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Review

The Saharan Metacraton

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Abstract

This article introduces the name "Saharan Metacraton" to refer to the pre-Neoproterozoic—but sometimes highly remobilized during Neoproterozoic time—continental crust which occupies the north-central part of Africa and extends in the Saharan Desert in Egypt, Libya, Sudan, Chad and Niger and the Savannah belt in Sudan, Kenya, Uganda, Congo, Central African Republic and Cameroon. This poorly known tract of continental crust occupies $\sim$5,000,000 km^2 and extends from the Arabian-Nubian Shield in the east to the Tuareg Shield to the west and from the Congo craton in the south to the Phanerozoic cover of the northern margin of the African continent in southern Egypt and Libya. The term "metacraton" refers to a craton that has been remobilized during an orogenic event but is still recognizable dominantly through its rheological, geochronological and isotopic characteristics. Neoproterozoic remobilization of the Saharan Metacraton was in the forms of deformation, metamorphism, emplacement of igneous bodies, and probably local episodes of crust formation related to rifting and oceanic basin development. Relics of unaffected or only weakly remobilized old lithosphere are present as exemplified by the Archean to Paleoproterozoic charnockites and anorthosites of the Uweinat massif at the Sudanese/Egyptian/Libyan boarder. The article explains why the name "Saharan Metacraton" should be used, defines the boundaries of the metacraton, reviews geochronological and isotopic data as evidence for the presence of pre-Neoproterozoic continental crust, and discusses what happened to the Saharan Metacraton during the Neoproterozoic. A model combining collisional processes, lithospheric delamination, regional extension, and post-collisional dismembering by horizontal shearing is proposed.

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1. Introduction

The African continent was the center of "Greater Gondwana" (Stern, 1994) or Pannotia (Dalziel, 1997) at the end of the Precambrian and comprises Archean cratons and Neoproterozoic orogenic belts (Fig. 1). However, the tract of continental crust which is dominated by medium- to high-grade gneissic and migmatitic terrains and occupies the area between the Arabian-Nubian Shield in the east and the Tuareg Shield to the west and the Congo craton in the south and the Phanerozoic cover of the northern continental margin of the African continent in southern Egypt and Libya (Fig. 2) was not a craton nor an orogenic belt in the classical meanings of the terms (Bates and Jackson, 1980) during Neoproterozoic time. Geochronological and isotopic

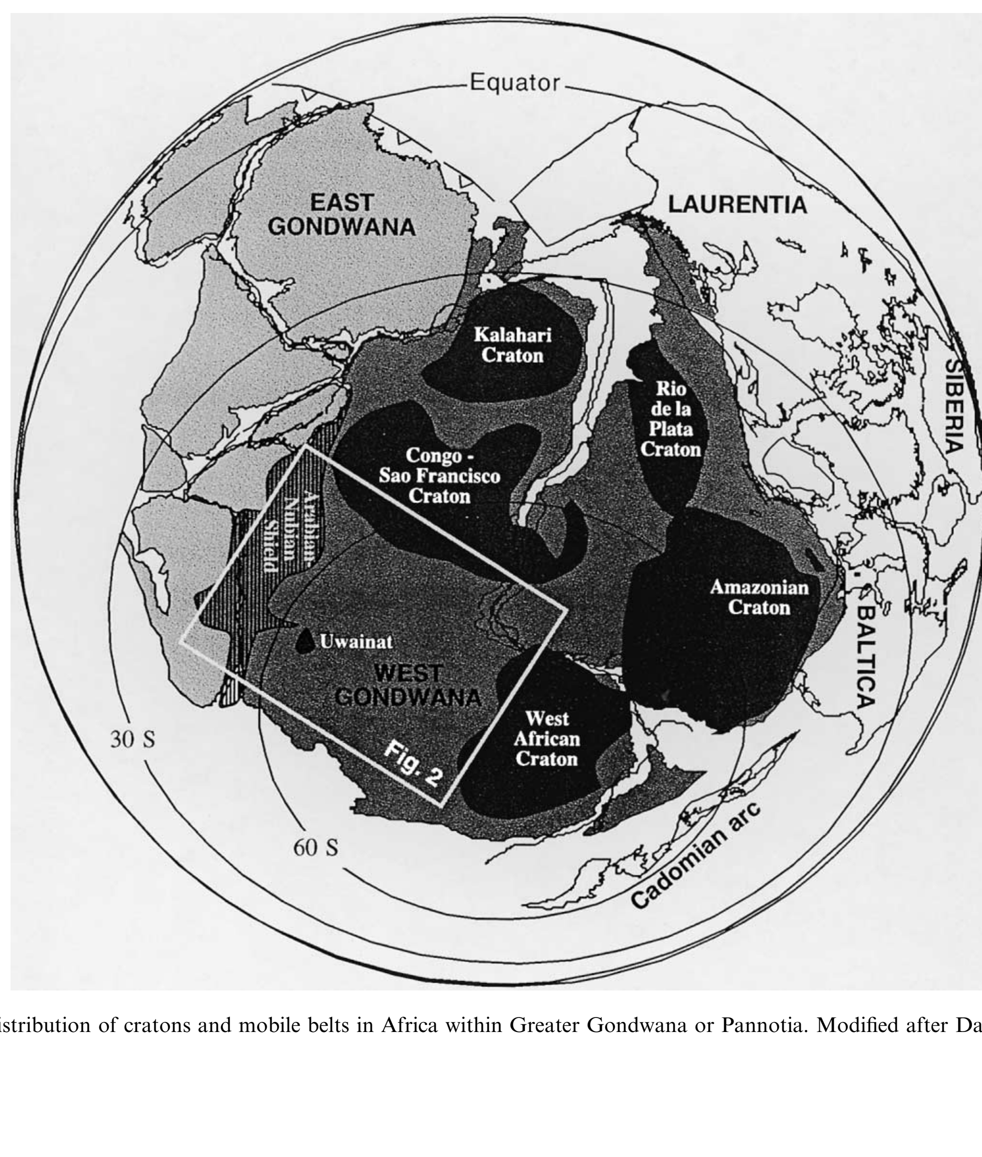

Fig. 1. Distribution of cratons and mobile belts in Africa within Greater Gondwana or Pannotia. Modified after Dalziel (1997).

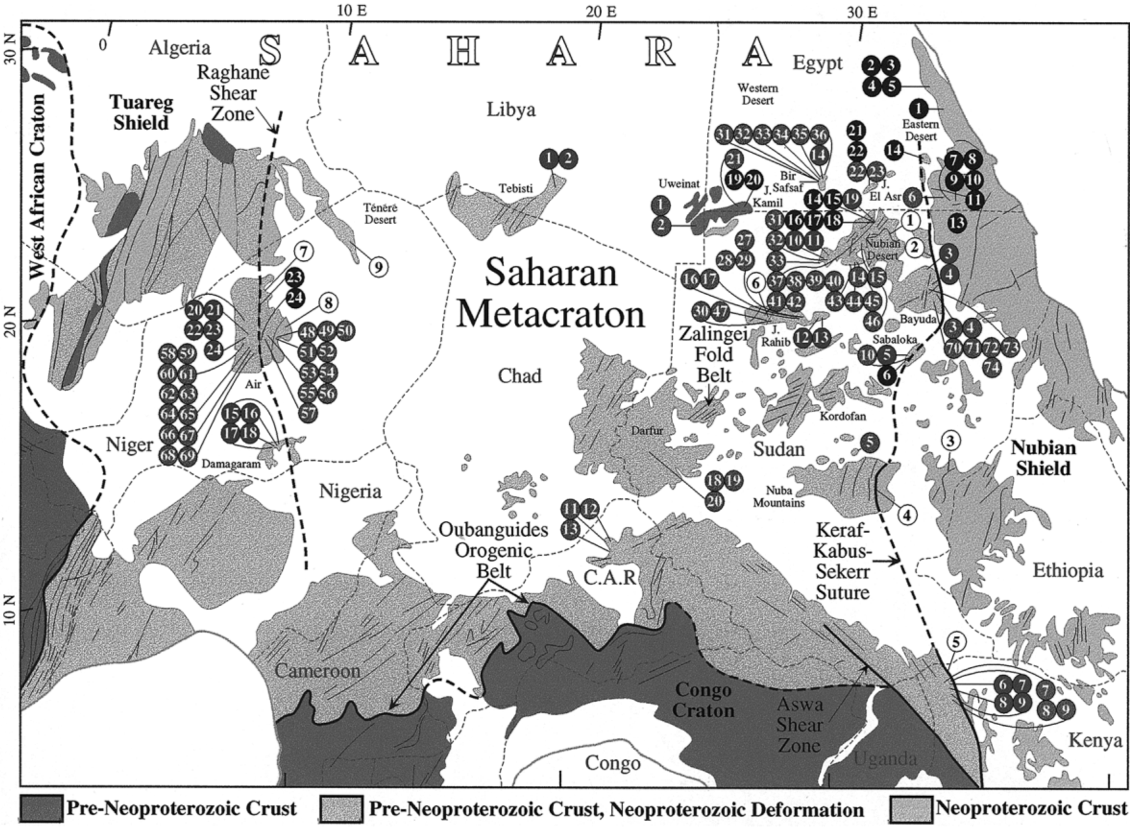

Fig. 2. The Saharan Metacraton (after Unrug, 1997) and location of geochronologic and isotopic data of the Saharan Metacraton which suggest the presence or involvement of pre-Neoproterozoic crust. Geochronological and isotopic data from Pegram et al. (1976), Klerkx and Deutsch (1977), Abdel-Monem and Hurley (1979), Dixon (1981), Harris et al. (1984), Kröner et al. (1987), Pin and Poidevin (1987), Schandelmeier et al. (1988), Wust (1989), Davidson and Wilson (1989), Key et al. (1989), Harms et al. (1990, 1994), Sultan et al. (1990, 1994), Black and Liégeois (1991), Stern et al. (1994), Liégeois et al. (1994) and Küster and Liégeois (2001). White circles indicate localites mentioned in the text. 1—Atmur-Delgo Suture; 2—Keraf Suture; 3—Ingessana Suture; 4—Kabus Suture; 5—Sekerr Suture; 6—Rahib Suture; 7—Barghot terrane; 8—Aouzegueuer terrane; 9—Edembo terrane. Black circles denote pre-Neoproterozoic U/Pb Zircon ages. 1—Wadi Rasheid gneiss; 2, 3, 4, and 5—Wadi Mubarak granitic conglomeratic cobble; 6—Sabaloka granulite; 7 and 8—Wadi Miyah greywacke; 9, 10 and 11—Wadi Allaqi greywacke; 12—Aswan granite; 13—Wadi Allaqi dike; 14, 15, 16, and 17—Duweishat gneiss; 18—Duweishat granitic conglomeratic cobble; 19 and 20—Jebel Kamil anorthosite; 21 and 22—Jebel Al Asr anorthosite; 23—Azan gneiss; 24—Raghane tonalite. Red circles denote I_{sr} higher than 0.704. 1 and 2—Ben Ghnema granite; 3—Shallal granite; 4—Deifallab granite; 5—Sabaloka granite; 6—Poloi granite; 7—Morupusti granite; 8—Lauraki granite; 9—Lolmungi granite; 10 and 11—Nubian Desert granite; 12 and 13—Wadi Hower granite; 14—Bir Safsaf granite; 15—Birnin Kazoe granite; 16—Moha granite; 17—Tyanza granite; 18—Dakouse granite; 19—Wadi Halfa granite; 20 and 21—Renatt leucogranite; 22—Iferouane monzogranite; 23 and 24—Taggar monzogranite. Blue circles denote T_{DM} ages older than 1500 Ma. 1 and 2—Karkur Murr charnockite; 3 and 4—Rahaba metasediment; 5—El Obeid granite; 6—Wadi Hafafit granitic conglomeratic cobble; 7 and 8—Marich granite; 9—Sekerr metapelite; 10—Sabaloka granite; 11—Mpoko charnockite; 12—Lere granulite; 13—Sibut Charnockite; 14—Nubian Desert gneiss; 15—Nubian Desert granite; 16—Wadi Hower migmatite; 17—Wadi Hower granite; 18, 19, and 20—Jebel Marra xenolith; 21—Jebel Kamil gneiss; 22 and 23—Jebel El Asr gneiss; 24—Nubian Desert gneiss; 25—Nubian Desert granite; 26—Nubian Desert granodiorite; 27—Wadi Hower migmatite; 28—Wadi Hower gneiss; 29—Wadi Hower granite; 30—Jebel Rahib gneiss; 31—Bir Safsaf gneiss; 32—Bir Safsaf diorite; 33 and 34—Bir Safsaf diorite enclave; 35—Bir Safsaf gabbro enclave; 36—Bir Safsaf basaltic dike; 37, 38, 39, 40, 41 and 42—Duweishat gneiss; 43—Nubian Desert granite; 44—Nubian Desert greywacke; 45—Nubian Desert cale-silicate; 46—Delgo ophiolite; 47—Jebel Rahib ophiolite; 48, 49 and 50—Eberjegui granodiorite; 51 and 52—Beurhot granite; 53, 54, 55 and 56—Takarakoum granite; 57—Tchebarlare granite; 58, 59, 60 and 61—Renatt leucogranite; 62, 63, 64 and 65—Iferouane monzogranite; 66 and 67—Teggar monzogranite; 68 and 69—Irsane graniote; 70, 71, and 72—El Melagi gneiss; 73 and 74—El Had metasediment.

data (Pegram et al., 1976; Klerkx and Deutsch, 1977; Meinhold, 1979; Abdel-Monem and Hurley, 1979; Dixon, 1981; Barth et al., 1983; Harris et al., 1984; Ries et al., 1985; Curtis and Lenz, 1985; Toteu et al., 1987; Kröner et al., 1987; Pin and Poidevin, 1987; Schandelmeier et al., 1988; Wust, 1989; Davidson and Wilson, 1989; Key et al., 1989; Harms et al., 1990, 1994; Black and Liégeois, 1991; Sultan et al., 1990, 1992, 1993, 1994; Stern and Dawoud, 1991; Stern et al., 1994; Liégeois et al., 1994; Küster and Liégeois, 2001) indicate that this crust is heterogeneous and contrasts with the Neoproterozoic juvenile crust of the Arabian-Nubian Shield to the east and the Archean continental crust of the Congo Craton in the south. This region constitutes pre-Neoproterozoic continental crust overprinted by Neoproterozoic tectonic events as well as containing Neoproterozoic juvenile material. These geochronological and isotopic data are interpreted as (1) indicating the existence of a craton prior to the Neoproterozoic orogenic events, but this was decratonized (Black and Liégeois, 1993) during Neoproterozoic time (Black and Liégeois, 1993; Liégeois et al., 1994); and/or (2) indicating the presence of a coherent continental crust prior to the Neoproterozoic orogenic events that was subjected to a Neoproterozoic extensional tectonic regime which produced rifting and subsequent oceanic basins which were subsequently closed, and as a result, drifted continental blocks collided with each other (Ghuma and Rogers, 1978; Dostal et al., 1985; El-Makhrouf, 1988; Schandelmeier et al., 1990, 1994; Abdel-Rahman et al., 1990; Pinna et al., 1994; Stern et al., 1994; Nzenti, 1998); or (3) Suggesting that this tract of continental crust constitutes, at least partly, a collage of exotic terranes which was assembled during Neoproterozoic time (Küster and Liégeois, 2001).

Precambrian rocks exposed in the above region form uplifted massifs within Cretaceous and younger cover rocks and have been collectively referred to as the Nile Craton (Rocci, 1965), the Sahara-Congo Craton (Kröner, 1977), the Eastern Saharan Craton (Bertrand and Caby, 1978), or the Central Saharan Ghost Craton (Black and Liégeois, 1993). The first three names are not precise in describing this crust in that (1) they used the term "craton" for a crust which did not act as a stable continental interior during the Neoproterozoic orogenic events; and (2) the "Nile" and "East Saharan" are too far to the east to define the geographic location and extent of this crust. The expression "Central Saharan Ghost Craton" satisfies the geographic extent and attempts to express the reworking nature of this crust. However, this term has not been precisely defined and has been loosely used afterwards. Moreover, in some cases the name "East Saharan Ghost Craton" (Liégeois et al., 1994) is used instead.

This paper: (1) Introduces and explains why the name "Saharan Metacraton" should be used to replace the previous names; (2) Defines and examines the nature of the boundaries of the Saharan Metacraton; (3) Reviews geochronological and isotopic geology data of the Saharan Metacraton as evidence for the presence of a pre-Neoproterozoic—but remobilized during Neoproterozoic time—continental crust; and (4) Discusses what happened to the Saharan Metacraton during Neoproterozoic time by examining the propositions of collision, decratonization, extensional tectonics and exotic terranes assembly as possible explanations.

2. What is a "Metacraton" and why "Saharan"?

The term "craton" refers to a stable part of the continent that has been only slightly deformed for a prolonged period (Bates and Jackson, 1980), hence acting as a stable interior during orogenic events at its margins. In Africa, continental blocks such as the Congo and West African Cratons (Fig. 1) behaved as true cratons during Neoproterozoic time. This can be linked to the acquisition of a thick lithospheric mantle, including a cold lower crust, strong crust–mantle coupling and an elastic thickness due to strong lithosphere that prevent later deformation (Black and Liégeois, 1993). Margins of present cratons behaved during Neoproterozoic orogenic events as passive margins. A craton may be decratonized when the passive margin turns into an active margin due to subduction initiation. In this case, when collision occurs, only traces of the craton will be preserved within the orogenic belt. There might be intermediate phases due, but not limited to, major thrusting events across passive margins inducing flexure of the cratonic lithosphere, lower crust fluxing or crust–mantle decoupling (Brown and Philipps, 2000) that could lead to decratonization, but retaining relics and/or isotopic inheritance of the former craton.

Using the term "craton" for the crust that occupies the north-central part of Africa is misleading because it implies that this lithosphere was stable during Neoproterozoic orogenic events, which was not the case. This continental block was at least partly, probably mostly, remobilized during Neoproterozoic time, although some coherence was maintained. Remobilization was in the form of deformation (Vail, 1971, 1972, 1976; Ghuma and Rogers, 1978; Grant, 1978; Annor and Freeth, 1985; Dumont et al., 1985; Dostal et al., 1985; Ngako, 1986; Toteu et al., 1987; Ekwueme, 1987; Schandelmeier et al., 1987; El-Makhrouf, 1988; Toteu, 1990; Stern et al., 1994; Sultan et al., 1994; Denkler et al., 1994; Harms et al., 1994; Abdelsalam et al., 1995, 1998, 2000; Nzenti, 1998; Küster and Liégeois, 2001), metamorphism (Kröner et al., 1987; Pin and Poidevin, 1987; Nzenti et al., 1988; Stern and Dawoud, 1991; Nzenti, 1992; Denkler et al., 1994), and emplacement of igneous bodies (Okujeni and Eduvie, 1985; Ghogomu et al., 1989; Harms et al., 1990, 1994; Stern et al., 1994; Sultan et al., 1990, 1992, 1994; Kusnir and Moutaye, 1997). Moreover, this crust might have been affected by extension to form restricted oceanic basins which subsequently closed, resulting in collision between drifted blocks (Ghuma and Rogers, 1978; Dostal et al., 1985; El-Makhrouf, 1988; Schandelmeier et al., 1990, 1994; Abdel-Rahman et al., 1990; Pinna et al., 1994; Stern et al., 1994; Nzenti, 1998). Nevertheless, geological, geochronological and isotopic evidence indicate that continental crust existed before Neoproterozoic time: (1) The region is dominated by medium to high-grade gneisses, migmatites and supracrustal rocks that are fundamentally different from the low-grade volcano-sedimentary-ophiolite assemblages which dominate the Neoproterozoic Arabian-Nubian Shield, even if some of these high-grade terrains appear to be Neoproterozoic juvenile rocks as in the Bayuda Desert (Küster and

Liégeois, 2001); and (2) Rb/Sr and U/Pb zircon ages as well as Nd model ages and initial Sr data indicate the presence of pre-Neoproterozoic continental crust (Klerkx and Deutsch, 1977; Abdel-Monem and Hurley, 1979; Dixon, 1981; Kröner et al., 1987; Wust, 1989; Davidson and Wilson, 1989; Harms et al., 1990, 1994; Sultan et al., 1990, 1992, 1993, 1994; Stern et al., 1994; Küster and Liégeois, 2001).

We introduce the term "metacraton" to refer to a craton that has been remobilized during an orogenic event but that is still recognizable dominantly through its rheological, geochronological and isotopic characteristics. The prefix "meta" in front of "craton" abbreviates "metamorphosis" in the general sense of the term and not only in the restricted geological meaning of "metamorphism". Remobilization of basement cratonic rocks has been discussed before plate tectonics within the context of geosynclinal theory (Stille, 1940, 1945; Khain and Scheinmann, 1962; Salop and Scheinmann, 1969). Terms such as "paraplatform" and "quasicraton" have been used in the past to describe basement reworking and "regenaration".

The previous names given to the Saharan Metacraton do not indicate its geographical extent. Rocci (1965) used the name "Nile" Craton but the Nile is in the extreme eastern part of the metacraton. Actually the course of the Nile in northern Sudan and southern Egypt follows the eastern boundary of the Saharan Metacraton (Stern and Abdelsalam, 1996). Kröner (1977) used the name "Saharan-Congo" Craton which implies that this crust is the northern continuation of the Congo craton. Geochronological and isotopic data indicate that the Saharan Metacraton is different from the Congo Craton in that the former, excluding unfolded sedimentary cover, does not contain Neoproterozoic material and is not overprinted by Neoproterozoic tectonic events (Toteu et al., 1987; Toteu et al., 1994; Pin and Poidevin, 1987; Toteu, 1990). Bertrand and Caby (1978) used the name "Eastern Sahara" Craton but the Saharan Metacraton also extends in central Sahara. Black and Liégeois (1993) were precise in using the name "Central Sahara" Ghost Craton in referring to the geographical extent of this crust. However, we propose the name "Saharan" for simplicity. Hence, we think that the name "Saharan Metacraton" is precise in referring to this tract of continental crust.

3. Geographical extent and outcropping nature of the Saharan Metacraton

The Saharan Metacraton extends from the Arabian-Nubian Shield in the east to the Tuareg Shield in the west and from the Congo Craton in the south to the Phanerozoic cover of the northern margin of the African continent in southern Egypt and Libya (Fig. 2). The metacraton occupies a rectangular-shaped region bounded by longitudes 8°30′ E and 35° E and latitudes 3° N and 24° N (Fig. 2) and covers more than 5,000,000 km^2. This poorly known region is just more than half the size of the USA.

More than 50% of the Precambrian rocks in the Saharan Metacraton are overlain by Cretaceous and younger unmetamorphosed and undeformed cover rocks or buried under sands of the Sahara Desert (Fig. 2). Precambrian rocks are exposed as uplifted massifs of different dimensions. The most prominent of these are Air, Tebesti, Uweinat, Bayuda, Nubian Desert, Kordofan, Darfur, and Nuba Mountains (Fig. 2). Smaller outcrops form inliers within cover rocks such as the Sabaloka in the Sudan, Bir Safsaf and Jebel El Asr in Egypt, and Damagaram in Niger (Fig. 2).

The southern part of the Saharan Metacraton extends from southern Sudan and northern Kenya, Uganda, and Congo, through Central African Republic to Cameroon and eastern Nigeria (Fig. 2). Geological, geochronological and isotopic data from Precambrian terranes of Kenya, Uganda, Congo, Chad, Central African Republic, Cameroon and eastern Nigeria are scarce but grossly similar to those from the northern part of the metacraton.

4. Lithology and structure

The Saharan Metacraton is dominated by medium to high-grade gneisses, metasediments, migmatites, and smaller outcrops of granulites, but low-grade volcano-sedimentary rocks are not uncommon. These metamorphic rocks are intruded by Neoproterozoic granitoids ranging in age between 750 and 550 Ma. The medium-to high-grade metamorphism suggests that these terrains could be pre-Neoproterozoic continental crust, but Rb/Sr and U/Pb zircon age determinations (Fig. 3) indicate that some, maybe most, of these gneisses were formed and/or metamorphosed during Neoproterozoic time. Nevertheless, Nd model ages and initial Sr isotopic data and some Rb/Sr and U/Pb zircon ages (Fig. 3) indicate involvement of a Paleoproterozoic or even Archean continental crust.

The presence of low-grade volcano-sedimentary rocks within the medium- to high-grade gneissic terranes has been considered as problematic in that the former are similar to those of juvenile Neoproterozoic rocks of the Arabian-Nubian Shield. Vail (1983, 1985) suggested that these outcrops are detached from the Arabian-Nubian Shield. The presence of the WSW-trending Atmur-Delgo ophiolite and low-grade volcano-sedimentary belt in the Nubian Desert of northern Sudan (Fig. 2; Schandelmeier et al., 1990; Harms et al., 1994; Denkler et al., 1994) has been explained as due to opening and

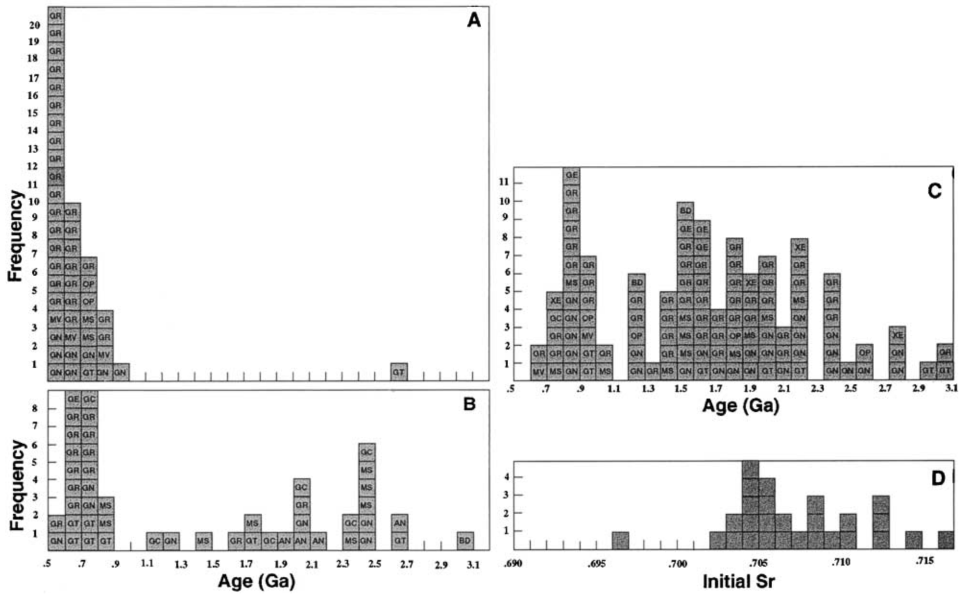

Fig. 3. Histogram representation of geochronologic and isotopic data of the Saharan Metacraton. (A) Rb/Sr and Sm/Nd ages. (B) Zircon U/Pb ages. (C) Nd model ages. (D) Sr initiais data. Data from Pegram et al. (1976); Klerkx and Deutsch (1977), Meinhold (1979), Abdel-Monem and Hurley (1979), Dixon (1981), Barth et al. (1983), Harris et al. (1984), Ries et al. (1985), Curtis and Lenz (1985), Toteu et al. (1987); Kröner et al. (1987), Pin and Poidevin (1987); Schandelmeier et al. (1988); Wust (1989); Davidson and Wilson (1989), Key et al. (1989), Harms et al. (1990, 1994), Sultan et al. (1990, 1994), Stern and Dawoud (1991), Black and Liégeois (1991); Stern et al. (1994), Liégeois et al. (1994) and Küster and Liégeois (2001). GT—Granulite, Charnockite, or Enderbite; GN—Gneiss and Migmatite; MS—Metasediment; MV—Metavolcanic; OP—Ophiolite; GR—Granite; GC—Granitic clast; GE—Granitic enclave; XE—Xenolith; BD—Basaltic dike.

closing of an aulacogen-like oceanic re-entrant which extended westward from the Mozambique ocean into the interior of the Saharan Metacraton (Stern, 1994; Stern et al., 1994). A similar explanation was adopted for low-grade volcano-sedimentary rocks in the Bayuda Desert of northern Sudan (Fig. 2; Abdelsalam et al., 1998). However, the dominantly juvenile Neoproterozoic age of the high-grade and low-grade rocks of the Bayuda Desert led Küster and Liégeois (2001) to conclude that the Bayuda is an accreted terrane on the eastern margin of the Saharan Metacraton. They explained the difference in metamorphic grade between the Bayuda terrane and the terranes of the Arabian-Nubian Shield as due to collisional and post-collisional horizontal movements. The presence of the N-trending Jebel Rahib ophiolite and low-grade sedimentary belt in Kordofan in central Sudan (Fig. 2) is explained as due to opening and closing of a Red Sea-like restricted oceanic basin without development of a subduction zone and magmatic arc (Abdel-Rahman et al., 1990; Schandelmeier et al., 1990). A similar explanation was given to explain the presence of low-grade volcano-sedimentary rocks in the Central African Republic (Dostal et al., 1985) and northern Cameroon (Pinna et al., 1994; Nzenti, 1998).

Two structural trends dominate the Saharan Metacraton; an early ENE–WSW trend and a younger N–S trend. The early ENE–WSW trend has been reported from many places in the Saharan Metacraton such as the Bayuda Desert (Vail, 1971, 1972; Abdelsalam et al., 1998), the Nubian Desert (Fig. 2; Schandelmeier et al., 1994; Stern et al., 1994; Abdelsalam et al., 1995), Kordofan and Darfur (Schandelmeier et al., 1987); Uweinat and Jebel El Asr inlier (Sultan et al., 1994). This trend has been referred to as the Zalingei folded zone (Vail, 1976; Schandelmeier et al., 1987) and is thought to continue SW into Cameroon where similar structures are reported (Dumont et al., 1985; Toteu et al., 1987; Toteu, 1990). This trend was interpreted in Central Africa to represent pre-Neoproterozoic structure (Grant, 1978; Annor and Freeth, 1985; Ngako, 1986; Ekwueme, 1987) but Toteu et al. (1987, 1994) obtained 630-620 Ma U/Pb zircon ages from granitoids deformed by these structures. These structures are interpreted by Küster and Liégeois (2001) as inherited from early Neoproterozoic terrane collisions.

The N-S trending structures in the Saharan Metacraton are similar in that they constitute N-trending upright folds deformed by N- to NW-trending strike-slip faults; both are more or less parallel to its eastern margin. In the northeastern part of the Saharan Metacraton, N-trending shear zones deform the dominantly medium to high-grade gneisses of the Nubian Desert in northern Sudan (Fig. 2; Abdelsalam et al., 2000). These are parallel to the Keraf suture (Fig. 2) which defines the eastern boundary of the metacraton if the Neoproterozoic Bayuda terrane has been thrust across it (Küster and Liégeois, 2001) as in the case of the Barghot and Aouzegueur terranes in the Tuareg Shield where they were thrust across the western boundary of the metacraton (Liégeois et al., 1994). The final deformation along the Keraf Suture is also defined by N-trending upright folds and NNW-trending sinistral strike-slip faults (Abdelsalam et al., 1995, 1998; Abdelsalam and Stern, 1996). Ar/Ar ages obtained from biotite and amphibole extracted from a granitic body deformed by sinistral strike-slip shearing along the Keraf suture indicates an age of 560 Ma for the end of this deformation. Similar N-trending structures are reported from as far west as northern Cameroon (Grant, 1978; Annor and Freeth, 1985; Ngako, 1986; Toteu et al., 1987; Ekwueme, 1987; Toteu, 1990). U/Pb zircon work by Toteu et al. (1987) suggests that these structures are younger than 620 Ma. Other N-trending structures in the Saharan Metacraton are in the form of fold and thrust belts such as those of the Tebisti massif in northern Chad and southern Libya and Jebel Rahib in Kordofan in northern Sudan (Fig. 2; Ghuma and Rogers, 1978; El-Makhrouf, 1988; Abdel-Rahman et al., 1990; Schandelmeier et al., 1990) and normal-slip shear zones such as those in the Nubian Desert (Fig. 2; Denkler et al., 1993, 1994; Harms et al., 1994; Stern et al., 1994). These fold and thrust belts have been interpreted as the product of opening and closing of oceanic basins (Ghuma and Rogers, 1978; El-Makhrouf, 1988; Abdel-Rahman et al., 1990; Schandelmeier et al., 1990), sometimes without the development of subduction zones (Schandelmeier et al., 1990; Abdel-Rahman et al., 1990). The normal-slip faults in the Nubian Desert have been interpreted as due to orogenic collapse following crustal thickening (Denkler et al., 1994). One explanation for the origin of the N-trending upright folds and strike-slip faults is offered by Abdelsalam et al. (2000) who advocated that far-reaching stresses due to E–W directed crustal shortening accompanying collision between the Saharan Metacraton and the Arabian-Nubian Shield. Küster and Liégeois (2001) considered some of these N–S trending faults as post-collisional deformation related to the ENE–WSW directed collision, where significant horizontal movements between terranes was accommodated. Association of collision and post-collision phases of deformation is comparable to the "hit and run orogeny" proposed by Maxson and Tikoff (1996).

5. Boundaries of the Saharan Metacraton

5.1. The eastern boundary

The eastern boundary of the Saharan Metacraton has been the focus of many studies. A tectonic boundary appears to separate it from lower grade, juvenile crust of the Neoproterozoic Arabian-Nubian Shield. Vail (1983, 1985, 1988) proposed that this boundary is marked by deformed sedimentary prisms overlying the medium to high-grade gneissic terrains, tracing it from the Sekerr region in northern Kenya, through the Ingessana region in east-central Sudan to the eastern Bayuda Desert in northern Sudan (Fig. 2). Abdelsalam and Dawoud (1991) argued that the boundary in central Sudan should be assigned to the Kabus Suture, which lies west of the Ingessana region (Fig. 2). The eastern boundary of the metacraton in northern Sudan is identified by Almond and Ahmed (1987) as the N-trending Keraf Suture. In the southern part of the Keraf Suture, Abdel-Rahman et al. (1993) interpreted a dismembered ophiolite belt as indicating opening and closing of a back-arc basin between the Saharan Metacraton and the Arabian-Nubian Shield. The northern part of the Keraf Suture is defined by N-trending upright folds which deform passive margin carbonate-rich turbidites (Abdelsalam et al., 1995, 1998). In contrast, the southern part of the Keraf suture is dominated by N- and NW-trending, sinistral strike-slip faults (Abdelsalam et al., 1995, 1998). The Keraf Suture truncates older E- to NE-trending structures of the Bayuda Desert (Abdelsalam et al., 1998). This led them to propose that the deformation in the Keraf Suture was due to sinistral transpression associated with oblique collision between the Arabian-Nubian Shield and the Saharan Metacraton. Concluding that the Bayuda Desert is dominantly Neoproterozoic juvenile crust, Küster and Liégeois (2001) suggest that the eastern boundary of the metacraton probably lies just west of, or along the Zalingei fold belt (Fig. 2), unless the Bayuda terrane is thrust across the metacraton.

The NE-trending Kabous ophiolitic melange in central Sudan (Fig. 2) separates the medium to high-grade gneissic terrane of the Nuba Mountains in the west from a low-grade volcano-sedimentary sequences of the Arabian-Nubian Shield to the east. The zone is characterized by imbricated ophiolitic fragments, arc volcanics, continental shelf conglomerates, and gneisses (Hirdes and Brinkmann, 1985; Brinkmann, 1986; Abdelsalam and Dawoud, 1991). The Kabous Suture is interpreted as due to closing of a marginal basin developed between the Saharan Metacraton and the Arabian-Nubian Shield (Abdelsalam and Dawoud, 1991).

The N-trending Sekerr Suture in northern Kenya (Fig. 2; Vearncombe, 1983; Shackleton, 1986; Ries et al., 1992) is a tectonic imbricate of arc volcanics and shelf sediments that was thrust, from east to west, across the eastern margin of the Saharan Metacraton (Fig. 2). Mosley (1993) proposed that the structures associated with the emplacement of the Sekerr ophiolite and the volcano-sedimentary sequence were modified by progressive shearing during collision between East and West Gondwana.

5.2. The western boundary

The western boundary of the Saharan Metacraton has been proposed on the basis of rheological considerations to lie east of the Tuareg shield (Liégeois et al., 1994, 2000). East of the N-trending Raghane Shear Zone, Barghot and Aouzegueur terranes (Fig. 2) have been thrust at ~700 Ma from west to east across a rigid cratonic block. These terranes include early molasse sediments, which were deposited due to flexure of a former passive cratonic margin. The cratonic shield protected these terrains during the 640–525 Ma late Neoproterozoic deformation that affected most of the Tuareg Shield. This suggests that the western boundary of the Saharan Metacraton is buried beneath thin-skinned allochthonous nappes including Neoproterozoic juvenile terranes and early molasse sediments. The easternmost Edembo terrane (Fig. 2) could either be part of the metacraton or former passive margin sediments back-thrust across it (Liégeois et al., 2000). Farther east, the metacraton disappears under the sands of Ténéré Desert (Fig. 2). East of the Ténéré Desert lies the poorly known Tebisti region of northern Chad and southern Libya. In this region granitoids of Neoproterozoic age have been reported (Pegram et al., 1976).

5.3. The southern boundary

The southern boundary of the Saharan Metacraton is not clearly defined. The geodynamic map of Gondwana (Unrug, 1996) shows it as partially defined by the NW-trending Aswa Shear Zone (Fig. 2). Further west Poidevin (1994) suggested that the boundary between the Congo Craton and the Saharan Metacraton is defined by the Oubanguides orogenic belt (Fig. 2). This belt trends NNW–SSE in northeastern Congo and E–W in the Central African Republic and Cameroon (Pin and Poidevin, 1987). The Oubanguides orogenic belt is bounded by a main basal thrust in the south and major dextral strike-slip fault in the north, both developed at 630 Ma (Pin and Poidevin, 1987). They interpreted N–S trending stretching lineations as indicating the direction of tectonic transport which produced S-verging thrusts along the collision zone between the Saharan Metacraton in the north and the Congo Craton to the south. Hence, the Oubanguides orogenic belt might be interpreted as a Pan-African mobile belt between the Congo Craton and the Saharan Metacraton. Hence, we consider the northern margin of the Oubanguides as the southern boundary of the Saharan Metacraton in Camerron and the Central African Republic. Pin and Poidevin (1987) also concluded, based on presence of a granulite belt that extends from Cameroon to southwestern Sudan and lack of calk-alkaline magmatism, that the Oubanguides represent a deeply eroded orogenic belt. West of Cameroon, the geodynamic map of Gondwana (Unrug, 1996) shows the Oubanguides orogenic belt (Fig. 2) extending into central Nigeria to define the northern margin of the Congo Craton (Fig. 2). Northern Nigeria is dominated by Neoproterozoic orogenic belts with older crustal element, hence may or may not be part of the Saharan Metacraton.

5.4. The northern boundary

The northern boundary of the Saharan Metacraton in southern Egypt and Libya is overlain by thick Phanerozoic sediments of the northern margin of the African continent.

6. Geochronology and isotope geology

Table 1 lists Rb/Sr and Sm/Nd ages, U/Pb zircon ages, T_{DM} ages, and initial Sr data for rocks of the Saharan Metacraton. Histograms of ages and initial Sr initial ratios are shown in Fig. 3. Fig. 2 shows the location of U/Pb zircon and T_{DM} ages and initial Sr data indicating the presence or involvement of pre-Neoproterozoic continental crust within the Saharan Metacraton.

6.1. Rb/Sr and Sm/Nd ages

The only Rb/Sr age indicating the presence of Paleoproterozoic to Archean continental crust came from a charnockitic sample from Uweinat (Fig. 2) where a whole-rock isochron age of 2617 ± 221 was reported by Klerkx and Deutsch (1977). Other Rb/Sr age determinations (Pegram et al., 1976; Meinhold, 1979; Barth et al., 1983; Ries et al., 1985; Curtis and Lenz, 1985; Kröner et al., 1987; Key et al., 1989; Harms et al., 1990, 1994; Stern and Dawoud, 1991; Black and Liégeois, 1991; Stern et al., 1994; Liégeois et al., 1994) give Neoproterozoic ages ranging between 550 and 1000 Ma (Table 1; Fig. 3A). Rb/Sr ages of granitoids approximate the ages of emplacement since these rocks did not experience metamorphism which is high enough to reset the isotopic System. On the other hand, Rb/Sr ages of gneisses, metasedimentary rocks and metavolcanic rocks are more likely to indicate the age of metamorphism.

Few Sm/Nd isochron ages have been determined for rocks of the Saharan Metacraton. Harms et al. (1994) obtained a whole rock Sm/Nd age of 752 ± 48 Ma and internal mineral isochron Sm/Nd age of 707 ± 54 Ma for gabbro from the Delgo ophiolite (Table 1). Küster and Liégeois (2001) obtained a whole-rock isochron age of 806 ± 19 Ma for juvenile oceanic island arc metavolcanics from the Bayuda Desert. They interpret this

Table 1
Geochronological and isotopic data of the Saharan Metacraton

Sample	Location	Rb/Sr	I_{Sr}	Sm/Nd	U/Pb	TDM	Reference
Ben Ghnema Granite	Tebesti, Libya	586 ± 27	0.7052 ± 0.0006(**1**)				Pegram et al. (1976)
”	”	550 ± 11	0.7062 ± 0.0004(**2**)				”
Karkur Murr Charnokite	Uweinat, Libya	2617 ± 21	?				Klerkx and Deutsch (1977)
Wadi Rasheid Gneiss	E Desert, Egypt				1770 ± 40(**1**)		Abdel-Monem and Hurley (1979)
Rahaba Gneiss	Bayuda, Sudan	874 ± 33	?				Meinhold (1979)
Kurmut Gneiss	”	757 ± 29	?				”
Nabati Granite	”	573 ± 72	?				”
Wadi Mubarak Cobble	E Desert, Egypt				2300 ± ?(**2**)		Dixon (1981)
Wadi Mubarak Cobble	E Desert, Egypt				1120 ± ?(**3**)		Dixon (1981)
”	”				1860 ± ?(**4**)		”
”	”				2060 ± ?(**5**)		”
Nabati Granite	Bayuda, Sudan	573 ± 70					Barth et al. (1983)
Karkur Murr Charnockite	Uweinat, Libya					3000(**1**)	Harris et al. (1984)
”	”					3200(**2**)	”
Rahaba Metasediment	Bayuda, Sudan					1500(**3**)	”
”	”					1550(**4**)	”
Abu Harrik Granite	”					750	”
Abu Hamed Quartzite	”					1050	”
Rashad Granite	Nuba Mts., Sudan					1000	”
Abbasiya Granite	”					950	”
El Obeid Granite	Kordofan, Sudan					2000(**5**)	”
Hafafit Cobble	E Desert, Egypt					750	”
Wadi Haimur metasediment	”					1800(**6**)	”
Jebel Zabara Granite	”					850	”
Aswan Granite	”					950	”
Marich Granite	Northwest Kenya					1650(**7**)	”
”	”					1950(**8**)	”
Sekerr Metapelite	”					1100	”
”	”					1500(**9**)	”
Sekerr Amphibolite	”					1000	”
Jebel Ed Dair Granite	Nuba Mts., Sudan	542 ± 12	0.6970 ± 0.0013				Curtis and Lenz (1985)
Jebel Liri Granite	Nuba Mts., Sudan	684 ± 05	0.70337 ± 0.0045				”
Abu Harik granite	Bayuda, Sudan	898 ± 51	0.7025 ± 0.0001				Ries et al. (1985)
Abu Hamed Quartzite	”	761 ± 22	0.7028 ± 0.0001				”
Shallai Granite	”	549 ± 12	0.7040 ± 0.0001(**3**)				”
Deifallab Granite	”	678 ± 43	0.7082 ± 0.0001(**4**)				”
El Kuro Metavolcanic	”	800 ± 83	0.7030 ± 0.0062				”
Sabaloka Granulite	Sabaloka, Sudan	720 ± 72	0.7044 ± 0.0003		2650 ± ?(**6**)		Kröner et al. (1987)
Sabaloka Migmatite	”	572 ± 16	0.7068 ± 0.0015				”
Sabaloka Granite	”	543 ± 26	0.7100 ± 0.0007(**5**)			1700(**10**)	”
Mpoko Charnockite	Oubanguides, C.A.R.				833 ± 66	1690(**11**)	Pin and Poidevin (1987)
Lere Granulite	”				652 ± 19	2210(**12**)	”
Sibut Charnockite	”				639 ± 03	2010(**13**)	”
Nubian Desert Gneiss	Nubian Des., Sudan					2230(**14**)	Schandelmeier et al. (1988)

(*continued on next page*)

Table 1 (*continued*)

Sample	Location	Rb/Sr	I_{Sr}	Sm/Nd	U/Pb	TDM	Reference
Nubian Desert Granite	"					1500(**15**)	"
Wadi Hower Migmantite	Kordofan, Sudan					1850(**16**)	"
Wadi Hower Granite	"					1600(**17**)	"
Wadi Miyah Greywacke	E Desert, Egypt				831 ± 20		Wust (1989)
"	"				2410 ± 100(**7**)		"
"	"				$2300 \pm ?$(**8**)		"
Wadi Allaqi Greywacke	"				820 ± 50		"
"	"				1460 ± 100(**9**)		"
"	"				2400 ± 140(**10**)		"
"	"				2450 ± 10(**11**)		"
Jebel Marra Xenolith	Darfur, Sudan					2210(**18**)	Davidson and Wilson (1989)
"	"					1930(**19**)	"
"	"					2820(**20**)	"
"	"					790	"
Mukogodo Migmatite	North-central Kenya	1206 ± 96	0.7035 ± 0.0002				Key et al. (1989)
Lolkoitoi Gneiss	"	622 ± 09	0.7033 ± 0.0001				"
Kotim Gneiss	"	818 ± 48	0.703 ± 0.0003				"
Poloi Granite	"	826 ± 33	0.704 ± 0.0015(**6**)				"
Morupusi Granite	"	572 ± 20	0.7045 ± 0.0006(**7**)				"
Lauraki Granite	"	582 ± 14	0.7044 ± 0.0002(**8**)				"
Lolmungi Granite	"	566 ± 30	0.7042 ± 0.0004(**9**)				"
Jebel Karail Gneiss	Uweinat, Egypt	673 ± 56	0.7050 ± 0.0004			2100(**21**)	Harms et al. (1990)
Jebel El Asr Gneiss	W Desert, Egypt					2500(**22**)	"
"	"					1900(**23**)	"
Nubian Desert Gneiss	Nubian Des., Sudan	918 ± 40	0.7161 ± 0.0007			2200(**24**)	"
Nubian Desert Granite	"	623 ± 37	0.7055 ± 0.0018(**10**)			1500(**25**)	"
"	"	565 ± 08	0.7064 ± 0.0002(**11**)			1200	"
Nubian Desert Granodiorite	"					1600(**26**)	"
Wadi Hower Migmatite	Kordofan, Sudan	686 ± 26	0.7064 ± 0.0008			1600(**27**)	"
Wadi Hower Gneiss	"					1700(**28**)	"
Wadi Hower Granite	"	562 ± 07	0.7080 ± 0.0040(**12**)			1400	"
"	"	585 ± 19	0.7086 ± 0.0008(**13**)			1800(**29**)	"
Jebel Rahib Gneiss	"					2600(**30**)	"
Bir Safsaf Gneiss	W Desert, Egypt					1500(**31**)	"
Bir Safsaf Granite	"	578 ± 09	0.7055 ± 0.0001(**14**)			1410	"
Bir Safsaf Diorite	"					1500(**32**)	"
Bir Safsaf Diorite Enclave	"					1610(**33**)	"
"	"					1640(**34**)	"
Bir Safsaf Gabbro Enclave	"					860	"
"	"					1580(**35**)	"
Bir Safsaf Basaltic Dike	"					1540(**36**)	"
"	"					1260	"
Nakhil Granite	W Desert, Egypt				578 ± 15	690	Sultan et al. (1990)
Aswan Granite	"				$1600 \pm ?$(**12**)	984	"
Roniers Valley Gneiss	Northern Cameroon				630 ± 05		Toteu (1990)
Jebel Moya Charnockite	Jebel Moya, Sudan				742 ± 02	970	Stern and Dawoud (1991)

Jebel Moya Granite	”	730 ± 31		0.7031 ± 0.0001	744 ± 02	980	”
Jebel Moya Enderbite	”				739 ± 02	960	”
Birnin Kazoe Granite	Damagaram, Niger	580 ± 69		0.7163 ± 0.0014(**15**)			Black and Liégeois (1991)
Moha Granite	”	530 ± 44		0.7051 ± 0.0008(**16**)			”
Tyanza Granite	”	556 ± 63		0.7120 ± 0.0050(**17**)			”
Dakousa Granite	”	579 ± 14		0.7125 ± 0.0008(**18**)			”
Wadi Allaqi Dike	”				3000 ± ?(**13**)		Kröner et al. (1992)
Wadi Halfa Metavolcanic	Nubian Des., Sudan	653 ± 20		0.7025 ± 0.0002			Stern et al. (1994)
”	”	581 ± 38		0.7031 ± 0.0005		660	
Wadi Halfa Granite	”	530 ± 10		0.7075 ± 0.0008(**19**)		720	”
Duweishat Gneiss	”				1232 ± 13(**14**)	2420(**37**)	”
”	”				2428 ± 07(**15**)	2810(**38**)	”
”	”				719 ± 14	1900(**39**)	”
”	”					1260	”
”	”				2406 ± 10(**16**)	2820(**40**)	”
”	”				1744 ± 11(**17**)	2400(**41**)	”
”	”					2030(**42**)	”
Duweishat Cobble	”				718 ± 15		”
”	”				2451 ± 05(**18**)		”
Jebel Kamil Anorthosite	Uweinat, Egypt				2629 ± 20(**19**)		Sultan et al. (1994)
”	”				2063 ± 08(**20**)		”
Jebel El Asr Anorthosite	W Desert, Egypt				2141 ± ?(**21**)		”
”	”				1922 ± ?(**22**)		”
SW Aswan Granite	”				741 ± 03		”
Aswan Granite	”				634 ± 04		”
Jebel Umm Shagir Granite	”				626 ± 04		”
Nubian Desert Migmatite	Nubian Des., Sudan			546 ± 19		830	Harms et al. (1994)
Nubian Desert Gneiss	”			592 ± 16	749 ± 12	1280	”
”	”					870	”
Nubian Desert Granite	”				718 ± 11	1620(**43**)	”
”	”				680 ± 12	800	”
”	”				661 ± 12	870	”
”	”					810	”
Nubian Desert Greywacke	”					1940(**44**)	”
Nubian Desert Cale-silicate	”			702 ± 27		1520(**45**)	”
”	”					1480	”
Delgo Ophiolite	”			752 ± 48		470	”
”	”			707 ± 54			”
”	”					1270	”
”	”					1810(**46**)	”
”	”					950	”
Jebel Rahib Ophiolite	Kordofan, Sudan					2650(**47**)	”
Azan Gneiss	Air, Niger				2086 ± 195(**23**)		Liégeois et al. (1994)
Raghane Gneiss	”				581 ± 48		”
Rhyolite Dyke	”	529 ± 23	0.7065 ± 0.0003				”
Raghane Tonalite	”				2081 ± 64(**24**)		”
Emzeggar Monzogranite	”				714 ± 19		”

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Table 1 (*continued*)

Sample	Location	Rb/Sr	I_{Sr}	Sm/Nd	U/Pb	TDM	Reference
Dabaga Forest Monzogranite	”				643 ± 11		”
Azan Gneiss Pendant	”				600 ± 28		”
Eberjegui Granodiorite	”					1627(**48**)	”
”	”					1985(**49**)	”
”	”					2027(**50**)	”
Beurhot Granite	”					1757(**51**)	”
”	”					1217	”
”	”					1601(**52**)	”
Takarakoum Granite	”					1816(**53**)	”
”	”					1865(**54**)	”
”	”					1616(**55**)	”
”	”					1886(**56**)	”
Tchebarlare Granite	”					2431(**57**)	”
”	”					1324	”
”	”					1461	”
Renatt Leucogranite	”	666 ± 11	0.7121 ± 0.0004(**20**)			1681(**58**)	”
”	”	666 ± 124	0.5106 ± 0.0001(**21**)			2279(**59**)	”
”	”					2496(**60**)	”
”	”					3187(**61**)	”
Iferouane Monzogranite	”	602 ± 14	0.7096 ± 0.0003(**22**)			1766(**62**)	”
”	”					1908(**63**)	”
”	”					2200(**64**)	”
”	”					1855(**65**)	”
Teggar Monzogranite	”	611 ± 11	0.7145 ± 0.0001(**23**)			2117(**66**)	”
”	”	586 ± 93	0.7183 ± 0.0014(**24**)			2011(**67**)	”
Irsane Granite	”					2495(**68**)	”
”	”					2102(**69**)	”
Dam Et Tor Amphibolite	Bayuda, Sudan			806 ± 19		903	Küster and Liégeois (2001)
Dam Et Tor Gneiss	”					858	”
”	”					899	”
Kurmut Gneiss	”					888	”
Kurmut Metasediment	”					787	”
Kurbi Metasediment	”					893	”
El Melagi Gneiss	”					2427(**70**)	”
”	”					2229(**71**)	”
”	”					2041(**72**)	”
El Had Metasediment	”					2244(**73**)	”
”	”					2084(**74**)	”

Data from Pegram et al. (1976), Klerkx and Deutsch (1977), Meinhold (1979), Abdel-Monem and Hurley (1979), Dixon (1981), Barth et al. (1983), Harris et al. (1984), Ries et al. (1985), Curtis and Lenz (1985), Toteu et al. (1987, 1994), Kröner et al. (1987), Pin and Poidevin (1987), Schandelmeier et al. (1988), Wust (1989), Davidson and Wilson (1989), Key et al. (1989), Harms et al. (1990, 1994), Sultan et al. (1990, 1994), Stern and Dawoud (1991), Black and Liégeois (1991), Stern et al. (1994), Liégeois et al. (1994) and Küster and Liégeois (2001). Numbers between brackets corresponds to those in Fig. 2.

isochron as manifesting a homogeneous depleted mantle source. These ages are interpreted to reflect crystallization ages since the Sm/Nd isotopic systematics are not easily reset during greenschist facies metamorphism.

6.2. U/Pb zircon ages

The first pre-Neoproterozoic U/Pb zircon age in the area was reported by Abdel-Monem and Hurley (1979) from Wadi Abu Rosheid in the Eastern Desert of Egypt (Fig. 2). They reported an age of 1770 ± 40 Ma for detrital zircons extracted from a psammitic gneissic sample and interpreted it as the age of the source region that supplied the detritus. Dixon (1981) reported U/Pb zircon ages of 1120, 1860, 2060, and 2300 Ma from granitic cobbles from a volcanoclastic and greywacke sedimentary sequence in the Wadi Mubarak area in the Eastern Desert of Egypt (Fig. 2). Similarly, these are interpreted to indicate the age of the source of the cobbles. Kröner et al. (1987) reported a U/Pb age of 2650 Ma from detrital zircon in granulite rocks of the Sabaloka inlier in northern Sudan (Fig. 2). This age is interpreted to indicate the presence of a pre-Neoproterozoic crust farther west. Wust (1989) reported pre-Neoproterozoic U/Pb single zircon evaporation ages ranging between 1460 and 2450 Ma for zircons extracted from greywackes in Wadi Miyah and Wadi Allaqi in the Eastern Desert of Egypt (Fig. 2) and interpreted these as indicating the age of the source rock. Stern et al. (1994) reported a single zircon U/Pb evaporation age of 2451 ± 5 Ma for a granitic cobble extracted from conglomerate near Wadi Halfa in the Nubian Desert in northern Sudan. This age was interpreted once more as indicating the age of the source rock.

U/Pb zircon ages which indicate the presence of "in situ" pre-Neoproterozoic crust in the Saharan Metacraton came from Sultan et al. (1994), Stern et al. (1994) and Liégeois et al. (1994). Sultan et al. (1994) reported U/Pb zircon ages of 2629 ± 3 and 2063 ± 8 Ma for a gabbroic anorthosite sample from Jebel Kamil immediately east of Uweinat (Fig. 2) as well as $\sim$2141 and $\sim$1922 Ma for an anorthosite sample from Jebel El Asr close to the Nile in southern Egypt (Fig. 2). Stern et al. (1994) reported single zircon Pb/Pb evaporation ages ranging between 1232 and 2428 Ma for orthogneisses near Wadi Halfa in the Nubian Desert of northern Sudan (Fig. 2). They concluded that these ages reflect several episodes of crust formation and/or reworking of older crust, another indication of the Saharan Metacraton. Liégeois et al. (1994) reported U/Pb zircon ages of 2086 ± 195 Ma from gneisses and 2081 ± 62 from tonalites of the Air massif in northern Niger.

Other U/Pb zircon ages from the Saharan Metacraton (Table 1; Fig. 3) dominantly indicate crust formation during Neoproterozoic time (Toteu et al., 1987; Pin and Poidevin, 1987; Wust, 1989; Sultan et al., 1990, 1994; Stern and Dawoud, 1991; Stern et al., 1994).

6.3. Nd model ages

Many Nd model ages were determined for different rock types in the Saharan Metacraton (Table 1; Fig. 3C; Harris et al., 1984; Kröner et al., 1987; Pin and Poidevin, 1987; Schandelmeier et al., 1988; Davidson and Wilson, 1989; Harms et al., 1990, 1994; Sultan et al., 1990, 1994; Stern and Dawoud, 1991; Stern et al., 1994; Liégeois et al., 1994; Küster and Liégeois, 2001). These range between 600 and 3200 Ma (Table 1; Fig. 4C). These data clearly indicate the heterogeneous nature of the Saharan Metacraton. Most of the analyses (66%) indicate the involvement of pre-Neoproterozoic continental crust in the formation of the Saharan Metacraton. It is worth noting here that Neoproterozoic ages were reported from as far as the Jebel Merra in the center of the metacraton (Fig. 2) where upper crustal xenoliths enclosed in Tertiary basait gave a T_{DM} age of 790 Ma (Table 1).

6.4. Initial Sr isotopic data

Initial Sr isotopic data for granitoids from the Saharan Metacraton ranging in age between 1000 and 550 Ma (Pegram et al., 1976; Meinhold, 1979; Barth et al., 1983; Ries et al., 1985; Curtis and Lenz, 1985; Key et al., 1989; Harms et al., 1990, 1994; Stern and Dawoud, 1991; Black and Liégeois, 1991; Stern et al., 1994; Liégeois et al., 1994; Küster and Liégeois, 2001) are dominantly higher than 0.704 (Table 1; Fig. 3D). The relatively high initial Sr is consistent with the involvement of an old continental crust component in the melt from which these granites were crystallized.

7. What happened to the Saharan Metacraton during the neoproterozoic?

Four models have been proposed to explain remobilization of the Saharan Metacraton during Neoproterozoic time: (1) collision; (2) delamination of the sub-continental mantle lithosphere; (3) extension; and (4) assembling of the metacraton from exotic terranes. We will show that these models are not mutually exclusive.

7.1. Collision processes in the Saharan Metacraton

Remobilization of the Saharan Metacraton during Neoproterozoic time through collision processes has been discussed by Schandelmeier et al. (1988) and Stern et al. (1994) based on observations in the northeastern part and along the eastern margin of the metacraton. There are at least two collision zones along the eastern and southern margins of the metacraton. The eastern margin has been defined by the N-trending Keraf-Kabus-Sekerr Suture (Fig. 2), thought to represent an arc-continental suture between the Saharan Metacraton and the Arabian-Nubian Shield (Vail, 1983, 1985; Almond and Ahmed, 1987; Abdelsalam and Dawoud, 1991; Abdel-Rahman et al., 1993; Abdelsalam et al., 1995, 1998; Abdelsalam and Stern, 1996). This can only be the case if the Bayuda terrane is thrust across the metacraton (Küster and Liégeois, 2001). On the other hand, the ENE–WSW trending Oubanguides orogenic belt (Fig. 2; Pin and Poidevin, 1987) can be interpreted as resulting from continent–continent collision between the Congo Craton and the Saharan Metacraton. This suggests that the Saharan Metacraton could have not been too far north if the former Neoproterozoic juvenile terranes between the two cratons had escaped. Scarce geochronologic data (Pin and Poidevin, 1987; Abdelsalam et al., 1998) constrained collision along these two sutures at ~630–560 Ma. In addition, structural styles along the two sutures can be interpreted as due to an overall NW–SE collision which was oblique to both the Keraf-Kabus-Sekker and Oubanguides orogenic fronts. This oblique collision produced sinistral transpression along the N-trending eastern margin of the metacraton, but was resolved as a south-verging fold and thrust belt and dextral strike-slip faults along the northern margin of the Congo Craton. Such collision might have been responsible for crustal thickening within the Saharan Metacraton where deformation related to collision was accommodated in the form of NE-trending (Vail, 1971, 1976; Grant, 1978; Annor and Freeth, 1985; Dumont et al., 1985; Ngako, 1986; Toteu et al., 1987; Ekwueme, 1987; Schandelmeier et al., 1987, 1994; Toteu, 1990; Sultan et al., 1994; Abdelsalam et al., 1995, 1998; Küster and Liégeois, 2001) and N-trending fold belts followed by sinistral strike-slip faults (Grant, 1978; Annor and Freeth, 1985; Ngako, 1986; Toteu et al., 1987; Toteu, 1990; Ekwueme, 1987; Denkler et al., 1994; Harms et al., 1994; Stern et al., 1994; Abdelsalam et al., 2000) followed by N-trending normal-slip faults (Denkler et al., 1994; Abdelsalam et al., 2000). Crustal thickening might have triggered lower crustal melting and emplacement of A-type granites within the Saharan Metacraton (Okujeni and Eduvie, 1985; Ghogomu et al., 1989; Harms et al., 1990, 1994; Stern et al., 1994; Sultan et al., 1990, 1992, 1994; Kusnir and Moutaye, 1997). This can be a favourable framework for the second model.

7.2. Sub-continental mantle lithosphere delamination of the Saharan Metacraton

Delamination of the continental lithospheric mantle has been proposed by Bird (1979) for the Colorado plateau and by Houseman et al. (1981) for the Tibetan Plateau to explain the high heat flow, rapid uplift and extension in an orogenic setting. Ashwal and Burke (1989) argued that the continental lithospheric mantle beneath North Africa was separated from the crust after thickening as a result of Neoproterozoic collision. Black and Liégeois (1993) expanded on this and proposed that delamination under the Saharan Metacraton occurred at ~700 Ma. They suggested, based on absence of a craton to the east of the Tuareg Shield at the end of the Neoproterozoic orogenic events, that destabilization of the Saharan Metacraton was due to the loss of its thick mechanical boundary layer and that Uweinat is a surviving fragments of this metacraton. They used this to explain why the Neoproterozoic orogenic belts in northern Africa are characterized by high-*K* calc-alkaline late-kinematic granitoids and high-temperature, low-pressure metamorphism, both of which indicate high heat flow. This can be achieved when significant vertical shortening is added on the former cratonic passive margin through stacking of over-thrust terrains such as the present case of the Himalayas and the Indian cratonic plate. This could induce crust–mantle decoupling (Brown and Phillips, 2000) leading to crust–mantle delamination beneath the passive margin, as suggested for the northernmost Indian plate by Molnar (1988). Such delamination process (Black and Liégeois, 1993) will induce regional extension due to the collapse of the hot lower crust. This leads to the third model.

7.3. Extension tectonics within the Saharan Metacraton

Extension of the lithosphere has not been discussed fully as a mechanism for remobilizing the Saharan Metacraton during Neoproterozoic time. Lithospheric extension can subject broad crustal tracts to high heat flow, uplift and intrusion or underplating of melts from the asthenosphere as exemplified by the 1000 km wide Basin and Range Province of the USA (Eaton, 1982). Large-scale extension brings the asthenosphere nearer to the surface, leading to high heat flow, resulting in metamorphism and emplacement of A-type granites (Lum et al., 1989). Similar relationships have been inferred for the northeastern part of the Saharan Metacraton (Denkler et al., 1994). These events can be integrated into a model of post-collisional collapse following lithospheric delamination. Such large-scale extension thus might have contributed significantly in remobilizing the Saharan Metacraton during the Neoproterozoic.

On the other hand, opening and closing of Red Sea-type restricted oceans, has been suggested for the evolution of the Jebel Rahib ophiolite (Fig. 2) in the interior of the Saharan Metacraton (Schandelmeier et al., 1990;

Abdel-Rahman et al., 1990). In addition, the E–W trending Atmur-Delgo Suture is interpreted as the opening and closing of an oceanic re-entrant extending westward into the interior of the Saharan Metacraton (Schandelmeier et al., 1994; Stern et al., 1994; Abdelsalam et al., 1995). Alternatively, the evolution of these terrains can also be explained as due to terrane accretion onto the Saharan Metacraton followed by strike-slip shearing (Küster and Liégeois, 2001) as in the case of the Tebisti Massif (Fig. 2; Ghuma and Rogers, 1978; El-Makhrouf, 1988). This leads to the last model.

7.4. Assembling the metacraton from exotic terranes

The proposition that continental crust west of the Nile is a collage of exotic terranes, assembled during Neoproterozoic time, was first made by Rogers et al. (1978). This was based on the absence of cratonic continental crust and used to explain the heterogeneous nature of the crust west of the Nile. Küster and Liégeois (2001) pointed out that such scenario has been proposed for the Tuareg Shield which forms an orogenic collage of thrust and/or shear zone bounded crustal blocks of different provenance and age (Fig. 2; Black et al., 1994; Liégeois et al., 1994). These terranes include non-remobilized cratonic fragments of microcontinents and remobilized Paleoproterozoic crustal blocks as well as Neoproterozoic island arcs. However, Black and Liégeois (1993) and Liégeois et al. (1994) argued that for the Tuareg Shield to have been assembled in such a fashion during the Neoproterozoic, there must have been a rigid craton to the east. Hence, they favoured the proposition that the Saharan Metacraton was a coherent continental block prior to Neoproterozoic orogenic events in Africa.

Review of these four scenarios shows that they can be combined into a single model to explain the evolution of the Saharan Metacraton during Neoproterozoic time. The metacraton was subjected to major collisional events at least along two of its passive margins, leading to crustal thickening that culminated in regional lithospheric delamination. This triggered extension, and was accompanied by horizontal movements along shear zones, followed by high-temperature metamorphism as well as magma generation. Such a process can dismember, and locally create Neoproterozoic juvenile terranes within the metacraton.

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