambique belt is discussed in terms of a collision between east and west Gondwana, but such a description oversimplifies both the geometry and timing of Gondwana formation. For example,

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most analyses view the assembly of the western Gondwana elements (e.g., South American and African blocks) as a series of collisions marked by near final assembly at around 600 Ma (Trompette, 1997). Therefore, if the collisions along the Mozambique belt occurred before 600 Ma, then only parts of west Gondwana were involved in the collision. Stern (1994) and, more recently, Blasband et al. (2000) document evidence of a series of arc-terrane accretions in the Arabian–Nubian shield (ANS) region that spanned at least 100 million years beginning at roughly 750 Ma, indicating a protracted assembly of juvenile terranes and older continental fragments along the northern segment of the Mozambique belt, and referred to this as the East Africa Orogen (EAO).

The term Pan-African (Kennedy, 1964) referred to a sequence of events of tectonothermal events at 500 ± 100 Ma within Africa and adjacent Gondwana elements. The term was broadened by Kröner (1984) to include orogenic events of the same time range (950–450 Ma) on a more global scale. In the intervening years, the 'Pan-African' orogenic cycle has been more narrowly defined both spatially and temporally such that it is possible to recognize individual orogenic events within Gondwana (Trompette, 1997; Stern, 1994; Meert et al., 1995). The term Pan-African is likely to remain popular, but it no longer provides a level of specificity commensurate with our knowledge of the tectonic history of Gondwana assembly. This paper outlines the orogenic events associated with the assembly of the eastern part of Gondwana.

Although the timing of Gondwana assembly will continue to be debated for some time, the history of eastern Gondwana amalgamation must begin with a discussion of where the cratonic elements originated prior to their late Neoproterozoic to early Cambrian fusion. The notion of a Meso–Neoproterozoic supercontinent is entrenched in the geologic literature (Piper, 1976; Bond et al., 1984; McMenamin and McMenamin, 1990; Dalziel, 1991; Hoffman, 1991; Karlstrom et al., 1999) although the configurations differ widely. A number of names have been proposed for this supercontinent including Ur-Gondwana (Hartnady, 1986, 1991), Paleopangea (Piper, 2000) and Rodinia (McMenamin and McMenamin, 1990). The name 'Rodinia,' which takes its name from the Russian prefix 'to beget,' is adopted here. According to Dalziel (1997), the Rodinia supercontinent formed during a series of late Meso to early Neoproterozoic collisions lumped under the term 'Grenvillian.' Its breakup is represented by the presence of Neoproterozoic IV-aged rift and passive margin-related sequences (Dewey and Burke, 1973; Bond et al., 1984; Dalziel, 1991; Moores, 1991; Hoffman, 1991; Knoll, 2001). Although there are debates regarding the exact position of various elements surrounding Rodinia (e.g., Piper, 2000; Sears and Price, 2000; Dalziel, 1997), there is clear evidence that Laurentia occupied the center of a major landmass. As noted by Dalziel (1997), whatever the final configuration of the supercontinent, the length of rifted margins surrounding Laurentia must be accounted for by a similar length of rifted margins in the formerly contiguous blocks. For simplicity, I adopt a variation of the Rodinia supercontinent shown in Fig. 1 (ca. 800 Ma; Dalziel, 1997; Weil et al., 1998; Torsvik et al., 1996). There is one caveat in adopting this model for the starting point of this paper. The popular model assumes that east Gondwana (Madagascar, Sri Lanka, Australia, India and Antarctica) was united by 1000 Ma. One of the conclusions of this paper is that east Gondwana never existed as a coherent block until all its constituent cratons were assembled in Neoproterozoic to early Cambrian time. Nevertheless, the geometry shown in Fig. 1 allows for a starting point in Gondwana assembly as the elements of Gondwana are more or less dispersed about the Laurentian continent. There are debates surrounding the exact timing of various rift events along the western margin of Laurentia; however, as discussed below, the available paleomagnetic evidence suggests that the rift-to-drift transition was prior to 750 Ma.

Owing to its position in east Gondwana (Fig. 2), the East Antarctic craton is viewed as the keystone continent in east Gondwana (see Yoshida, 1995; Rogers, 1996). Links between the Albany–Fraser belt (Australia) and the Wilkes Province (Antarctica, Fig. 2) were used to argue in support of a Mesoproterozoic (1300–1200 Ma) link between the two continents (Sheraton et al., 1995; Nelson et al., 1995; Post et al., 1997). Dalziel (1992) considered the collisional events in the Wilkes Province (Antarctica), the eastern Ghats region of India and the Prince Charles Mountains (Antarctica) as broadly coeval provinces formed during final assembly of Rodinia; however, recent work in the eastern Ghats region and the northern Prince Charles Mountains (nPCMs) of East Antarctica

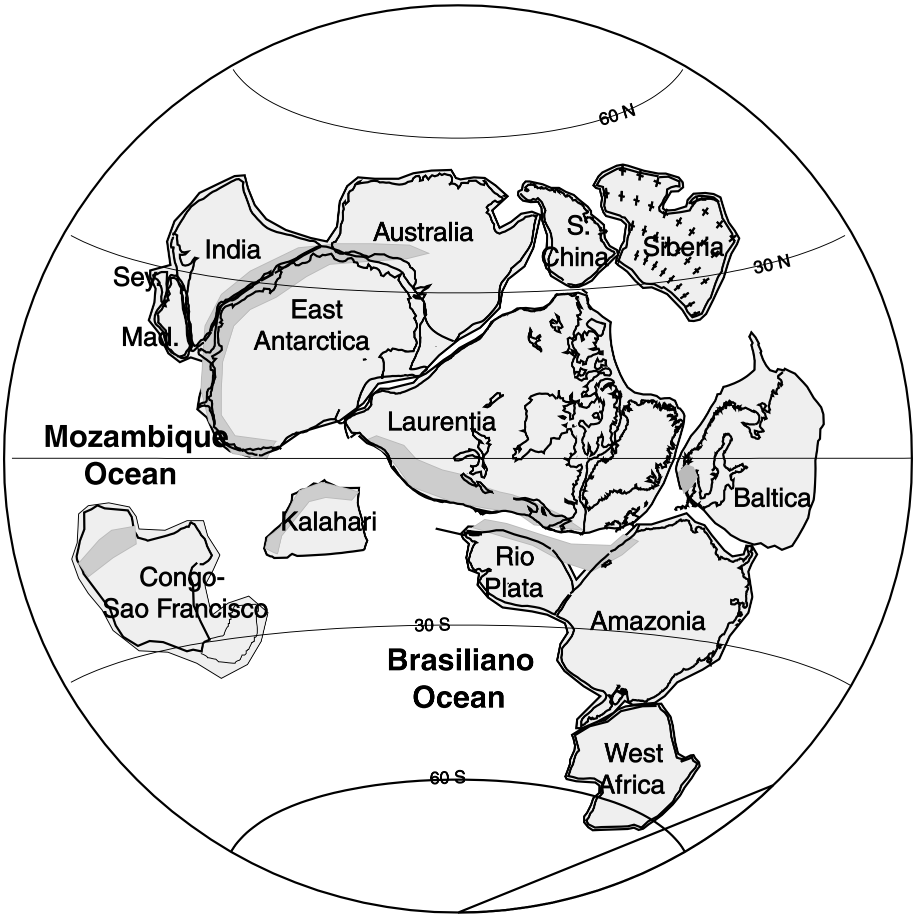
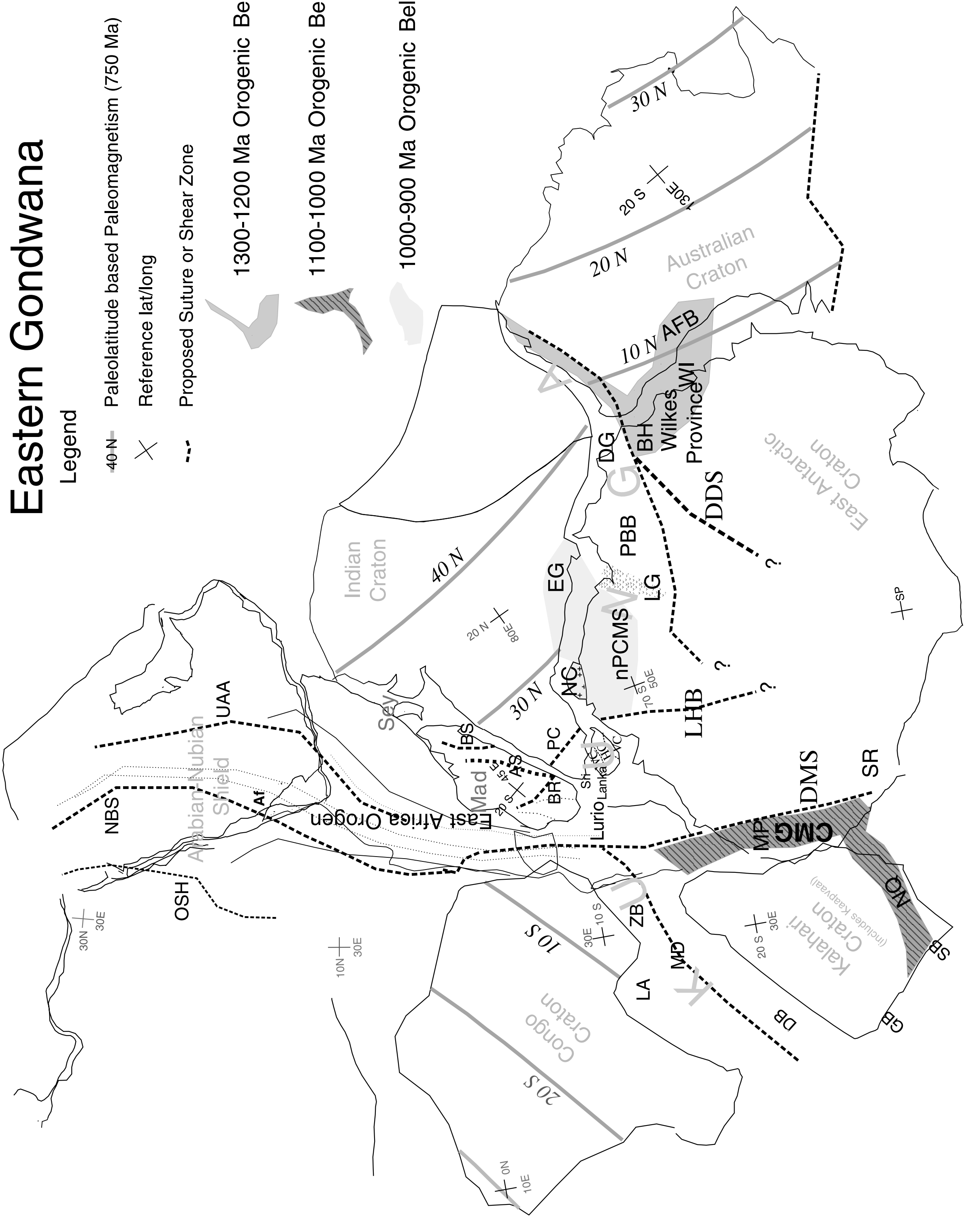


Fig. 1. The supercontinent Rodinia at ca. 800 Ma (Laurentian coordinates). The fit is slightly modified from those of Torsvik et al. (1996), Dalziel (1997), Weil et al. (1998). Darker-grey shading represents the 'Grenvillian-age' belts marking the Meso-Neoproterozoic suturing of the supercontinent.

(Mezger and Cosca, 1999; Boger et al., 2000) indicate that these blocks were incorporated into east Gondwana during a much younger 900–1000 Ma orogenesis (Fig. 2).

Additional hints that East Antarctica might not comprise a single block came about through the recognition of the similar-aged tectonic histories of the Maud Province and the Kaapvaal craton in the interval 1100–1000 Ma (Thomas et al., 1994; Cornell et al., 1996; Jacobs et al., 1996, 1998). Gose et al. (1997) argued, on paleomagnetic grounds, that the CMG terrane (Coats Land, Maudheim, Grunehogna) was juxtaposed against the Kalahari craton in early Neoproterozoic times (Fig. 2). Despite the recognition that the CMG terrane was likely a part of Africa, the notion of an undivided east Gondwana during the Neoproterozoic remained largely unchallenged (see Kröner, 1991; Meert et al., 1995; Kröner et al., 2000a,b for alternative suggestions). Recently, Fitzsimons (2000a,b) has noted the possible existence of Cambrian-aged suture zones in East Antarctica that separate the nPCMs/Ghats regions from the Wilkes province. In addition, the possible southern extension of the East Africa Orogen into the Lützow–Holm region would juxtapose the CMG terrane with the remainder of east Gondwana during the Cambrian as argued by Fitzsimons (2000b) and Grunow et al. (1996). Other authors have also argued for a multiphase late Neoproterozoic–early Cambrian assembly of eastern Gondwana (Meert and Van der Voo, 1997;


Eastern Gondwana
Legend
40 N Paleolatitude based Paleomagnetism (750 Ma)
Reference lat/long
Proposed Suture or Shear Zone
1300-1200 Ma Orogenic Belt
1100-1000 Ma Orogenic Belt
1000-900 Ma Orogenic Belt
Australian Craton
30 N
20 N
10 N
20 S 130E
AFB
Wilkes Province
WI
BH
DG
DDS
East Antarctic Craton
PBB
LG
EG
nPCMS
NC
70 S 50E
LHB
SP
Indian Craton
40 N
30 N
20 N 80E
Sey
BS
AS
PC
BR
SH
Lanka
VC
Mad
20 S 45 E
Lurio
East Africa Orogen
Arabian-Nubian Shield
UAA
NBS
Af
OSH
30N 30E
10N 30E
DMS
SR
MP
CMG
NO
Kalahari Craton
(includes Kaapvaal)
20 S 30E
SB
GB
DB
MD
ZB
LA
30E 10 S
10 S
20 S
Congo Craton
0N 10E
G
A
K
U

Hensen and Zhou, 1997). Madagascar and Sri Lanka were usually considered 'minor' elements of east Gondwana, but the recent recognition that Madagascar may itself contain several sutures has sparked renewed interest in the geochronological and tectonic setting of this continental block (Paquette and Nédélec, 1998; Cox et al., 1998; Handke and Tucker, 1999; Kröner et al., 2000a,b; de Wit et al., 2001).

One of the major advances in understanding the complexity of Gondwana assembly is the ability to date metamorphic and igneous events with high-precision U–Pb geochronology using a variety of methods (e.g., SHRIMP, isotope dilution, Pb–Pb evaporation and electron microprobe). The U–Pb system is particularly robust because a single zircon may contain information regarding both crystallization and metamorphic events. Each of the cratons involved in the assembly of eastern Gondwana now has reliable U–Pb age data that can be tied to a tectonic framework. Other isotopic dating methods, such as ^{40}Ar/^{39}Ar and ^{147}Sm/^{144}Nd, along with knowledge of closure temperatures in those systems have facilitated the development of detailed cooling histories of orogenic belts. The events in some regions are better constrained than in others (e.g., Arabian–Nubian Shield vs. Kenya–India); however, collectively, they yield important information regarding the assembly of eastern Gondwana. The geochronologic data for each of the elements is reviewed below and placed in a regional tectonic framework.

Paleomagnetic studies from these blocks have the potential to discriminate amongst the various tectonic models (Meert, 2001; Torsvik et al., 2001b; Meert, 1999), but the relatively poor quality of the extant database has hindered progress (Meert and Powell, 2001). Nevertheless, recent work (described below) hints that a polyphase assembly of eastern Gondwana is possible. This paper describes the geologic, geochronologic and paleomagnetic evidence for a polyphase assembly of the eastern Gondwana region.

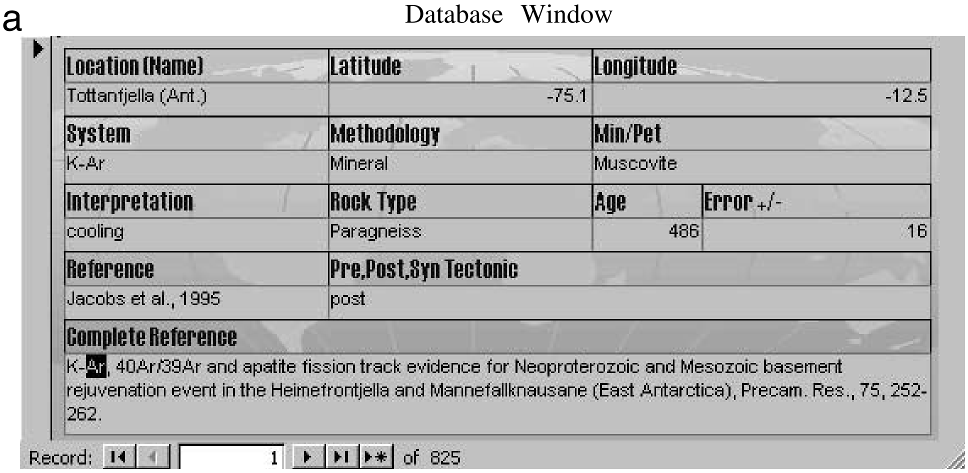
2. Geology–geochronology

2.1. The database

The author has compiled a database of radiometric ages from eastern Gondwana covering the ~ 800 to 400 Ma interval that have been published since 1985. In regions where there is poor geochronological coverage, the search was extended to include older publications. The database is assembled in Microsoft Access 1997 format and is available online (http://www.clas.ufl.edu/users/jmeert). A total of 1057 age determinations are included in the present database and it is updated through a literature search every 3 months (currently through March 2001). An example of the database window is shown in Fig. 3a and the age–frequency distribution for the database is shown in Fig. 3b (ages are binned in 5 Ma intervals). The areal distribution of ages is shown in Fig. 4 and each region is discussed in detail below. Version 1.0 of the database is released with full access rights for any user. Future versions of the database will be more restrictive in terms of editing individual entries.

An important factor for evaluating the available geochronologic data is establishing a correlation between ages and tectonic/metamorphic events. In an effort to place ages in their proper tectonic setting, the designation of 'pre,' 'syn' and 'post' assembly was made based on the author's interpretation of the ages whenever possible. These designations must be

Fig. 2. A map of eastern Gondwana with the circum-Antarctic mobile belts highlighted by age differences (See Fitzsimons, 2000a). The paleolatitude lines for Congo, India and Australia are based on 750 Ma data from those continents listed in Table 1. Abbreviations for this figure are as follows: **Af** = Afif terrane (Saudi Arabia); **AFB** = Albany–Fraser Belt (Australia); **AS** = Angavo shear zone (Madagascar); **BH** = Bunger Hills (Antarctica); **BR** = Bongolava–Ranotsara shear zone (Madagascar); **BS** = Betsimisaraka suture zone (Madagascar); **CMG** = Coats Land–Maudheim–Grunehogna (Antarctica); **DB** = Damara Belt (Africa); **DDS** = Darling–Denman suture (Antarctica–Australia); **DG** = Denman Glacier (Antarctica); **DMS** = Dronning Maud suture (Antarctica); **EG** = Eastern Ghats (India); **GB** = Gariep Belt (Africa); **HC** = Highland Complex (Sri Lanka); **LA** = Lufilian Arc (Africa); **LG** = Lambert Graben (Antarctica); **LHB** = Lützow–Holm Belt (Antarctica); **MD** = Mwembeshi Dislocation (Africa); **MP** = Maud Province (Antarctica); **NBS** = Nabitah Suture (Arabia); **NC** = Napier Complex (Antarctica); **nPCSM** = Northern Prince Charles Mountains (Antarctica); **NQ** = Namaqua Belt (Africa); **OSH** = Onib–Sol Hamed suture (Arabia); **PBB** = Prydz Bay Belt (Antarctica); **PC** = Paughat–Cauvery (India); **SB** = Saldania Belt (Africa); **SR** = Shackelton Range (Antarctica); **UAA** = Urd Al Amar suture (Arabia); **VC** = Vijayan Complex (Sri Lanka); **WC** = Wanni Complex (Sri Lanka); **WI** = Windmill Islands (Antarctica); **ZB** = Zambezi Belt (Africa).



Location (Name)	Latitude	Longitude	
Tottanfjella (Ant.)	-75.1	-12.5	
System	**Methodology**	**Min/Pet**	
K-Ar	Mineral	Muscovite	
Interpretation	**Rock Type**	**Age**	**Error +/-**
cooling	Paragneiss	486	16
Reference	**Pre,Post,Syn Tectonic**		
Jacobs et al., 1995	post		

Complete Reference

K-Ar, 40Ar/39Ar and apatite fission track evidence for Neoproterozoic and Mesozoic basement rejuvenation event in the Heimefrontjella and Mannefallknausane (East Antarctica), Precam. Res., 75, 252-262.

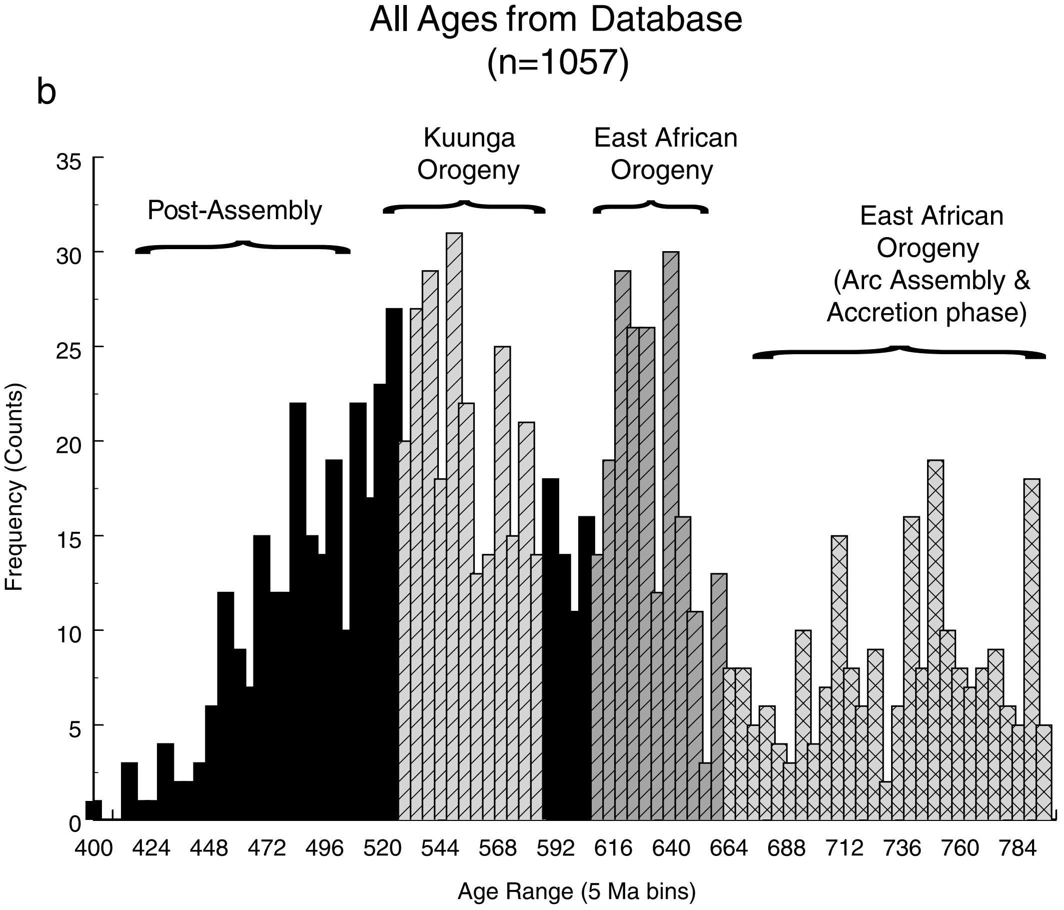


Fig. 3. (a) Sample database window from the geochronologic database used in this paper (Microsoft Access 1997). (b) A histogram showing the frequency of published ages from the entire database (in 5 Ma bins). Orogenic events described in this paper are given at the top of the diagram.

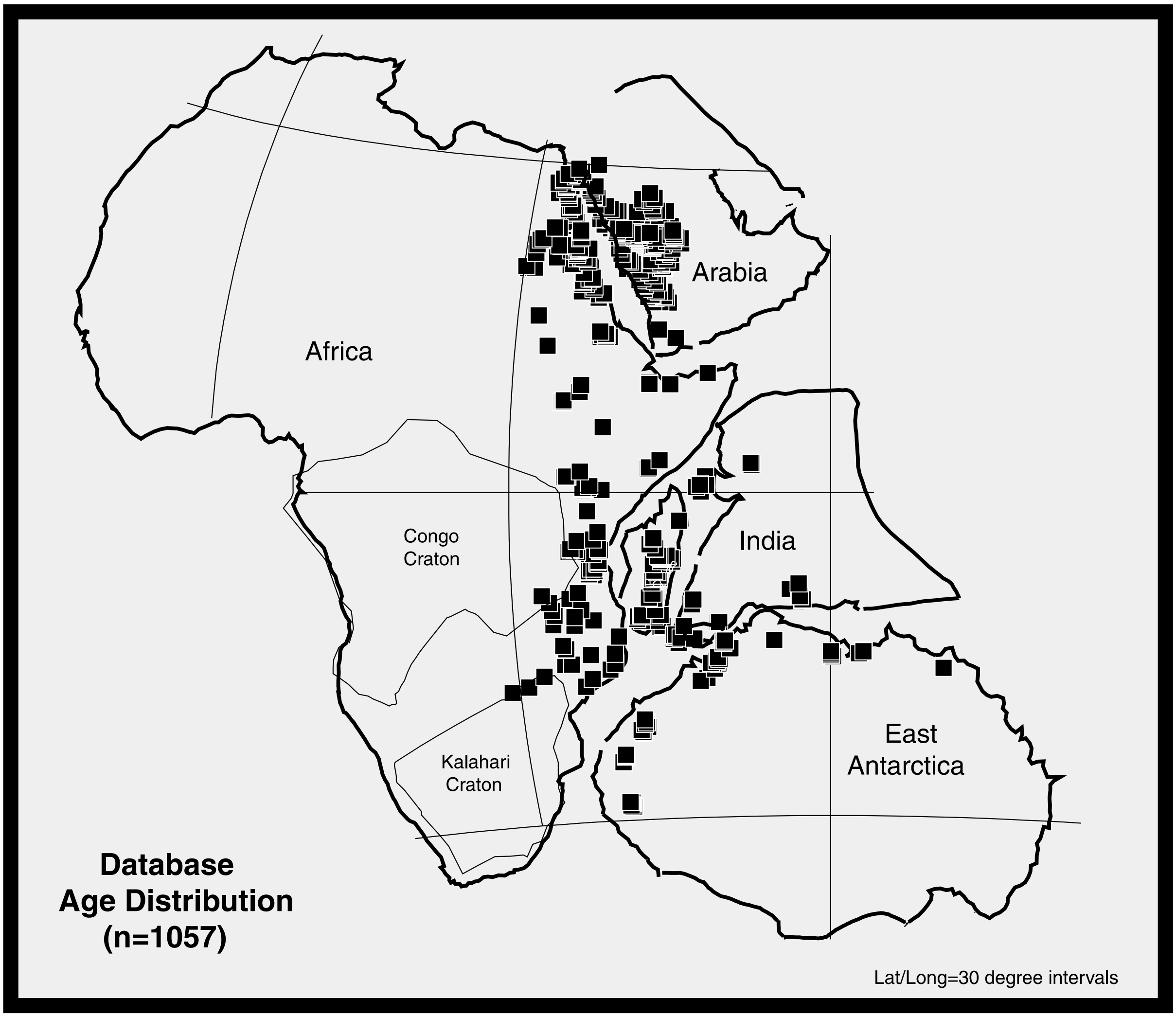


Fig. 4. Location of geochronologic studies within eastern Gondwana used in the database. Due to the scale of the map, multiple-age determinations may be represented by a single point.

supported by structural, geochemical and geological arguments. Clearly, all ages are related to some form of tectonic process, so the designation of 'Pre' assembly signifies that the ages reflect the crystallization of rocks prior to Gondwana amalgamation (e.g., protoliths, ophiolites and arc-related igneous rocks) or precollisional metamorphism. 'Syn' assembly signifies that the ages of the rocks were interpreted as having formed during a collisional event related to Gondwana assembly (as documented in the original manuscripts) and postassembly ages signify either cooling ages, extensional phases of tectonism following collision or late shear events that followed the amalgamation of Gondwana elements. Many older studies interpreted K–Ar mineral ages and Rb–Sr ages as crystallization ages in metamorphic terranes. Unless these ages are independently verified by U–Pb or Pb–Pb ages, they were reinterpreted as cooling ages in the database in keeping with the understanding of their behavior during metamorphism.

The database is provided to the user without any quality 'filter' on individual entries as various schemes can be devised depending on the particular research initiative; however, each datapoint in this study was evaluated according to the criteria given in Table 1. The goal of this particular grading scheme was to eliminate age determinations that have large errors, are poorly documented or use methods that yield ages of uncertain tectonic significance (e.g., errorchrons or 'model' ages). The cutoff grade for inclusion into this analysis was "C" or better (with exceptions noted). Application of this filter eliminates about 15% of the database and reduces the average error to ± 10 Ma. In practical terms, this means that the typical uncertainty in the binned histogram distributions is ± two bins. Fig. 5 demonstrates that inclusion of "C" graded ages does not result in any degradation of the 'signal' of the syn-assembly distribution.

There are several additional observations with regard to the database and the analysis presented in this paper. Fig. 6a shows the frequency distribution of unfiltered ages based on the U–Pb decay scheme ($n = 607$), and Fig. 6b shows the frequency distribution of ages based on other decay schemes ($n = 450$). Although the U–Pb distribution is sawtooth, there are three broad modalities in the data (e.g., 530–570, 610–650 and 740–800 Ma). Radiometric ages using other methods are skewed toward the lower end of the age range (< 600 Ma; mean 570 Ma).

Fig. 7a shows the frequency distribution of graded (C and above) 'pre' assembly ages from eastern Gondwana. The pattern toward the high end of the age range likely reflects the formation and amalgamation of arc terranes in the Arabian–Nubian sector as described below. There are three broad modalities in the syn-assembly data (Fig. 5). The first is at about 750 Ma and represents the beginning of arc amalgamation in the Arabian–Nubian shield sector as described below. The second ranges from ~ 610 to 650 Ma and the youngest from about 520 to 570 Ma. Postassembly ages are shown in Fig. 7b and show a broader range of ages, but most are younger than 640 Ma (mean 544 Ma). The last point is that Section 5 is framed in terms of current geographic boundaries for convenience and does not imply specific boundaries within Gondwana.

2.2. The Arabian–Nubian shield sector

The tectonic setting and geochronology of the Arabian–Nubian shield (ANS) and its environs were discussed by a number of authors (Stoeser and Van Camp, 1985; Kröner et al., 1990; Stern, 1994; Windley et al., 1996; Al-Saleh et al., 1998; Cosca et al., 1999; Blasband et al., 2000; Johnson and Kattan, 2001; Whitehouse et al., 2001). In general, the region is viewed as a collage of juvenile arc terranes and associated ophiolitic remnants formed in the Mozambique Ocean beginning around 900 Ma (Stern, 1994). In his seminal paper on the East Africa Orogen (EAO), Stern (1994) argued for collision and amalgamation of these terranes by ~ 650 Ma. More recently, Stern and Abdelsam (1998) discuss alternative models for the formation of juvenile crust in the Arabian–Nubian shield in considerable detail. Their

Table 1
Age 'grading' criterion

Criteria used	Are these criteria Met (Y/N) and Grade?
(1) Age is provided with analytical details and diagrams (e.g., isochron plots, concordia diagrams, plateaus). Age quoted is not an errorchron or model age. Age error is within ± 75 Ma.	*Yes*—Proceed to #2 *No*—Grade '**E**'
(2) Age is tied to specific structural/orogenic/magmatic event. Age error is within ± 50 Ma.	*Yes*—Proceed to #3 *No*—Grade '**D**'
(3) Age is based on mineral separates (e.g., single zircons) or whole-rock/mineral combined analysis (Rb–Sr, Sm–Nd). Age error is within ± 25 Ma. Ar–Ar ages are not integrated or total gas ages.	*Yes*—Proceed to #4 *No*—Grade '**C**'
(4) Error is within ± 10 Ma.	*Yes*—Grade '**A**' *No*—Grade '**B**'

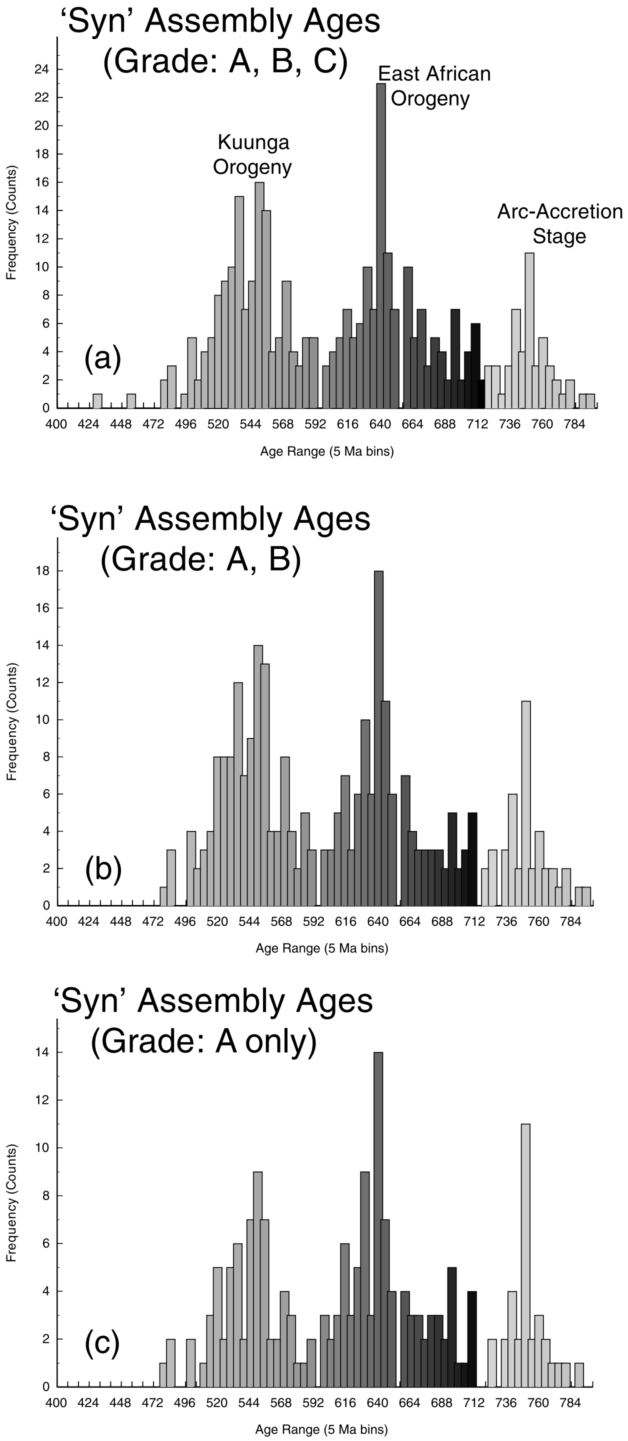


Fig. 5. (a) Histogram of ‘syn’ assembly ages graded ‘C’ or better along with the authors’ suggested interpretation of the orogenic events. (b) Histogram of ‘syn’ assembly ages graded ‘B’ and better and (c) histogram of ‘syn’ assembly ages rated ‘A’.

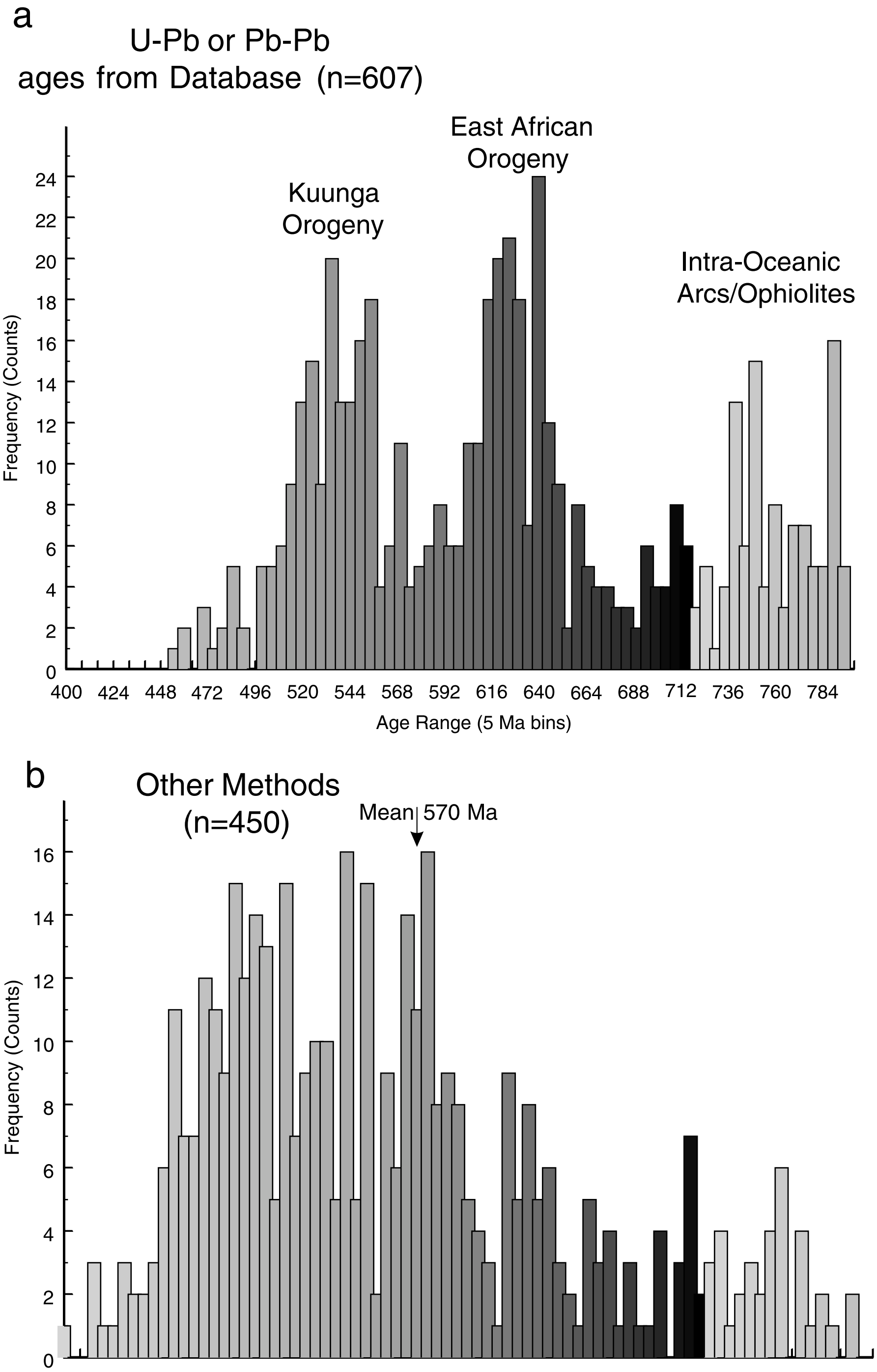


Fig. 6. (a) Histogram showing the frequency of U–Pb ages from the current database (in 5 Ma bins) and the interpreted major orogenic events withing eastern Gondwana. (b) Histogram showing the frequency of ages determined by other methods from the current database (in 5 Ma bins).

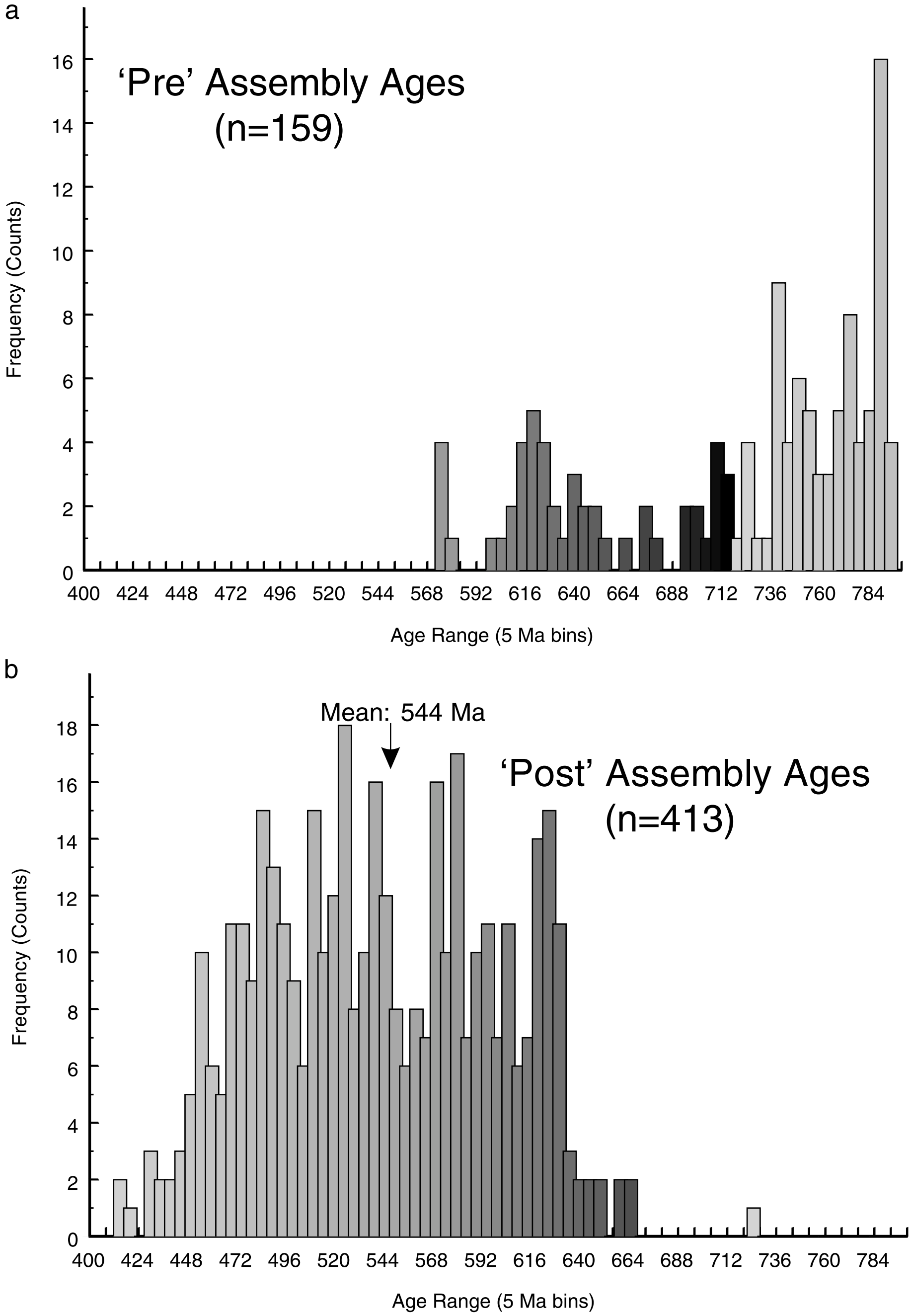


Fig. 7. (a) Histogram showing the frequency of graded 'pre' assembly ages from the current database (in 5 Ma bins). (b) Histogram showing the frequency of graded 'post' assembly ages in the current database.

conclusion, based on both new and existing age and geochemical data, is consistent with the idea that most of the juvenile crust in the ANS region formed in intraoceanic convergent margin settings. Amalgamation of these terranes began as early as 800 Ma and continued to about 620 Ma (Stern and Abdelsam, 1998; Cosca et al., 1999). A detailed analysis of each of the many arc-ophiolitic terranes in the Arabian–Nubian shield can be found in the references listed at the beginning of this section and will not be repeated here.

In addition to the data cited in Stern (1994), there are a number of new studies that provide additional geochronologic constraints on the development of the Arabian–Nubian shield and all paint a consistent picture of the tectonic history. Cosca et al. (1999) document a single collisional event in the Elat area (southern Israel) at 620 ± 10 Ma. This was followed by unroofing, extensional collapse and intrusion of postorogenic grantoids (post-600 Ma). The authors considered these ages to document final collision of east and west Gondwana at ca. 620 Ma followed rapidly by extensional collapse or tectonic escape in the EAO. Beyth and Heimann (1999) report mean K–Ar ages of doleritic dikes in the Mt. Timna region (Israel) of the ANS between 548 and 509 Ma. These dikes are the youngest igneous rocks below the basal Cambrian unconformity. A ^{40}Ar/^{39}Ar plateau age of 531.7 ± 4.6 was obtained from one of the Mt. Timna dikes and is considered to be the age of the youngest postcollision extensional magmatism in the ANS. Emplacement of these dikes followed an earlier period of posttectonic alkaline magmatism in the same region dated to 610 Ma (Beyth et al., 1994).

Loizenbaur et al. (2001) report new single zircon age data from the Meatiq metamorphic core complex in eastern Egypt. According to their interpretation, this complex underwent rifting between 800 and 700 Ma, convergence between 660 and 620 Ma and extensional tectonism with the emplacement of granitoids between 620 and 580 Ma consistent with the Wilson-cycle hypothesis of Stern (1994).

Blasband et al. (2000) provide an recent compilation of geochronology and tectonic settings for the ANS. They consider that island arcs and ophiolitic fragments formed during the interval from ~ 740 to 900 Ma and were accreted by about 650 Ma. Gneissic domes, considered by Blasband et al. (2000) to represent metamorphic core complexes, formed during extensional collapse of the EAO between 620 and 530 Ma. Fig. 8 shows a generalized tectonic history for the ANS and the data are consistent with consolidation of the region between 750 and 620 Ma presumably marking the first stage of the assembly of eastern Africa and the elements of eastern Gondwana.

An important consideration in any tectonic model is the timing of ‘escape’ tectonics in the Arabian–Nubian sector. Extrusion tectonics in the EAO appears to be restricted to the northern part of the orogen, suggesting that oblique collision to the south resulted in a northerly ‘free-face.’ Motion along the Najd fault is interpreted to provide a minimum age for continent–continent collision further south at 624.9 ± 4.2 Ma and possible cessation or transfer of motion to other regions of the Najd system at 576.6 ± 5.3 Ma (Stacey and Agar, 1985; Kusky and Matesh, 1999).

Fig. 9a and b shows the frequency distribution of all/syn-assembly ages for the eastern part of Africa including the ANS, and Fig. 9c shows the areal distribution of sampling localities.

2.3. Somalia–Ethiopia–Eritrea–Sudan–Yemen

Progress in geochronologic, geologic and isotopic studies in this important region of eastern Africa is hindered by a host of sociopolitical problems. Nevertheless, significant progress in understanding this region has been made in the past decade. The transition between the juvenile domains of the ANS sector and older continental crust of the African craton is located in this region and several of the studies outlined below provide constraints on the location of suture(s) within the area.

Ayalew et al. (1990) report ages of 828 and 814 Ma on ‘prekinematic’ plutons in the Birbir domain in western Ethiopia. Anatectic melts within the high-grade Baro domain yielded an age of 780 + 19/ – 14. Rb–Sr dating of one of the syn-tectonic plutons yielded a whole-rock isochron age of 759 ± 18 Ma. These ages were considered to date a regional metamorphic event in western Ethiopia coeval with low-grade metamorphism of arc-related rocks to the east in the Arabian–Nubian shield. Reset Rb–Sr ages from the western Ethiopian shield cluster around 630 Ma and are interpreted by the authors as dating the development of a major transcurrent fault system similar to the timing of motion suggested along the Najd fault in Saudi Arabia (see above). A second amphibolite-grade metamorphic event in the Baro domain is dated by a lower concordia intercept age of zircons from anatectic melts at 582 + 29/ – 33 Ma.

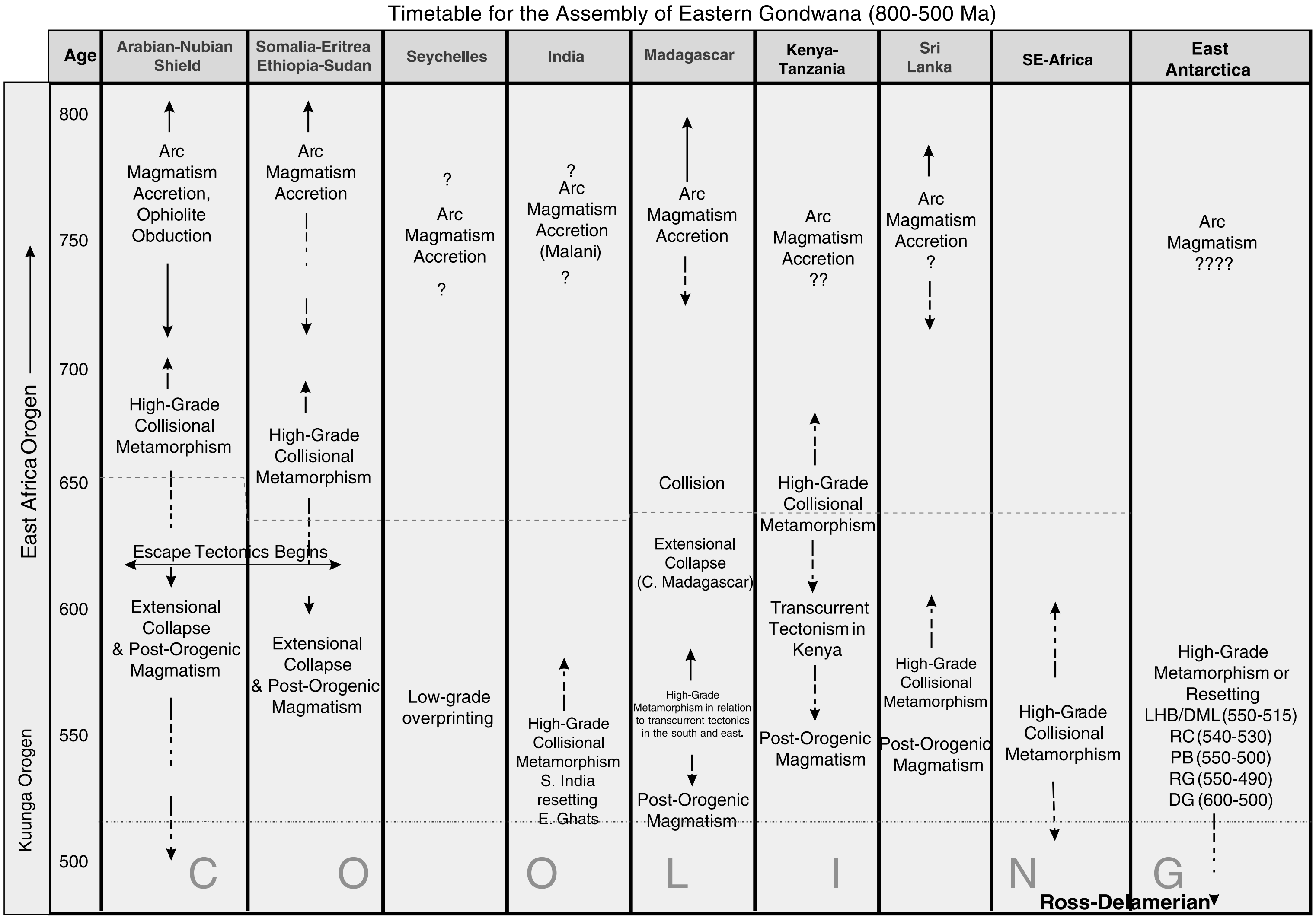


Fig. 8. Schematic summary of the tectonic events within the different regions of eastern Gondwana discussed in this paper.

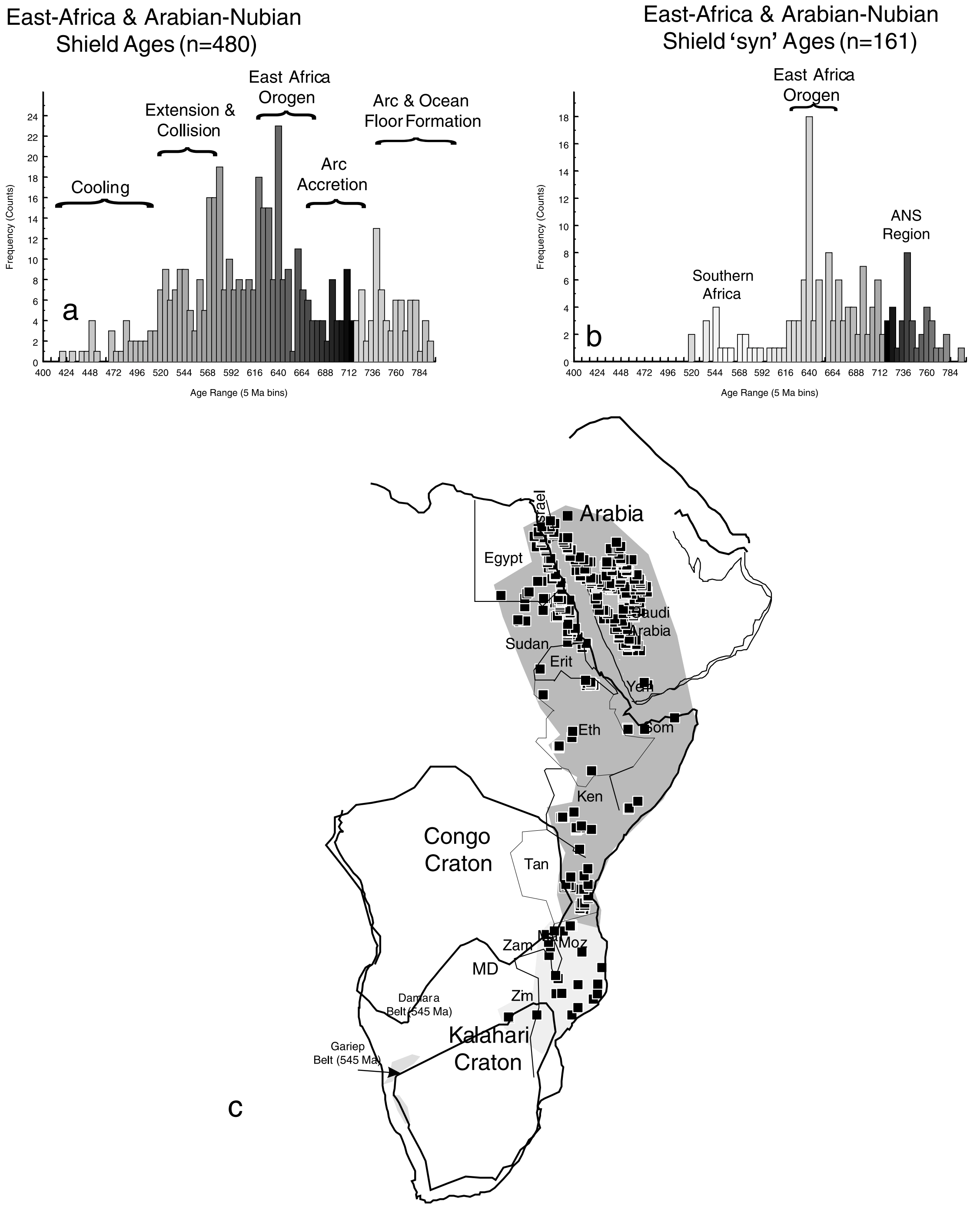


Fig. 9. (a) Histogram of all 'C' and better ages (5 Ma bins) from East Africa and the Arabian–Nubian shield region along with the interpreted major orogenic events in that region. (b) Histogram of 'syn' assembly ages from the same region. (c) Geographic distribution of geochronologic studies used in the analysis. Dark shading refers to regions with 'syn' assembly ages older than 620 Ma and lighter shading refers to regions with 'syn' assembly ages younger than 600 Ma.

Posttectonic plutonism in the region took place after 550 Ma.

Mock et al. (1999) conducted ^{40}Ar/^{39}Ar thermochronology on Ethiopian and Yemeni basement rocks in order to elucidate the thermal effects of the younger Afar plume. In western Ethiopia, post to late syntectonic rocks gave equivocal results; however, muscovite from a leucogranite gave a plateau age of 594.0 ± 10.7 Ma and an augen gneiss sample yielded a biotite plateau age of 564.5 ± 10.2 Ma. Results from western Ethiopia are consistent with the results of Ayalew et al. (1990); however, as the Ar–Ar results are from minerals with blocking temperatures between 300–450 °C (McDougall and Harrison, 1999), the age of the younger metamorphic event is likely to be older ~ 600 Ma. In northern Ethiopia, post or late tectonic granitoids gave plateau ages between 600 ± 11 Ma (biotite) and 663.7 ± 14 Ma (muscovite). Tonalitic samples gave overlapping biotite plateau ages of 665.8 ± 13.0 and 647.2 ± 13.0 Ma. High-grade metamorphic samples showed disturbed biotite and K-spar age spectra, but were interpreted to reflect cooling between ca. 736 and 510 Ma. Southern Ethiopian samples yielded muscovite plateau ages between 518 and 543 Ma. Two Yemeni basement metamorphic samples showed K-spar age gradients with maxima ranging between 540 and 565 Ma. Mock et al. (1999) considered all of the ^{40}Ar/^{39}Ar ages to reflect diachronous cooling from north to south following crustal thickening or denudation following a single orogenic episode older than 600 Ma.

Whitehouse et al. (1998, 2001) documented basement correlations between Yemen, Saudi Arabia and Somalia. Yemen contains some Archean-age crust sandwiched between a collage of arc terranes in the Arabian–Nubian shield region and high-grade polycyclic gneissic rocks of the East Africa Orogen to the south and west. Windley et al. (1996) noted that following the accretion of the Arabian arc terranes (~ 750 Ma) and the formation of the Nabitah suture (680–640 Ma), subduction took place beneath the Afif terrane of Yemen and several arc and continental fragments were welded to the Afif terrane during the latter stages of the East African Orogeny (< 640 Ma).

Recent work in the Axum area of northern Ethiopia (Tadesse et al., 2000) also yielded ages for arc-related magmatism between 750 and 800 Ma and postorogenic magmatism at ~ 550 Ma (based on Sm–Nd, Rb–Sr and Th–U–Pb dating). These ages are broadly similar to those in the ANS sector and Somalia.

Ages of felsic volcanics and granitoid intrusions between 850 and 811 Ma (Pb–Pb evaporation) are found in the Nafka terrane of Eritrea (Teklay, 1997). Tadesse et al. (2000) argued that the consistency in ages from Ethiopia and Eritrea indicated a similar arc setting between 750 and 850 Ma. Geochemical data from the Adobha belt of northern Eritrea (Woldemhaimanot, 2000) indicate an island-arc/ophiolitic/island-arc assemblage presumably juxtaposed during the arc-accretion phase of the East Africa Orogen (see also Beyth et al., 1997).

Kröner and Sassi (1996) dated the intrusion of 840–720 Ma gabbroic and granitoid magmas (Pb–Pb zircon) into Meso to Paleo Proterozoic crust (1400–1820 Ma) in northern Somalia and considered the northern Somali basement as part of a continental terrane sandwiched between arc terranes. This conclusion is supported by the subsequent work of Teklay et al. (1998), who examined the ages and geochemistry of granitoid intrusions in southern (cratonic Africa) and eastern (ANS arc terranes) Ethiopia and concluded that the granites formed along a boundary between the ANS terranes and this continental block. Kröner et al. (2000a,b) tentatively correlated this intrusive event with similar-aged intrusions in Central Madagascar (see below) possibly related to a continental arc (see also Handke et al., 1999). These ages are consistent with those reported by Ayalew et al. (1990) for western Ethiopia. This work echoes that of Whitehouse et al. (2001), who argue that the arc-gneiss terranes of Yemen pass along strike into the purely continental blocks of the East Africa Orogen in Ethiopia.

The Delgo basement area of northern Sudan contains a microcosm of the tectonic history of northeastern Africa (Harms et al., 1994). There is evidence for oceanic floor forming in an arc-type setting along with arc-related magmatism in the 650–800 Ma interval. A younger limit for high-grade metamorphism is placed at 600 Ma and was followed by the emplacement of posttectonic granites in an extensional setting at 550 Ma.

Extensional collapse and postorogenic magmatism in the Somalia–Eritrea–Ethiopia–Sudan region is only poorly known, but the extant ages (Tadesse et al., 2000) are entirely consistent (Figs. 8–10) with the extensional tectonic setting described by Blasband et al. (2000) and Cosca et al. (1999).

2.3.1. Kenya–Tanzania–Congo Cratonic margin

The timing of orogenic events in northeast Africa is well constrained; however, one region of the EAO that contains many of the high-grade metamorphic rocks (Kenya) is poorly known. Most of the ages from Kenya (Key et al., 1989) provide only loose constraints on the collisional history. Key et al. (1989) reported a number of Rb–Sr and K–Ar ages from north-central Kenya and used these data to define four major tectonothermal events in Kenya. The oldest of these at ~ 820 Ma was attributed to continent–continent collision between east and west Gondwana, but this age was based on a Rb–Sr errorchron whose significance is debatable. Younger, ~ 620, ~ 570 and ~ 530 Ma events were considered to date postcollisional episodes in Kenya. In contrast, Stern (1994) considered the ages compatible with the thesis of terminal collision in the Kenyan sector of the EAO at ~ 650 Ma as did Meert and Van der Voo (1997). Late postcollisional extension is also evidenced in Kenya. A series of mafic dikes intrudes granulite-facies metamorphic rocks and crosscuts regional high-grade shear zones (Key et al., 1989; Meert and Van der Voo, 1996). Meert and Van der Voo (1996) obtained an age of 547 ± 4 Ma (^{40}Ar/^{39}Ar, biotite) for one of these mafic dikes in Kenya (Sinyai dolerite) similar to ages obtained by Key et al. (1989).

Mosley (1993) highlighted the importance of placing proper age control on the history of tectonism in Kenya as it relates to Gondwana assembly. The metamorphosed sedimentary rocks in Kenya may represent the former passive margin sequence formed adjacent to the Congo craton and there are a number of dismembered ophiolites (Shackleton, 1986) that might represent the former Mozambique Ocean. It was suggested that Kenya marks a transitional region between the main continental collision (southern and central parts of the country) to continental escape tectonics and arc collisions (northeastern Kenya and adjacent region of Ethiopia; Mosley, 1993). The age of the oldest metamorphic event in Kenya that reached amphibolite–granulite facies metamorphism is not known with any certainty, but the 630 to 550 Ma age Baragoin–Barsaloin of Key et al. (1989) involved extensive transcurrent shearing that is approximately time-equivalent to that seen in Madagascar (see below). This observation, coupled with the new geochronologic data from northeastern Tanzania immediately adjacent to Kenya suggest that metamorphism related to continent–continent collision was coeval at ~ 640 Ma.

The age constraints on high-grade metamorphism and continental collision in Tanzania can be tentatively correlated with those to the immediate north in Kenya. Coolen et al. (1982) reported a U–Pb zircon lower intercept age of 652 ± 10 Ma for granulite-facies rocks of the Furua Complex in southern Tanzania. Subsequent K–Ar hornblende dating of the Furua granulites yielded ages ranging from 614 to 655 Ma (Andriessen et al., 1985). Maboko et al. (1985) obtained U–Pb ages from the Wami River granulites (N. Tanzania) with upper and lower intercepts at 714 and 538 Ma, respectively. These ages are based on bulk zircon analyses of mixed samples; however, as noted by Muhongo et al. (2001), the analyses were likely influenced by mixing of metamorphic and igneous zircons and overestimates the actual time of granulite-facies metamorphism. Collectively, these earlier studies concluded that the main phase of granulite formation occurred at ~ 715 Ma followed by a slightly younger ~ 650 Ma amphibolite-facies event in Tanzania.

Maboko et al. (1989) obtained ^{40}Ar/^{39}Ar hornblende ages from the Uluguru granulites of ~ 630 Ma and much younger muscovite and K-feldspar ages. Muhongo and Lenoir (1994) reported U–Pb zircon ages from the Uluguru Complex of 695 ± 4 Ma and from the Pare Mountains 645 ± 10 Ma. Maboko and Nakamura (1995) also conducted Sm–Nd garnet dating on the same Uluguru granulites and obtained ages of 633 ± 7 and 618 ± 16 Ma. Maboko and Nakamura (1995) suggested that these ages reflected extremely slow-cooling following continent–continent collision at 695–715 Ma. A recent study by Maboko (2000) yielded a number of whole-rock–biotite Rb–Sr ages from granulite and amphibolite-facies rocks in Tanzania. The ages range from ~ 648 to ~ 490 Ma and are interpreted as diachronous cooling ages across the Mozambique belt in Tanzania. Maboko (2000) suggested that both the granulite and amphibolite deformational events were related to a single continent–continent collision at ≥ 650 Ma and that the range in ages reflected differences in depth of burial and, therefore, timing of exhumation and cooling.

Two recent papers on granulites from Tanzania by Muhongo et al. (2001) and Möller et al. (2000) tightly constrain the timing of metamorphism in Tanzania (and by inference, in Kenya). Muhongo et al. (2001) report Pb–Pb evaporation and SHRIMP zircon ages from garnet–sillimanite gneisses in the Wami River area with an age of 641.2 ± 0.9 Ma that are interpreted as dating the peak of granulite-facies metamorphism. Metamorphic zircons from the Uluguru region of Tanzania yielded similar ages of 642.3 ± 0.9, 642 ± 5 and 638.8 ± 1.0 Ma. A population of igneous zircons from an orthopyroxene granulite was analyzed on SHRIMP and yielded a nearly concordant age of 725 ± 14 Ma, and a group of metamorphic grains from the same rock yielded concordant ages of 639 ± 13 Ma. Möller et al. (2000) report monazite ages from enderbite and metapelites in the northern and eastern Uluguru region between 653 and 625 Ma and a concordant zircon fraction from the enderbite produced a minimum age of 626 ± 2 Ma.

Metamorphic monazite ages from the Usambara mountains in Tanzania (Möller et al., 2000) ranged between 625 and 630 Ma. Muhongo et al. (2001) analyzed granulite-facies metasediments from the same region and obtained Pb–Pb evaporation ages of 641.4 ± 0.9 and 641.1 ± 0.9 Ma.

Cooling ages based on U–Pb/Pb–Pb dating of rutile in the Pare-Usumbara and Uluguru mountains range between 500 and 550 Ma (Möller et al., 2000). According to Möller et al. (2000), these rutile ages reflect cooling immediately following ca. 550 Ma collision in the region (see Section 5).

A possible extension of the Congo craton to the north of Kenya and Sudan is described by Sultan et al. (1994) based on ages from a gneissic terrane west of the Nile in southern Egypt. The region contains evidence of Archean crust and led the authors to suggest that the Uweinat region represents a continuation of the Congo craton to the northeast. The youngest metamorphic event in this region is dated to 604 ± 5 Ma (U–Pb multi grain sphene, Gabel El Asr) and a series of granitic gneisses yielded intrusive ages between ~ 743 and 626 Ma.

Collectively, the data from the regions at the eastern margin of the Congo craton are compatible with the idea of continent–continent collision at ~ 640 Ma followed by the intrusion of posttectonic granitoids and extensional collapse sometime after 600 Ma (Figs. 8–10). Additional evidence for igneous activity between 843 and 665 Ma suggest that at least part of the EAO in Tanzania region originated as a convergent margin during this interval (Muhongo et al., 2001; Kröner et al., 2000a,b). This earlier episode of arc magmatism is similar in age and style to some units in Madagascar (Muhongo et al., 2001; Kröner et al., 2000a,b).

The conclusion that continent–continent collision occurred at ~ 640 Ma is in direct contrast to that reached by Appel et al. (1998) and Möller et al. (2000). They report isobaric slow cooling throughout much of the Tanzania region and concluded that the earlier granulite-facies metamorphism in the region resulted from magmatic underplating preceding continent–continent collision at 550 Ma. The conclusion of Appel et al. (1998) and Möller et al. (2000) that continent–continent collision in the region did not occur until ~ 550 Ma begs the question as to what collided with the Kenya and Tanzania region since postcollisional extension and shearing is well underway in Kenya and Madagascar by 550 Ma and escape tectonics in the ANS region began at ~ 630 Ma (Key et al., 1989; Meert and Van der Voo, 1996; Martelat et al., 2000; Nédélec et al., 2000). Furthermore, slightly metamorphosed dikes that crosscut regional high-grade fabrics in Kenya are dated to 550 Ma, suggesting that granulite-facies metamorphism was earlier than 550 Ma (Meert and Van der Voo, 1996; Key et al., 1989). An alternative explanation is that magmatic underplating was caused by lithospheric delamination beneath a collision thickened crust and subsequent intrusion of asthenospheric mantle (Platt and England, 1994).

2.3.2. NE-Zimbabwe–Malawi–Mozambique

Pinna et al. (1993) obtained a number of Rb–Sr ages from Precambrian rocks of northern Mozambique. They concluded that the main phase of ‘Mozambican’ orogenesis occurred between 1100 and 850 Ma as did Wilson et al. (1993). Younger, ca. 538 Ma ages were inferred to date a younger thrusting and shearing episode in an intracontinental environment. This interpretation is broadly consistent with the interpretation of Key et al. (1989) for deformational events in

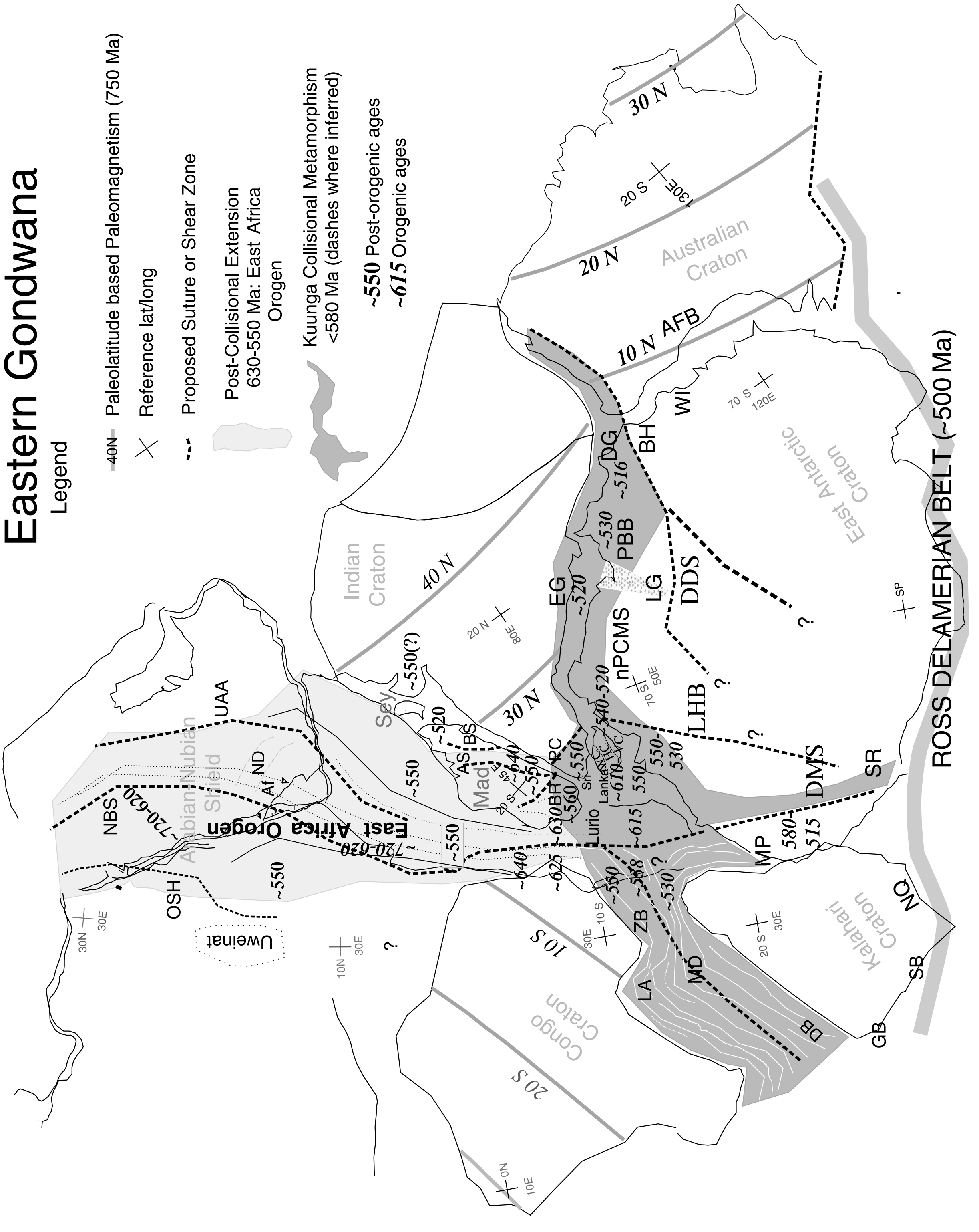
Kenya, but seems difficult to reconcile with an increasing number of U–Pb zircon ages in Tanzania (see above) and the few new ages from Mozambique–Malawi–Zimbabwe described below. As noted by Goscombe et al. (1998), this older 800–900 Ma orogenesis may be unrelated to the formation of eastern Gondwana.

Vinyu et al. (1999) report new U–Pb titanite–zircon ages along with $^{40}Ar/^{39}Ar$ ages from the easternmost part of the Zambezi belt in NE-Zimbabwe. These data are cast in a regional tectonic framework and the authors argue for extensional magmatism at ca. 800 Ma followed by amphibolite-grade metamorphism at 535 Ma. Goscombe et al. (1998) provide geochronologic and thermobarometric data from the Chewore inliers (northernmost Zimbabwe) and concluded that the older granulite terranes in the region were partially reworked in the interval from 510 to 560 Ma (peak 538 ± 15 Ma). Metamorphic zircon overgrowths in the granulite terrane yielded a SHRIMP U–Pb age of 524 ± 16 Ma (Goscombe et al., 1998). Hanson et al. (1993) note that the style of tectonic activity to the east of the Mwembeshi dislocation in central Zambia (~ 14°S, 28°E, Fig. 10) is distinct from that in the Damara and Lufilian arc region (Fig. 10) although they are temporally correlated. The Damara–Gariep Orogenic belts both reached peak metamorphic conditions at ~ 545 Ma (Frimmel et al., 1996; Rozendal et al., 1999), suggesting that tectonic activity throughout the Zambezi–Lufilian–Damara–Gariep belts was related to the final stages of Gondwana assembly.

In northern Mozambique, high-grade metamorphism is dated to ~ 615 Ma by Pb–Pb, conventional zircon and SHRIMP methods (Kröner et al., 1997). There are limited reliable Neoproterozoic geochronologic data from Malawi on syn-tectonic rocks. A syn-tectonic syenite from Songwe yielded a Rb–Sr whole-rock age of 671 ± 62 Ma (Ray, 1974) that was considered equivalent to the Mbozi syenite in southern Tanzania (755 ± 25 Ma, biotite K–Ar, recalculated from Brock, 1968). Mineral Rb–Sr ages on the Lwakwa granite yield 599 ± 6 Ma (biotite) and 560 ± 6 Ma (feldspar) that were considered to date cooling of the granite (Ray, 1974). These cooling ages are slightly younger than those found in the Furua region of Tanzania to the immediate north (Andriessen et al., 1985). Definite posttectonic granites from the Malawi granite province yield Rb–Sr whole-rock ages between 443 ± 30 and 489 ± 14 Ma (Haslam et al., 1983). $^{40}Ar/^{39}Ar$ cooling ages from the posttectonic Ntonya ring structure range from 471.5 ± 7 Ma (biotite) to 510.0 ± 7 Ma on hornblende (Briden et al., 1993) that are consistent with the Rb–Sr results of Haslam et al. (1983).

Kröner et al. (2000b) report new zircon ages on granitoid gneisses and charnockites from southern Malawi. These yield protolith ages ranging from 1040 to 580 Ma and metamorphic ages in the range of 550–560 Ma. Kröner et al. (2000b) note the apparent correlation between the younger (~ 550 Ma) ages in Malawi with those in southern Madagascar, Sri Lanka and southern India (see also Fig. 8–10 and Section 5). Geochemical and thermobarometric ($P–T$) data from the protoliths led Kröner et al. (2000b) to suggest that Malawi, together with Tanzania and other regions of eastern Gondwana, behaved as an active continental margin with diachronous accretion of terranes during the 100 million year interval from ~ 640 to ~ 530 Ma. However, the $P–T$ evolution of continent–continent collision zones is not always straightforward and the direction of $P–T$ path

Fig. 10. Geographic distribution of age provinces within eastern Gondwana showing the regions of 620–550 Ma postcollisional extension (light shading), 570–530 Kuunga collisional metamorphism (darker shading). Abbreviations are: **Af** = Afif Terrane (Arabian–Nubian Shield); **AFB** = Albany–Fraser Belt (Australia); **BH** = Bunger Hills (Antarctica); **BR** = Bongolava–Ranotsara shear zone (Madagascar); **BS** = Betsimisaraka suture zone (Madagascar); **DB** = Damara Belt (Africa); **DDS** = Darling–Denman suture (Antarctica–Australia); **DG** = Denman Glacier (Antarctica); **DMS** = Dronning Maud suture (Antarctica); **EG** = Eastern Ghats (India); **GB** = Gariep Belt (SW Africa); **HC** = Highland Complex (Sri Lanka); **LA** = Lufilian Arc (Africa); **LG** = Lambert Graben (Antarctica); **LHB** = Lützow–Holm Belt (Antarctica); **MD** = Mwembeshi Dislocation (Africa); **MP** = Maud Province (Antarctica); **NBS** = Nabitah Suture (Arabia); **NC** = Napier Complex (Antarctica); **ND** = Najd fault; **nPCSM** = Northern Prince Charles Mountains (Antarctica); **NQ** = Namaqua Belt (Africa); **OSH** = Onib–Sol Hamed suture (Arabia); **PBB** = Prydz Bay Belt (Antarctica); **PC** = Paughat–Cauvery (India); **SB** = Saldania Belt (SW Africa); **SR** = Shackelton Range; **UAA** = Urd Al Amar suture (Arabia); **VC** = Vijayan Complex (Sri Lanka); **WC** = Wanni Complex (Sri Lanka); **WI** = Windmill Islands (Antarctica); **ZB** = Zambezi Belt. The 500 Ma Ross–Delamerian Orogen followed the final suturing of Gondwana elements.


Eastern Gondwana
Legend
40N Paleolatitude based Paleomagnetism (750 Ma)
Reference lat/long
Proposed Suture or Shear Zone
Post-Collisional Extension 630-550 Ma: East Africa Orogen
Kuunga Collisional Metamorphism <580 Ma (dashes where inferred)
~550 Post-orogenic ages
~615 Orogenic ages
Australian Craton
East Antarctic Craton
Indian Craton
Congo Craton
Kalahari Craton
Arabian-Nubian Shield
East Africa Orogen
ROSS DELAMERIAN BELT (~500 Ma)
30 N
20 N
10 N
40 N
30 N
10 S
20 S
AFB
WI
BH
DG
PBB
LG
DDS
EG
nPCMS
LHB
SR
DMS
MP
NQ
SB
GB
DB
MD
LA
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580-515
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30N 30E
10N 30E
0N 10E
10 S 30E
20 S 30E
20 S 45E
20 N 80E
70 S 50E
70 S 120E
20 S 130E
SP
?

may depend on the location within the orogenic belt. For example, Goscombe and Hand (2000) report both clockwise and counterclockwise $P-T$ paths in a Himalayan paired continent–continent collision metamorphic belt. It is possible that Malawi (together with points south) remained an active continental margin until ~ 570 Ma, but collision in Tanzania occurred earlier (~ 640 Ma) as discussed below.

2.3.3. Age distribution and tectonic events in eastern Africa

Fig. 9a shows the frequency–age distribution for eastern Africa and the Arabian–Nubian shield region ($n = 480$). There is a clear peak in between 615 and 650 Ma and a smaller peak between 560 and 600 Ma although the younger segment is not apparent in the frequency distribution graph of 'syn' assembly ages (Fig. 9b). Fig. 10 shows the spatial significance of these ages in the broader eastern Gondwana assembly process. The most important conclusion is that the ages from southern Tanzania northward suggest that oblique continent–continent collision occurred at 640 ± 20 Ma and was followed by extensional collapse beginning sometime after ~ 600 Ma. This oblique collision resulted in tectonic escape along the northern free face (ANS region) of the EAO beginning around 630 Ma and continuing to about 550 Ma. The makeup of the colliding blocks is difficult to evaluate, but likely include Madagascar, Sri Lanka, Somalia, India and part of East Antarctica. Unfortunately, some of these regions are overprinted by younger orogenies and the exact boundaries between continental blocks are difficult to unravel (Fig. 10). The region south of Tanzania and northern Mozambique shows no indication of the ~ 615 to 650 Ma orogenesis, but does show evidence of a regional tectonothermal event during the interval from 570 to 520 Ma consistent with metamorphic ages in the Dronning Maud Province of East Antarctica discussed below. High-grade metamorphism in the Damara and Gariep belts of southwestern Africa also occurs during this same interval and may be partly related to collisional events to the east.

2.4. Seychelles

Torsvik et al. (2001a) and Tucker et al. (2001) provide an up-to-date picture of the tectonic setting of the Seychelles (see Fig. 8 summary). The islands are dominated by granitoids with subordinate mafic dykes. Previous geochronology using Rb–Sr gave a range of ages between 570 and 713 Ma (Suwa et al., 1994). Recent U–Pb zircon–baddelyite geochronology on the granitoid rocks yielded ages between 748 and 764 Ma (Stephens et al., 1997; Tucker et al., 1999a,b). Basaltic dykes intruding one of the granites (Takamaka dike) yielded a U–Pb zircon age of 750 Ma (Torsvik et al., 2001a). Stephens et al. (1997) argued that the Seychelles magmatism resulted from extensional tectonism during the purported breakup of a supercontinent between 703 and 755 Ma.

New ages from the Seychelles further define the tectonic setting and duration of magmatism in the Seychelles (Tucker et al., 2001). An arithmetic mean age based on 11 samples of granitic rocks and 1 sample of a mafic dyke is 751.6 ± 3.3 Ma (all ages are U–Pb ages). Tucker et al. (2001) conclude that the majority of the rocks were emplaced between 750 and 755 Ma; however, three samples yielded older ages from 758 to 808 Ma. Combined with the younger age of 703 Ma reported by Stephens et al. (1997), the Seychelles documents an ~ 100 Ma duration of magmatic activity. Tucker et al. (2001) and Ashwal et al. (2002) argue that the geochemistry of the Seychelles rocks, the coexistence of mafic and silicic magmatism and the duration of magmatic activity are consistent with an active (Andean-type) setting for the Seychelles.

2.5. India

Reliable geochronologic data from India during the period from 800 to 500 Ma are sparse and all ages are represented in the histogram (Figs. 8, 10 and 11). In the northern and western regions of India, there is little hint of Neoproterozoic orogenesis with the exception of a ^{40}Ar/^{39}Ar 'thermal disturbance' dated between 550 and 500 Ma for the Jalore granites and Malani rhyolites of Rajasthan (Rathore et al., 1999). These authors also reported a variety of Rb–Sr ages on granitic and rhyolitic rocks from Rajasthan and argued for several pulses of magmatic activity between 779 and 683 Ma. New U–Pb ages from the Malani rhyolites and the Jalore granite (Torsvik et al., 2001b) range between 748 and 767 Ma. It appears that these represent a single pulse of granitic magmatism

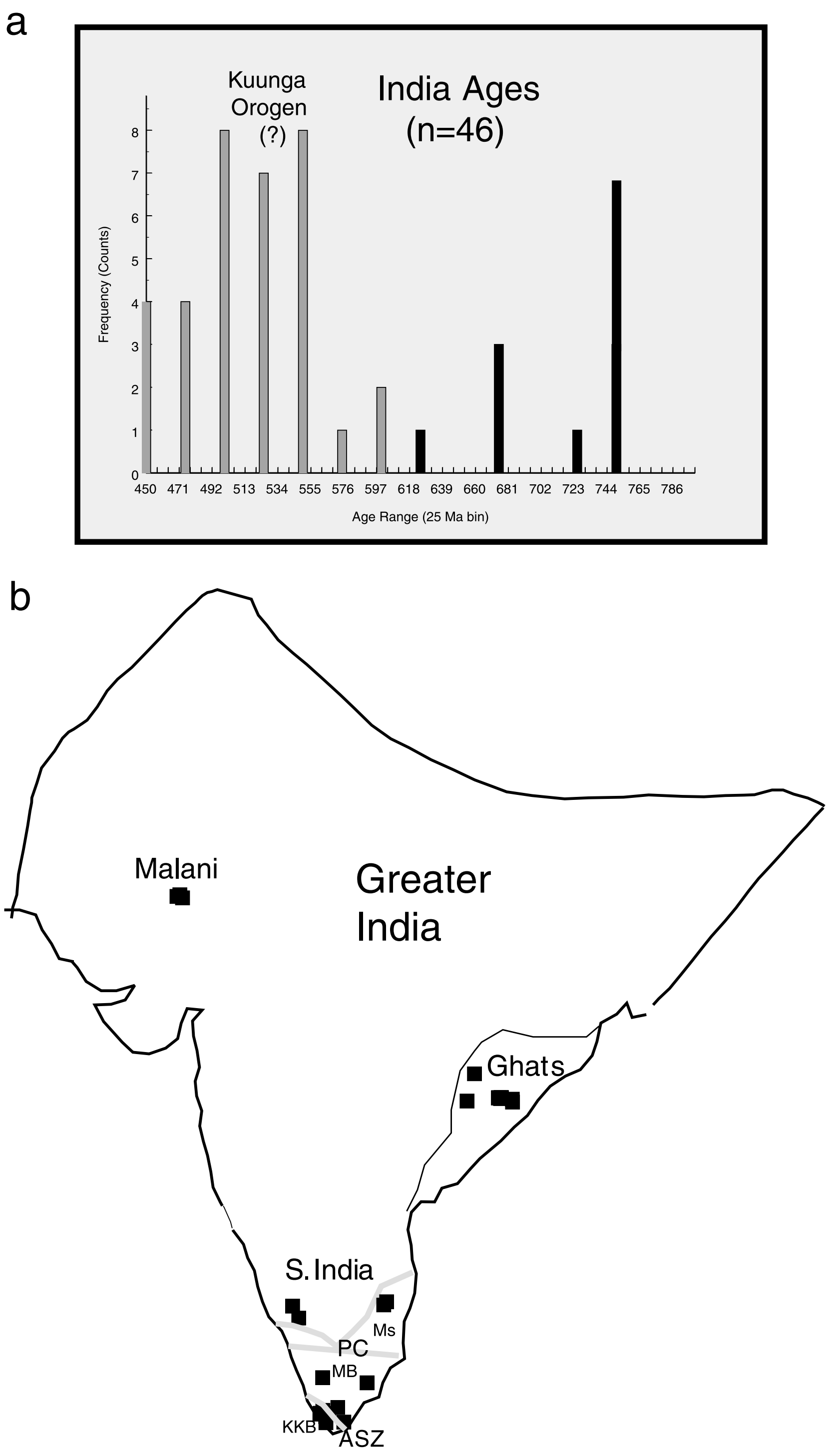


Fig. 11. (a) Histogram of ages (25 Ma bins) from India along with the major orogenic events in that region. (b) Geographic distribution of geochronologic studies used in the analysis. PC = Palghat–Cauvery shear zone; ASZ = Ankankovil shear zone; MB = Madurai block; KKB = Kerala–Khondalite belt; MB = Madurai Block; Ms = Madras block.

in the Rajasthan region that is consistent with a continental arc setting (Torsvik et al., 2001b) although previous interpretations favored an extensional setting (Bushan, 2000). There is some evidence for a similar-age magmatic episode in southern India (Santosh et al., 1989). They report Rb–Sr whole-rock ages of 638 ± 24 Ma for the Angadimogar syenite and 750 ± 40 for the Peralimala alkaline granites north of Trivandrum. Miyazaki et al. (2000) report similar Rb–Sr whole-rock ages for the Yelagiri syenite (757 ± 32 Ma) and the Sevattur syenite (756 ± 11 Ma) to the northeast of the Trivandrum samples. These newer ages on the Sevattur syenite are consistent with previously published ages of 767 ± 8 Ma (Rb–Sr whole rock), 771 ± 8 Ma (Rb–Sr whole-rock–mineral age) and 773 ± 18 Ma (Rb–Sr whole-rock–mineral age) on the Sevattur syenite, carbonatite and pyroxenite, respectively (Kumar et al., 1998; Kumar and Gopolan, 1991). All of the ca. 750 Ma magmatism in the southern Indian shield is attributed to extensional tectonism by Santosh et al. (1989).

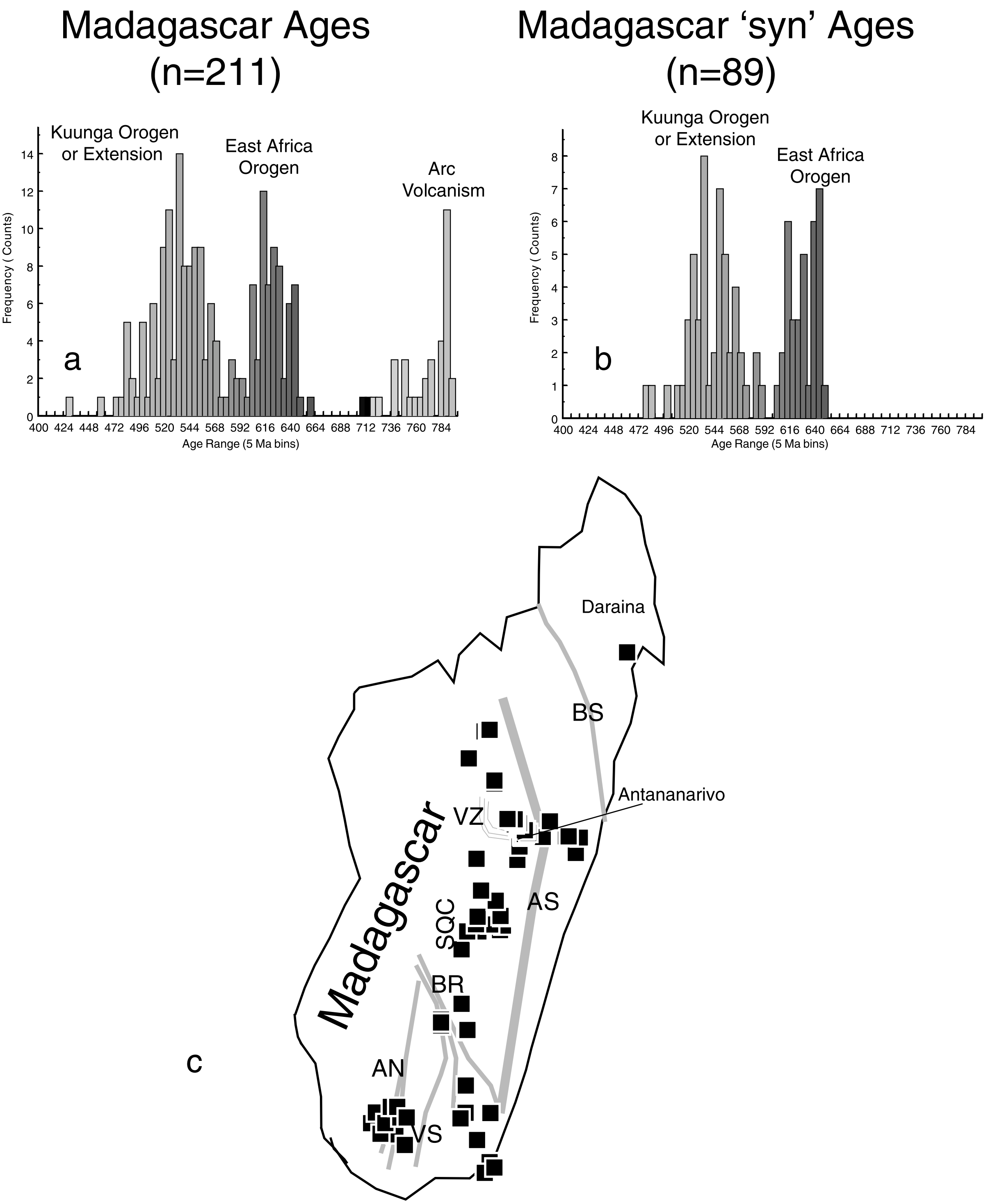
Hansen et al. (1985) reported a whole-rock Rb–Sr age from amphibolite-facies migmatites near Madurai of 550 ± 15 Ma. Bühl (1987, unpublished PhD thesis) reported a 552 Ma U–Pb monazite age from the Ponmundi quarry north of Trivandrum. Electron microprobe dating of monazites by Bindu et al. (1998) and Braun et al. (1998) from the same region indicated ages ranging from 520 to 605 Ma. Bindu et al. (1998) related these ages to a thermal overprinting, whereas Braun et al. (1998) considered the ages to date peak metamorphism in the region during Gondwana assembly. Choudhary et al. (1992) reported Sm–Nd ages on garnets from gneisses and granulites from Ponmundi and also concluded that high-grade metamorphism occurred at ca. 558 Ma and was followed by uplift and cooling between 440 and 460 Ma. Santosh et al. (1992) dated charnockitic rocks at Nellikala using Sm–Nd whole-rock mineral systematics at 539 ± 20 Ma. Unnikrishnan-Warrier et al. (1995) used the same technique to date a charnockite from Kottaram at 517 ± 26 Ma and a Rb–Sr whole-rock–mineral isochron of 484 ± 15 Ma. A U–Pb zircon age on the Putetti alkaline syenite body of Tamil Nadu in southern India yielded a well-defined upper intercept age of 572 ± 2 Ma (Kovach et al., 1998). Fonarev and Konilov (1999) report metamorphic events in the Karnataka craton and Niligiri blocks of southern India, with the youngest between 580 and 550 Ma.

Other hints of a late Neoproterozoic to early Paleozoic metamorphic event are provided by recent work in the eastern Ghats province (Mezger and Cosca, 1999). A concordant zircon from an undeformed apatite–magnetite vein yielded an age of 516 ± 1 Ma. A combined regression on partially reset sphenes resulted in a poorly defined lower intercept age of 504 ± 20 Ma. Hornblende ^{40}Ar/^{39}Ar ages from the eastern Ghats were also reset between 573 and 650 Ma (Mezger and Cosca, 1999) leading to the suggestion that an amphibolite-grade metamorphic event occurred in the region between 500 and 550 Ma. Shaw et al. (1997) also identified this ~ 550 Ma metamorphic event in zircons from the eastern Ghats and attributed it to a subsequent heating or fluid flux event. Sarkar and Paul (1998) indicate that this younger thermal event is not associated with any deformation in the eastern Ghats and conclude that it is similar to the effects noted in the Rayner Complex of East Antarctica (Tingey, 1991). Collectively, the data from the eastern Ghats suggest relatively minor effects associated with the younger tectonic (560–530 Ma) episode in eastern Gondwana. It is clear that additional work is needed in southern India to fully constrain the type and timing of orogenic activity in that region.

2.6. Madagascar

The island of Madagascar is located in a key position within the assembled Gondwana supercontinent (Figs. 2 and 9). Geochronologic work in the past decade has provided important constraints on the evolution of the island (Paquette et al., 1994; Kröner

Fig. 12. (a) Histogram of all 'C' and better ages (5 Ma bins) from Madagascar along with the major orogenic events in that region. (b) Histogram of 'syn' assembly ages from Madagascar. (c) Geographic distribution of geochronologic studies used in the analysis. Light shadings are shear zones within Madagascar. Shear zones (or proposed tectonic boundaries) in Madagascar abbreviated as: AN = Anosyan shear zone; AS = Angavo shear belt; BR = Bongolava–Ranotsara shear zone; BS = Betsimisaraka suture; SQC = 'Itremo Group'; VS: Vohibory shear zone; VZ: Antananarivo virgation zone.


Madagascar Ages
(n=211)
Kuunga Orogen
or Extension
East Africa
Orogen
Arc
Volcanism
a
Frequency (Counts)
0
2
4
6
8
10
12
14
400 424 448 472 496 520 544 568 592 616 640 664 688 712 736 760 784
Age Range (5 Ma bins)
Madagascar 'syn' Ages
(n=89)
Kuunga Orogen
or Extension
East Africa
Orogen
b
Frequency (Counts)
0
1
2
3
4
5
6
7
8
400 424 448 472 496 520 544 568 592 616 640 664 688 712 736 760 784
Age Range (5 Ma bins)
c
Daraina
BS
Antananarivo
VZ
AS
SQC
BR
AN
VS
Madagascar

et al., 1996; Ito et al., 1997; Paquette and Nédélec, 1998; Handke et al., 1999; Tucker et al., 1999a,b; Kröner et al., 2000a,b). An excellent summary of these data is provided by Kröner et al. (2000a,b). Figs. 7, 9 and 11) summarize the geochronologic data and tectonic setting for the basement rocks of this island.

Widespread magmatic activity in central Madagascar (gabbroic and granitoid intrusions) are dated between 824 and 740 Ma (Handke et al., 1999; Tucker et al., 1999a,b; Kröner et al., 2000a,b). Handke et al. (1999) concluded that the magmatism resulted from Andean-type subduction beneath central Madagascar (Itremo region). Zircon geochronologic data from gabbroic and granitic intrusions in central Madagascar overlap with the ages provided by Handke et al. (1999) and are compatible with subduction beneath Archean to early Proterozoic continental crust (Kröner et al., 2000a,b). Broadly coeval ages (715 to 754 Ma, U–Pb) are also found within the metamorphosed Daraina sequence in northernmost Madagascar (Tucker et al., 1999a,b) that are correlated with magmatism in the Seychelles and the Rajasthan region of India. Kröner et al. (2000a,b) and Muhongo et al. (2001) extend this active tectonic margin to Tanzania and Sri Lanka to the west and south and to Somalia–Ethiopia and Eritrea to the north.

Paquette and Nédélec (1998) dated ‘stratoid’ granites (layered granitic sills) in central Madagascar. They obtained U–Pb zircon ages of 627–633 Ma and interpreted the emplacement of these rocks in a postcollisional tectonic setting (Nédélec et al., 1995, 2000). According to Paquette and Nédélec (1998), Madagascar collided with East Africa at about 650 Ma and underwent extensional collapse at roughly 630 Ma. This intepretation is consistent with the timing of events in the ANS region described above. Meert et al. (in press) provide ^{40}Ar/^{39}Ar biotite ages from the stratoid granites north of Antananarivo. Their data suggests that these stratoid granites cooled rapidly following their 630 Ma emplacement (25 °C/Ma). Structural studies on the stratoid granites and high-grade rocks in the Antananarivo region by Nédélec et al. (2000) suggest that the structural trends are rotated from N–S (to the north of Antananarvio) to E–W in the so-called ‘virgation zone.’ U–Pb ages from within the virgation zone suggest that this structure formed at around 560 Ma (Paquette and Nédélec, 1998; Kröner et al., 2000a,b) slightly predating 550 Ma activity in the Angavo shear belt to the east. The deformation in the Angavo belt and virgation zone were relatively short-lived events (20–30 Ma). The late-syn to post tectonic Carion granite was emplaced at 532 Ma along the margin of the Angavo shear belt (SHRIMP U–Pb age, Meert et al., 2001a,b).

It is also clear that much of the island of Madagascar has experienced thermal overprinting during the 500–590 Ma interval (see Kröner et al., 2000a,b and references therein). Meert (1999) attributed these ages to a younger collision between Antarctica and Madagascar. Kröner et al. (2000a,b) argue that the granulite-facies metamorphism in central Madagascar was due to extensional collapse as did Collins et al. (2000). Geochronologic data from the Antananarivo region, along with new zircon, sphene and titanite U–Pb ages from southern Madagascar (discussed below) suggest that this younger metamorphism followed the main collision by some 80–100 million years (de Wit et al., 2001). In southern Madagascar (south of the Ranotsara shear zone), Paquette et al. (1994) report metamorphic zircon ages between 523 ± 5 and 577 ± 8 Ma based on upper and lower concordia intercept ages on granulites. They also reported Sm–Nd whole-rock–mineral ages from the same rocks that yielded ages ranging from 491 ± 5 to 588 ± 13 Ma. Collectively, these ages argue for a high-grade regional metamorphic event between 565 and 590 Ma with the younger ages representing a thermal fluid disturbance in the isotopic systematics (mostly along shear zones) following regional metamorphism. This ~ 570 Ma orogeny resulted in extensive granulite formation in close association with transcurrent tectonics along N–S shear zones. Similar ages were obtained for the southern Madagascar domains by Andriamarofahatra et al. (1990), Nicollet et al. (1997) and Montel et al. (1994, 1996). Martelat et al. (2000) define two deformation events in southern Madagascar that are continuous and overlapping from 590 to 500 Ma based on the electron micrprobe dating of monazite by Montel et al. (1994, 1996). The older deformation (590–530 Ma) was attributed to either crustal thickening or postcollisional extension, whereas the younger event (530–500 Ma) was attributed to a transpressional regime in a region of complex anastomosing shear zones.

Ashwal et al. (1999) interpreted the effects of prolonged metamorphism on U–Pb systematics of

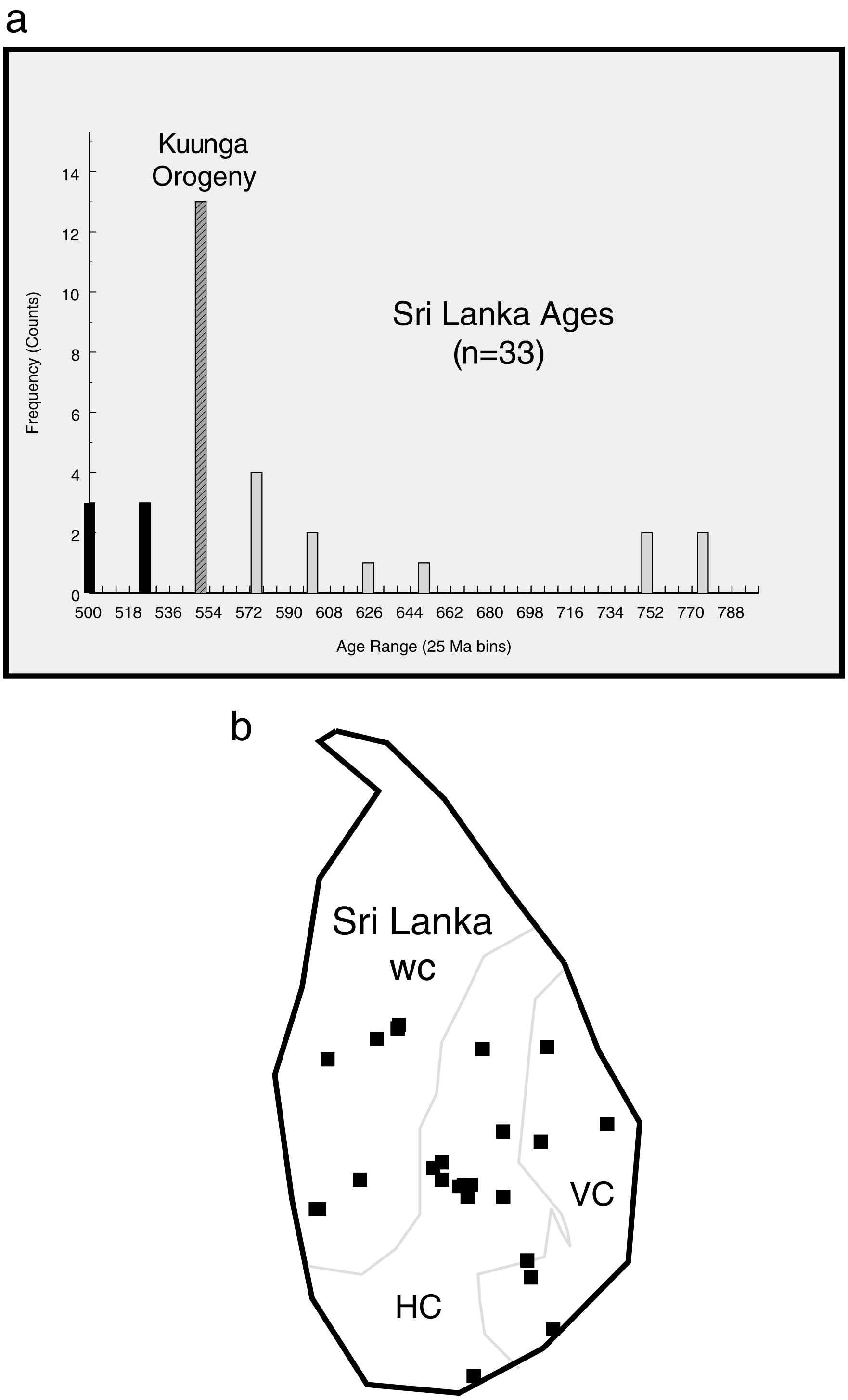


Fig. 13. (a) Histogram of ages (25 Ma bins) from Sri Lanka along with the major orogenic events in that region. (b) Geographic distribution of geochronologic studies used in the analysis. HC = Highland Complex, VC = Vijayan Complex and WC = Wanni Complex.

zircons from an anorthosite in southern Madagascar. They argue that a series of amphibolite-grade metamorphic events occurred in southern Madagascar between 630 and 530 Ma. This suggestion is further supported by the detailed structural and geochronologic studies of de Wit et al. (2001). According to those authors, collision in southern Madagascar occurred between 630 and 610 Ma based on U–Pb dating of zircon and monazite. The collisional metamorphism was followed by an extended period of static annealing at mid-crustal levels resulting in monazite growth and final orogenic collapse during the 530–490 Ma interval. Collectively, these results echo the conclusions of other workers who have attributed the younger high-grade event in southern Madagascar to either postcollisional extension and shearing or a separate collision (see Meert and Van der Voo, 1997; Hensen and Zhou, 1997; Paquette and Nédélec, 1998; Kröner et al., 2000a,b; Fitzsimons, 2000b).

2.6.1. Sri Lanka

Sri Lanka consists of three distinct tectonic provinces called the Highland, Wanni and Vijayan Complexes (Cooray, 1994; Fig. 12). The boundary between the Highland and Wanni Complexes is poorly defined, but both appear to have similar Neoproterozoic and younger histories as discussed below. The Highland Complex is in thrust contact with the Vijayan Complex. A detailed summary of geochronologic studies in Sri Lanka is given in Kröner and Williams (1993). Nd-model ages for the three domains in Sri Lanka were compiled by Möller et al. (1998; after Millisenda et al., 1988, 1994) and show a clear difference between the Highland and Vijayan–Wanni Complexes, the former consisting of Nd-model ages between 2 and 3.1 Ga and the latter two between 1.9 and 1.0 Ga. The Highland Complex consists of a Paleoproterozoic supracrustal assemblages and granitoid rocks that underwent granulite-grade metamorphism between 540 and 550 Ma and intense charnockitisation at ca. 550 Ma (Kröner et al., 1994). The Wanni Complex preserves a record of I-type intrusives dated between 750 and 1040 Ma (Baur et al., 1991; Kröner and Jaeckel, 1994) that were correlated with similar-age intrusions in Madgascar by Kröner et al. (2000a,b).

Neither the Wanni nor the Vijayan Complexes contain crustal rocks older than early Neoproterozoic (1000–1100 Ma); however, the late Neoproterozoic though Cambrian tectonic history of the Wanni Complex is identical to that of the Highland Complex (Kröner and Williams, 1993). The Vijayan Complex underwent upper amphibolite-grade metamorphism during the 465–558 Ma interval. The marked difference in rock types, metamorphic grade, timing of deformation and nature of the tectonic contact between the HC and VC led Kröner and Williams (1993) to suggest that they were juxtaposed during final assembly of Gondwana. According to their hypothesis, the HC–WC were part of a tectonic block consisting of S. Madagascar, S. India and Tanzania. In contrast, Hözl et al. (1994) concluded that peak metamorphism in all three complexes occurred at ca. 610 Ma and that separate tectonothermal events in Sri Lanka were not resolvable with the available geochronologic data. Rb–Sr whole-rock–mineral ages of posttectonic granites in Sri Lanka were reported (without details) by Fernando et al. (1999). The ages ranged from 464 ± 28 Ma (Tonigala-B granite) to 533 ± 6 Ma for the Kotadeniya granite. The geochronologic data for Sri Lanka are summarized in Figs. 8 and 13.

3. East Antarctica

Two recent comprehensive reviews of geochronologic data from East Antarctica and their tectonic significance are given by Fitzsimons (2000a,b). These data are summarized in Figs. 8, 10 and 14). There are several key observations that are repeated here. The first is that the circum-Antarctic mobile belt that was thought to represent the major orogenic cycle responsible for the assembly of east Gondwana consists of three distinct orogens separated by much younger (≤ 600 Ma) mobile belts (Moores, 1991). The oldest of the three Meso–Neoproterozoic orogenic belts suggest a connection between the Wilkes Province (East Antarctica) and the Albany–Fraser belt of Australia by ∼ 1200 Ma (Sheraton et al., 1995). As noted above, Jacobs et al. (1998) provide strong evidence that the Maud Province of East Antarctica was juxtaposed with the Namaqua–Natal provinces of South Africa during an 1100–1000 Ma orogenic event. Gose et al. (1997) provided paleomagnetic support for this link. The Rayner Province and north-

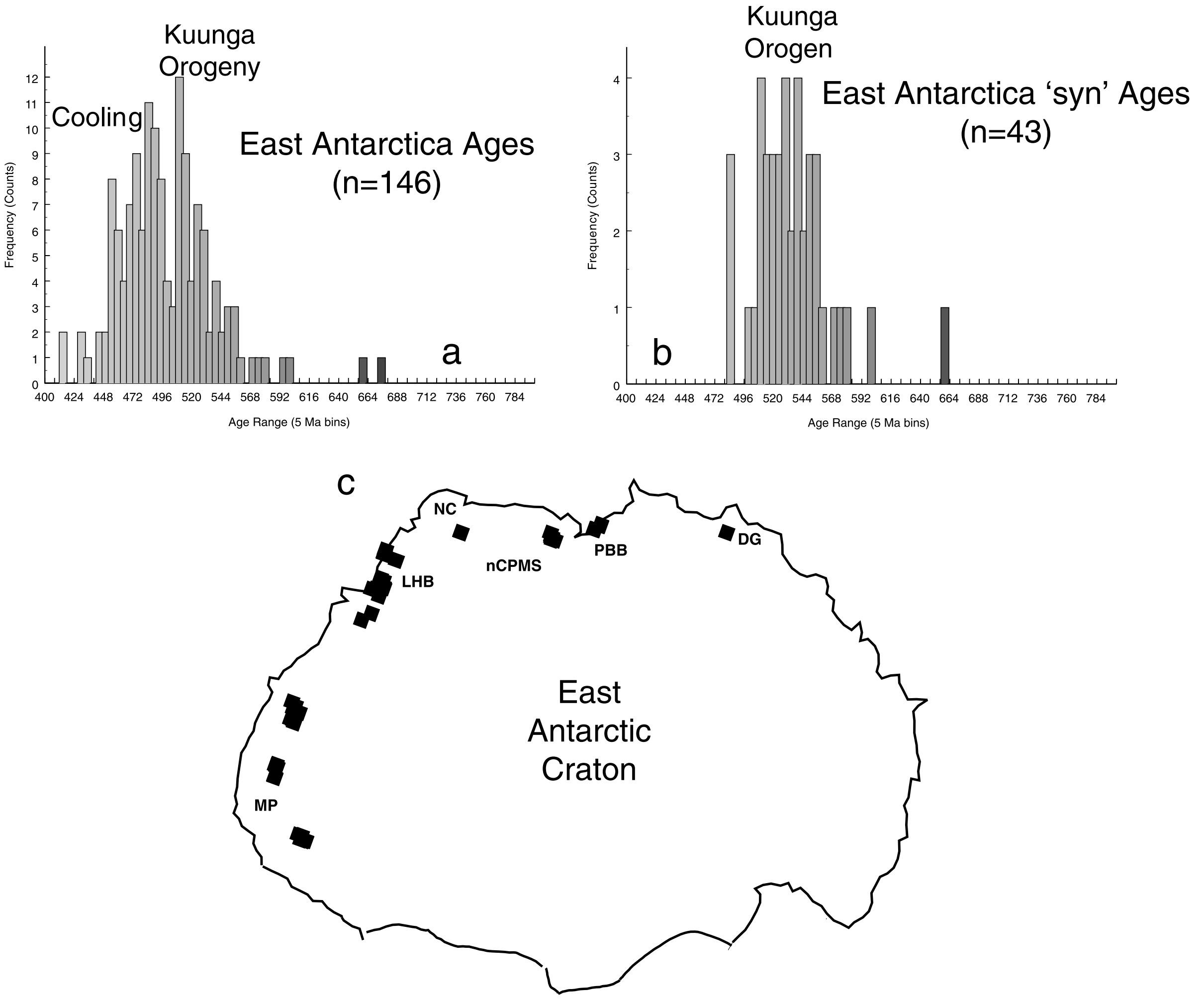


Fig. 14. (a) Histogram of all 'C' and better ages (5 Ma bins) from East Antarctica along with the major orogenic events in that region. (b) Histogram of 'syn' assembly ages from Antarctica. (c) Geographic distribution of geochronologic studies used in the analysis. LHB = Lützow–Holm Belt; MP = Maud Province; NC = Napier Complex; nCPMS = northern Prince Charles Mountains; PBB = Prydz Bay Belt.

ern Prince Charles Mountains regions of East Antarctica are correlated with the eastern Ghats region of India. Recent geochronologic data from both the eastern Ghats region (Mezger and Cosca, 1999) and the northern Prince Charles Mountains region (Boger et al., 2000) provide evidence for major tectonism and amalgamation of the two regions sometime between 1000 and 900 Ma followed by a relatively minor thermal reworking in the latest Neoproterozoic.

The presence of a significant Pan-African overprint in Antarctica was first realized by Shiriashi et al. (1994). These authors reported U–Pb ages between 550 and 530 in the Lützow–Holm region of Antarctica. Significant Pan-African orogenesis is further supported by the work of Jacobs et al. (1998) in central Dronning Maud Land. Their data documented two metamorphic episodes between 570 and 550 Ma and between 530 and 515 Ma with the latter episode recording granulite-facies metamorphism.

A number of high-grade metamorphic rocks crop out in the Prydz Bay region of Antarctica. These rocks record a number of late Neoproterozoic to early Paleozoic ages. Zhao et al. (1992) reported Pb–Pb zircon evaporation ages of 540–560 Ma for the synkinematic Progress granite. Subsequent work by Carson et al. (1996) resulted in a significantly younger age for the same granite of 515 Ma (SHRIMP dating). Hensen and Zhou (1995) also reported Sm–Nd whole-rock–garnet ages ranging from 467 to 517 Ma for metamorphic rocks in the southern Prydz Bay region. Fitzsimons et al. (1997) obtained SHRIMP U–Pb ages between 550 and 500 Ma for paragneisses in the same region. Harley (1998) reports clear evidence for ultrahigh $P-T$ metamorphism in the Rauer Group (Prydz Bay region), but the temporal constraints (e.g., ~ 1000 Ma or ~ 530 Ma) on this metamorphism are poor. Collectively, the data from the Prydz Bay region suggest that a major tectonothermal event occurred in that region sometime between 550 and 500 Ma. Farther east, in the Denman glacier region, Black et al. (1992) give evidence for thermal resetting of the region sometime between 600 and 500 Ma.

4. Paleomagnetism

Comprehensive reviews of Neoproterozoic paleomagnetic data from the Gondwana crustal components are provided by Meert and Van der Voo (1997), Meert (1999) and Meert (2001). Table 2 lists paleomagnetic data compiled from these previous reviews along with recent updates. Meert and Van der Voo (1997) concluded on the basis of the (then) extant database that Gondwana assembly had been completed by ~ 550 Ma. Subsequent updates to the paleomagnetic database for the interval from 550 to 500 Ma have not drastically altered this conclusion, and Meert (2001) concludes that the available paleomagnetic data are compatible with final Gondwana assembly occurring between 550 and 530 Ma. This conclusion was misinterpreted by some authors (Appel et al., 1998; Möller et al., 2000) as arguing against earlier collisions. However, the paleomagnetic interpretation does not preclude the earlier collision of other terranes within Gondwana; small-scale motions between individual cratonic nuclei; or continued collision between the elements of Gondwana—only that they fall outside of paleomagnetic resolution at the present time. There is one other important note with regard to the database. There are no paleomagnetic data from Antarctica older than about 515 Ma. This makes it difficult to test for cratonic coherence of both east Gondwana and East Antarctica, in particular. Australian data provide a better test for the age of coherence within the blocks thought to comprise east Gondwana because they form a nearly continuous record from 600 to 500 Ma. The Dokhan volcanics pole (Davies et al., 1980) was typically ignored in recent compilations due to poor age control. Recent SHRIMP dating of these rocks yielded an age of 593 ± 15 Ma (Wilde and Youssef, 2000). This pole is quite distinct from coeval Australian rocks and suggests that either the ANS or Australia was not in its final Gondwana configuration. Assuming that the ANS region had amalgamated with eastern Africa prior to 590 Ma as discussed above, this pole provides further evidence for a younger collision related to the union of Australia with Gondwana.

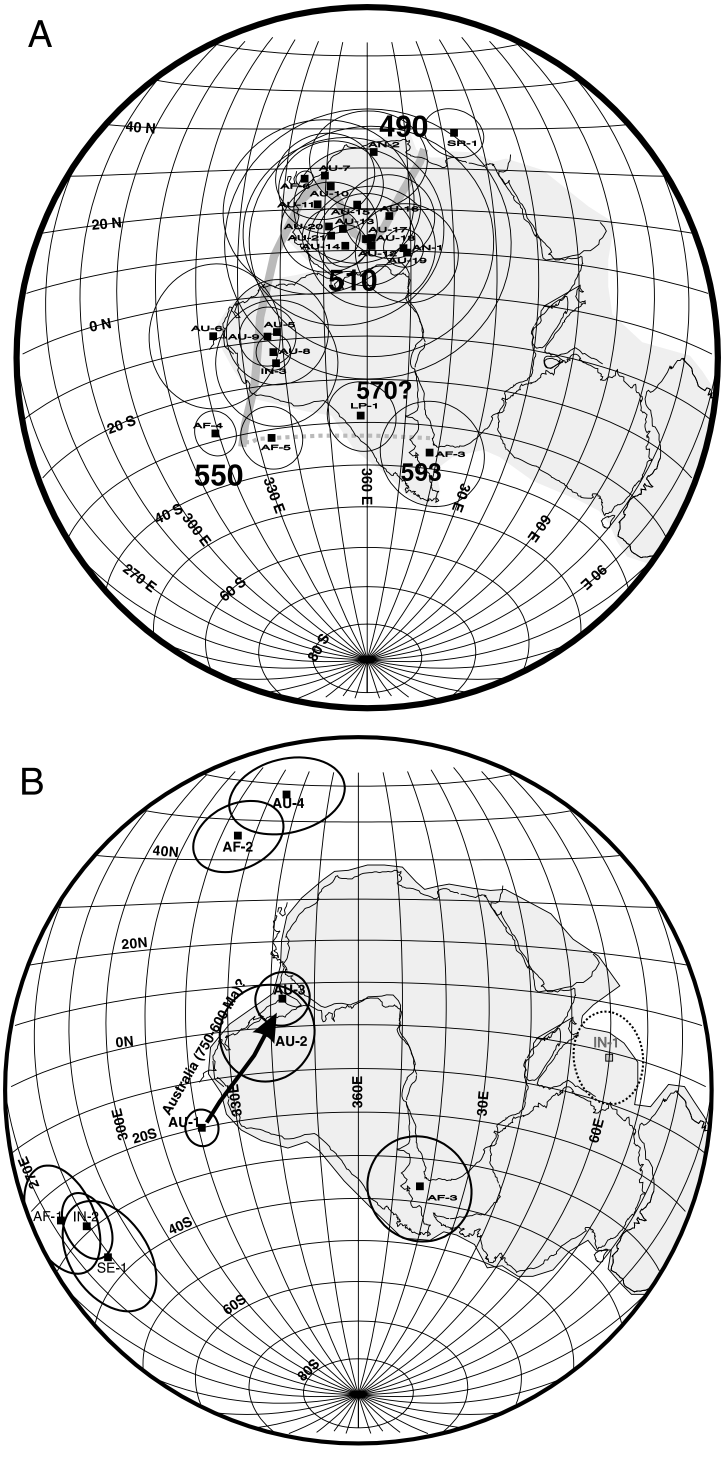
Despite the limitations of the younger segment of the Gondwana path, there are several tantalizing clues regarding the coherence of the east Gondwana continent prior to 600 Ma. Table 2 lists the available data from Gondwana including several new studies that are in press. The paleomagnetic data in Fig. 15 are rotated to Gondwana coordinates in order to test possible coherence of the blocks. Unfortunately, the database is so sparse as to preclude any details on the drift histories of individual cratonic blocks. Nevertheless, spot paleomagnetic readings at ~ 750 Ma present problems for the notion of a united east Gondwana.

Wingate and Giddings (2000) reported paleomagnetic data from the Mundine dyke swarm in Australia. These dykes have a well-established age of 755 Ma and their pole falls at 20°S, 321°E (Gondwana coordinates, Africa fixed). Coeval paleomagnetic data from the Seychelles (Torsvik et al., 2001a) and from the Malani rhyolites of India (Klootwijk, 1975; Torsvik et al., 2001b) constrain the position of these two blocks. The poles fall at 29°S, 279°E (Malani) and 36°S, 279°E (Seychelles) when rotated to Gondwana coordinates (Africa fixed). The poles from India–Seychelles are displaced by nearly 40° in a conventional east Gondwana fit with respect to the Mundine dikes pole from Australia (Wingate and Giddings, 2000). Paleolongitudes are unconstrained by paleomagnetic data so it is important to note that not only are the poles from these two east Gondwana blocks displaced, the closest fit that can be made between Australia and India results in a mismatch of geologic

Table 2
Selected paleomagnetic poles 800–500 Ma for eastern Gondwana

Pole #	Name	α_{95}	Plat	Plong	Age (S,I)[a]	Reconstr.	References
African poles/Arabian							
AF-1	Gagwe Lavas	10°	25.0S	273.0E	813.0 (I)	fixed	Meert et al., 1995
AF-2	Mbozi Complex	9°	46.0N	325.0E	743.0 (I)	fixed	Meert et al., 1995
AF-3	Dokhan Volcanics	12.1°	36.1S	017.1E	593.0 (I)	fixed	Davies et al., 1980
AF-4	Sinyai Dolerite	5.0°	28.4S	319.1E	547.0 (I)	fixed	Meert and Van der Voo, 1996
AF-5	Mirbat SS (Oman)	7.2°	31.9S	333.9E	550.0 (S,I)	ANS–AFR	Kempf et al., 2000
AF-6	Ntonya Ring Structure	1.9°	27.5N	355.2E	522.0 (I)	fixed	Briden et al., 1993
LP-1	Callander/Catoctin Mean	8.0°				LAU–AFR	Meert et al., 1994
Antarctic poles							
AN-1	Sor Rondane	4.5°	10.6N	008.3E	515.0 (I)	ANT–AFR	Zijderveld, 1968
AN-2	Mt. Loke/Killer Ridge	8.0°	34.0N	001.6E	499.0 (S,I)	ANT–AFR	Grunow and Encarnacion, 2000a,b
Australian poles							
AU-1	Mundine Dykes	4°	46.0N	135.0E	755.0 (I)	20S, 321E	Wingate and Giddings, 2000
AU-2	Yaltipena Fm	11°	44.2N	172.7E	600.0 (S)	01S, 339E	Sohl et al., 1999
AU-3	Elatina Fm	6.2°	39.7N	181.9E	600.0 (S)	07N, 343E	Sohl et al., 1999
AU-4	Bunyeroo Fm (Impact)	10.7°	18.1S	16.3E	590.0 (S,I)	57N, 335E	Williams and Schmidt, 1996
AU-5	Lower Arumbera SS	12°	8.2S	338.8E	550.0 (S)	AUS–AFR	Kirschvink, 1978
AU-6	Brachina Fm	16°	7.4S	323.5E	550.0 (S)	AUS–AFR	McWilliams and McElhinny, 1980
AU-7	Bunyeroo Fm	10.7°	28.2N	349.7E	550.0 (S)	AUS–AFR	McWilliams and McElhinny, 1980
AU-8	U. Arumbera SS	4.1°	12.6S	337.6E	535.0 (S)	AUS–AFR	Kirschvink, 1978
AU-9	Todd River Dolomite	6.7°	9.0S	336.5E	532.0 (S)	AUS–AFR	Kirschvink, 1978
AU-10	Hawker Gp. A	11.4°	25.5N	351.2E	525.0 (S)	AUS–AFR	Klootwijk, 1980
AU-11	Hawker Gp. B	21.2°	21.2N	348.3E	525.0 (S)	AUS–AFR	Klootwijk, 1980
AU-12	Aroona–Wirealpa-A	14.4°	11.1N	001.0E	510.0 (S)	AUS–AFR	Klootwijk, 1980
AU-13	Aroona–Wirealpa-B	22.6°	15.2N	354.4E	510.0 (S)	AUS–AFR	Klootwijk, 1980
AU-14	Tempe Fm	5.0°	11.1N	355.0E	510.0 (S)	AUS–AFR	Klootwijk, 1980
AU-15	Hudson Fm	14.0°	21.0N	357.6E	508.0 (S)	AUS–AFR	Luck, 1972
AU-16	Lake Frome-A	10.1°	18.2N	005.1E	505.0 (S)	AUS–AFR	Klootwijk, 1980
AU-17	Lake Frome-B	27.7°	13.0N	001.0E	505.0 (S)	AUS–AFR	Klootwijk, 1980
AU-18	Giles Creek Dol. L.	32.6°	12.7N	000.2E	505.0 (S)	AUS–AFR	Klootwijk, 1980
AU-19	Giles Creek Dol U.	11.7°	9.6N	009.1E	505.0 (S)	AUS–AFR	Klootwijk, 1980
AU-20	Illara SS	10.8°	15.8N	351.1E	505.0 (S)	AUS–AFR	Klootwijk, 1980
AU-21	Deception Fm	6.5°	13.6N	351.7E	500.0 (S)	AUS–AFR	Klootwijk, 1980
Indian poles							
IN-1	Harohalli Dikes Combined	9.2°	27.3N	78.9E	823.0 (I)	01N, 060E	Radhakrishna and Joseph, 1996
IN-2	Malani Igneous Suite[2]	6.4°	74.5N	71.2E	758.0 (I)	29S, 279E	Torsvik et al., 2001a
IN-3	Bhander–Rewa Mean	11°	23.0S	333.0E	550.0 (S)	IND–AFR	McElhinny et al., 1978
Madagascan poles							
MA-1	Carion Granite	11°	12.7N	359.7E	508.0 (I)	MAD–AFR	Meert et al., 2001
Seychelles poles							
SE-1	Mahe Island Rocks	11.2°	54.8N	57.6E	750.0 (I)	36S, 279E	Torsvik et al., 2001a
Sri Lanka poles							
SR-1	Tonigala Granite	6.2°	39.0N	23.2E	500 (I)	SRI–AFR	Yoshida et al., 1992

[a] S = stratigraphic age, I = isotopic age, SI = both stratigraphic and isotopic age. Rotation parameters: ANS–AFR: 26.5°N, 21.5°E, −7.6°; ANT–AFR: −7.78°N, −31.42E, +58°; AUS–AFR: 25.1°N, 110.1°E, −56.7°; IND–AFR: 27.9°, 43.64°, −64.4°; MAD–AFR: −3.41N, −81.7E, +19.7°; SAM–AFR: 45.5°N, −32.2°E, +58.2°; SRI–AFR: 17.3°N, 52.5°E, −89.8°; LAU–AFR: 17.7°N, 350.5° E, +145.3.


A
490
SR-1
AN-2
AU-7
AF-6
AU-10
AU-11
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AU-18
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AU-13
AU-14
AU-12
AN-1
AU-19
510
AU-6
AU-5
AU-9
AU-8
IN-3
570?
LP-1
AF-4
AF-5
AF-3
550
593
40 N
20 N
0 N
20 S
40 S
60 S
80 S
270 E
300 E
330 E
360 E
30 E
60 E
90 E
B
AU-4
AF-2
40N
20N
0N
20S
40S
60S
80S
AU-3
AU-2
Australia (750-600 Ma)?
AU-1
270E
300E
330E
360E
30E
60E
IN-1
AF-1
IN-2
SE-1
AF-3

features. Although the evidence from spot readings of the paleomagnetic data do not provide compelling evidence for the separation of the east Gondwana blocks at 750 Ma, the data are clearly compatible with such a proposal.

5. Discussion

Did the assembly of eastern Gondwana take place through a series of collisional events between formerly unrelated elements or did east Gondwana represent a coherent fragment of Rodinia? If so, where are the sutures marking the final closure of the intervening oceans and how strong is the evidence to support these collisions? Both the geochronologic and paleomagnetic data can be used to argue that east Gondwana assembly paralleled the assembly of greater Gondwana during the interval from 750 to 530 Ma. The main problem is determining what elements behaved as coherent entities and when final assembly occurred. The age distributions discussed above show that the history of amalgamation in eastern Gondwana is not a simple merging of east and west Gondwana. Arguments favoring a single collision followed by progressive crustal thickening from west to east (Windley et al., 1994; Collins et al., 2000) are difficult to reconcile with the areal distribution of high-grade events in eastern Gondwana described above. For example, why do the regions south of Tanzania (including the Maudheim Province of east Antarctica) indicate peak metamorphism during the interval from 570 to 530 Ma, while data from Tanzania northward indicate peak metamorphism at ~ 640 Ma (Fig. 10)?

5.1. A polyphase model for eastern Gondwana assembly

The available geochronologic data, combined with limited paleomagnetic data, can be used to argue for a polyphase assembly of eastern Gondwana. Assuming that the geochronologic data catalogued and described in this study faithfully record at least two distinct pulses of orogenic activity in eastern Gondwana, it is possible to focus on the makeup of the major elements involved in those orogenies.

The East Africa Orogen (EAO, Stern, 1994) records the amalgamation of arc terranes in the Arabian–Nubian shield region and the collision of a major landmass to the south. Stern and Abdelsam (1998) argue that terrane amalgamation in the ANS region began around 750 Ma. A number of geochronologic studies cited above suggest that the entire region was assembled by 630 Ma with continental escape in the ANS following shortly thereafter. Farther south in Kenya and Tanzania, the available geochronology indicates a major episode of high-grade metamorphism at ~ 640 Ma although others have suggested that this metamorphism was unrelated to continental collision (Möller et al., 2000; Kröner et al., 2000b). This paper favors the conclusion reached by Stern (1994), wherein the collision in the southern part of the EAO involved oblique continent–continent collision and the EAO further north evidences the accretion of juvenile terranes and continental escape. The isobaric cooling noted in the Tanzania granulites may have resulted from the thermal effects of lithospheric delamination beneath Tanzania. Suggested postorogenic extensional magmatism in central Madagascar at 630 Ma, derived from the stratoid granites (Paquette and Nédélec, 1998), is compatible with the suggestion of de Wit et al. (2001) for continental collision at ca. 650 Ma (see also Ashwal et al., 1999). There is no evidence of a ca. 640 Ma orogenic episode in India, Sri Lanka or East Antarctica, but this may also be due to the fact that collision in those regions was younger than southern and central Madgascar and that these areas were relatively unaffected by the older deformation. Kröner et al. (2000b) suggest that east Gondwana was a collage of terranes that amalgamated during a 100+ million year time period, but they did not provide details on the makeup of the individual blocks that collided during the 640–530 Ma interval. Indeed, the boundaries between elements suggested in the model below are also ill-defined, but investigations into the isotopic characteristics, detrital zircon populations and mineral distributions within the for-

Fig. 15. (A) Paleomagnetic data from 590–490 Ma from Gondwana taken from Table 2 and Meert (2001). (B) Paleomagnetic data from Table 2 for the time period from 600 to 750 Ma. Note the discordance between 750 Ma paleomagnetic data from India–Seychelles (IN-2, Malani; SE-1, Seychelles) and Australia's Mundine dykes (AU-1).

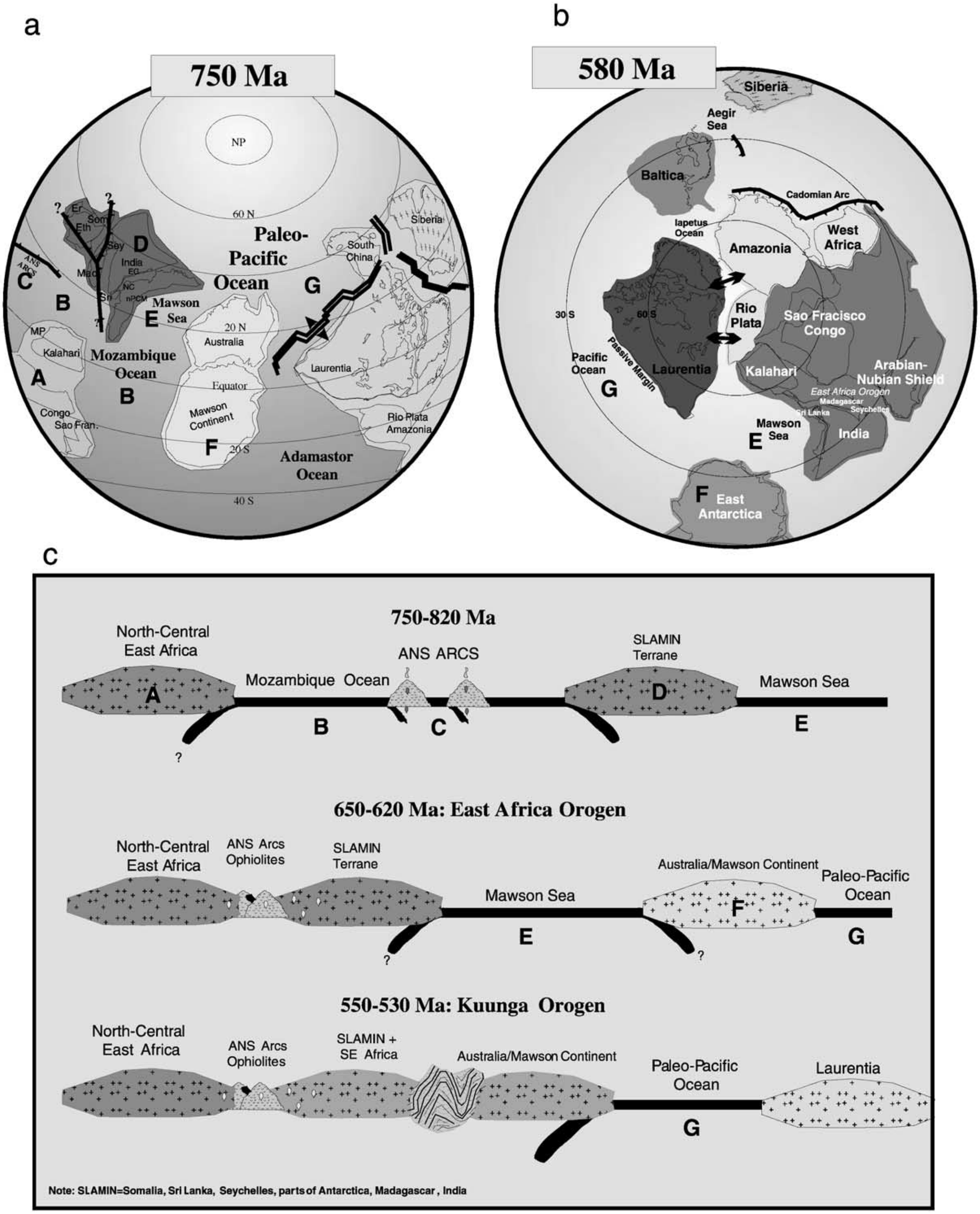
merly contiguous regions of Gondwana (e.g., Cox et al., 1998; Möller et al., 1998; Stern and Abdelsam, 1998; Dissanayake and Chandrajith, 1999) may prove key in delineating the assembly and fragmentation history of eastern Gondwana.

The development of the EAO is shown schematically in Fig. 16a–c (a text summary is given in Fig. 8). The plate reconstruction in Fig. 16a shows the cratonic elements involved in the formation of eastern Gondwana at ca. 750 Ma and Fig. 16b shows the reconstruction at 580 Ma (based on paleomagnetic data). At the present time it is difficult to delineate exact boundaries between individual arc terranes in the ANS region so they are simply labeled as 'arc terranes' in the figure. Kröner et al. (2000a,b) and Kröner and Sassi (1996) suggested that Somalia and part of Ethiopia were continental blocks sandwiched between arc terranes, but the exact boundaries are poorly known. Kröner et al. (2000a,b) hint that arc magmatism in Somalia may have a genetic link to magmatism in central Madagascar. There is also controversial evidence for arc-related magmatism in the Seychelles, the Malani region of India and parts of eastern Tanzania (see above). It is possible that the EAO formed through a series of arc collisions in the ANS region followed by the collision of a continental block comprised of at least central Madagascar, India, a sliver of continental crust from Antarctica (Rayner Province, nPCMs and LHB—see Fig. 9) and Sri Lanka with Kenya and Tanzania at ~ 640 Ma. The remainder of eastern Africa (Kenya) behaved as a passive margin until the ~ 640 Ma orogenesis.

The 550–530 Ma Kuunga Orogeny (Swahili meaning 'to unite') was originally defined by Meert et al. (1995) in an effort to explain the difference between geochronologic evidence, suggesting a ca. 650 Ma amalgamation age and the paleomagnetic data that favored a 550 Ma amalgamation age for Gondwana. Temporal constraints on the Kuunga orogeny now suggest that the timing should be expanded to encompass the 570–530 Ma interval (see Fig. 10). The boundaries of the Kuunga orogeny were not sharply drawn, but the suggestion was that the orogeny followed the coastal margins of East Antarctica, southern India, Sri Lanka, southern Madagascar and south-eastern Africa. In a subsequent paper (Meert, 1999), the Kuunga Orogeny was correlated with the margins of the ancient craton called 'Ur' (see Rogers, 1996). The Kuunga Orogeny caused isotopic resetting and/or new zircon–monazite growth in central Madagascar (primarily along shear zones), along the eastern Ghats region of India and high-grade metamorphism and transcurrent tectonism in southern Madagascar, Sri Lanka, the Maud Province, Lützow–Holm and Prydz Bay regions of East Antarctica, southeastern Africa and along the Darling belt of Australia. The Kuunga orogeny may also responsible for changes in the loci of transcurrent tectonism and continental escape in the regions from Kenya to the Arabian–Nubian shield (Kusky and Matesh, 1999). The suture (Fig. 10) for the Kuunga Orogeny may run from the Darling belt of Australia through the Prydz Bay and southern Prince Charles Mountains region of East Antarctica and on into the Maudheim Province as suggested by Fitzsimons (2000a,b) or alternatively, it would more closely parallel coastal East Antarctica as outlined in Boger et al. (2001). It is interesting to note that tectonism in the Damara–Gariep belts occurs during the Kuunga time span. If the Kalahari craton was 'hinged' in the region of the Zambezi belt, an oblique collision between the East Antarctic and Kalahari cratons may have resulted in the closure of the Damara belt by a clockwise rotation of the Kalahari craton about the hinge.

Grunow et al. (1996) suggested that the final stages of Gondwana assembly were genetically related to the opening of the Iapetus Ocean between Laurentia and elements of west Gondwana. The drift history of

Fig. 16. (a) Reconstruction at 750 Ma (based partly on the paleomagnetic data given in Table 2). Individual arc elements in the Arabian–Nubian sector are shown schematically as ANS arcs in the figure. Crustal block(s?) comprised of Somalia, Ethiopia–Eritrea, Madagascar, Sri Lanka, Seychelles, Antarctica and India are also shown schematically as active continental margins. The exact boundaries between these crustal blocks are ill-defined. (b) Reconstruction at 580 Ma (based partly on paleomagnetic data in Table 2). Rifting of Laurentia from western portions of Gondwana occurred sometime in the 565 to 510 Ma interval. The positions of Baltica and Siberia are based on data in Torsvik and Rehnstrom (2001). (c) Cartoon showing the development of eastern Gondwana during the interval 820–530 Ma. The top and middle diagrams show the closure of the Mozambique Ocean between the ANS arcs, SLAMIN terrane and eastern Africa during the East Africa Orogen; the bottom diagram represents the collision between the bulk(?) of the East Antarctic craton and Australia with the previously assembled elements of the EAO during the Kuunga Orogen. Letters A–G relate regions in the paleoreconstructions to those in the cross-sectional diagrams.


a
750 Ma
NP
60 N
Paleo-Pacific Ocean
Siberia
South China
G
Er
Som
Eth
Sey
D
India
EG
Mad
NC
Sri
nPCM
ANS ARCS
C
B
Mawson Sea
E
MP
Kalahari
Mozambique Ocean
A
B
20 N
Australia
Equator
Mawson Continent
F
20 S
Laurentia
Rio Plata
Amazonia
Congo
Sao Fran.
Adamastor Ocean
40 S
b
580 Ma
Siberia
Aegir Sea
Baltica
Cadomian Arc
Iapetus Ocean
Amazonia
West Africa
30 S
60 S
Rio Plata
Sao Fracisco Congo
Pacific Ocean
Passive Margin
Laurentia
Kalahari
Arabian-Nubian Shield
East Africa Orogen
Madagascar
Sri Lanka
Seychelles
G
Mawson Sea
E
India
F
East Antarctica
c
750-820 Ma
North-Central East Africa
A
Mozambique Ocean
B
ANS ARCS
C
SLAMIN Terrane
D
Mawson Sea
E
?
650-620 Ma: East Africa Orogen
North-Central East Africa
ANS Arcs Ophiolites
SLAMIN Terrane
Mawson Sea
E
Australia/Mawson Continent
F
Paleo-Pacific Ocean
G
?
?
550-530 Ma: Kuunga Orogen
North-Central East Africa
ANS Arcs Ophiolites
SLAMIN + SE Africa
Australia/Mawson Continent
Paleo-Pacific Ocean
G
Laurentia
Note: SLAMIN=Somalia, Sri Lanka, Seychelles, parts of Antarctica, Madagascar, India

Laurentia is poorly known between 565 and 510 Ma, but the opening of the Iapetus most likely occurred during that interval as 510 Ma paleomagnetic data from Gondwana and Laurentia show a clear separation, while 565 Ma data do not. Therefore, the existence of the supercontinent Panottia (Powell et al., 1995) is unresolved, but the connection between final Gondwana assembly and opening of the Iapetus ocean is supported by this study.

Shackleton (1996) asked "Where is the final collision zone between east and west Gondwana?" A number of collisional sutures have been proposed for the assembly of this part of Gondwana and there has been considerable debate surrounding the tracing of the suture from north to south (see Fig. 10). In my opinion, the difficulty stems from the fact that there was not a single collision between east and west Gondwana; therefore, we should not be able to identify a single suture. A better question, though still unresolved, is "Where are the suture(s) between the elements of eastern Gondwana?"

6. Conclusions

The assembly of the eastern part of Gondwana (eastern Africa, Arabian–Nubian shield, Seychelles, India, Madagascar, Sri Lanka, East Antarctica and Australia) resulted from a complex series of orogenic events spanning the interval from ~ 750 to ~ 530 Ma. A detailed examination of the geochronologic database from key cratonic elements in eastern Gondwana suggest a multiphase assembly. The model outlined in this paper precludes the notion of a united east Gondwana until after the East African and Kuunga orogenies and strongly suggests that its assembly paralleled the final assembly of greater Gondwana. It is possible to identify at least two main periods of orogenesis within eastern Gondwana. The older orogeny resulted from the amalgamation of arc terranes in the Arabian–Nubian shield region and continent–continent collision between eastern Africa (Kenya–Tanzania and points northward) with an, as of yet, ill-defined collage of continental blocks including parts of Madagascar, Sri Lanka, Seychelles, India and East Antarctica during the interval from ~ 750 to 620 Ma. This is referred to as the East Africa Orogeny in keeping with both the terminology and the focus of the paper by Stern (1994). The second major episode of orogenesis took place between 570 and 530 Ma and resulted from the collision between Australia and an unknown portion of East Antarctica with the elements previously assembled during the East African Orogen and closure of the Damara belt. This is referred to as the Kuunga Orogeny following the suggestion of Meert et al. (1995). The problem in identifying a single suture and its extension beyond Tanzania stems from the fact that there are at least two collisions. The older EAO resulted in the suture known as the Mozambique belt (N–S trending) and the second suture is ill-defined at present, but may run from the west coast of Australia through the Prydz Bay–Southern Prince Charles Mountains of East Antarctica through to the Maudheim Province. Subduction of the Paleo-Pacific Ocean along the Ross–Delamerian belt followed closely after the Kuunga Orogeny at ~ 500 Ma (Goodge et al, 1993; Foster and Gray, 2000).

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