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OPHIOLITES AND ISLAND ARCS IN THE LATE PROTEROZOIC NUBIAN SHIELD

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ABSTRACT

The late Proterozoic ophiolites of the Nubian shield can be classified into two groups. One is mainly found in shallow structural positions thrust over presumably older continental crust. The other group predominantly occurs in steep structural orientations within suture zones separating arc terranes. These structural differences reflect the classification of the first group as Tethyan-type and of the second group as Cordilleran-type. The Cordilleran-type ophiolites characterize the ensimatic realm of the Arabian-Nubian shield, whereas the occurrences of the Tethyan-type ophiolites are restricted to the presumed continental margin environment of the Archaean to mid Proterozoic Nile craton. Using the obduction ages of the ophiolites as time markers of ocean closure, the time span of the subduction-related igneous activity and thus the life time of the distinct arc terranes can be estimated to be about 150 Ma.

INTRODUCTION

The Arabian-Nubian shield (ANS, Fig. 1) is one of the largest exposed areas of late Proterozoic rocks where plate tectonic processes are known to have taken place (e.g. Kröner et al., 1987). The continental crust of the shield, which is exposed on both sides of the Red Sea, is believed to have been built up by island arc terranes and microplates welded together and accreted to the African craton. The arc accretion hypothesis is based on the predominant association of calc-alkaline igneous rocks and the occurrence of ophiolites. The ophiolites are interpreted to represent the obducted remnants of former ocean basins, once separating the arc terranes, that now occur as nappes or along ophiolite-decorated belts (e.g. Bakor et al., 1976; Neary et al., 1976; Greenwood et al., 1976; Kröner et al., 1987).

The ophiolites of the Arabian part of the shield have been studied by Kröner (1985) and Pallister et al. (1987; 1988) who suggested that they are arc-related ophiolites formed in an environment that has Phanerozoic analogues in Alaska or Indonesia. In Arabia and as well in the northernmost part of the Nubian shield, geochronological studies of the ophiolites showed that they are among the oldest rocks of the arc terranes with which they are associated (Kröner et al., 1992). This study also demonstrated that the formation of oceanic crust accompanied the subduction-related formation of the arc terranes from the beginning of the so-called Pan-African evolution (Kennedy, 1964; Kröner, 1984) of the Arabian-Nubian shield.

The palaeogeographic reconstruction of the Arabian shield provides evidence for the existence of arcs and a pre-Pan-African (>950 Ma) microcontinent, the Afif terrane. These are now separated by ophiolite-decorated suture zones (Camp, 1984; Stoeser and Camp, 1985). The eastern part of the Nubian shield is considered to consist of arc terranes bordered by a Proterozoic to Archaean craton to the west (Kröner et al., 1986). This pre-Pan-African continent west of the Nubian shield is known from various studies (e.g. Klerx and Deutsch, 1977), although the exact postition of the ancient continental margin is not precisely known. This craton was named the East Sahara craton by Kröner (1979) or Nile craton by Rocci (1965), and the latter term will be used in the following.

The geodynamic position of ophiolites emplaced along a continental margin is significantly different from those in an inter-arc setting, as demonstrated by Moores (1982). Therefore, it can be expected that ophiolites in Arabia and the inter-arc setting of the ANS differ from those obducted onto the margin of the Nile craton.

Moores (1982) has divided Phanerozoic ophiolites into Tethyan (T-type) and Cordilleran (C-type). The major difference between the two is that the Tethyan-type ophiolites such as the Semail ophiolite in Oman are structurally emplaced onto a crystalline basement and/or platform sediments and show a tectonized and metamorphic unit at their base. These parts of the ophiolite sequence are not developed in the Cordilleran-type ophiolites such as the Josephine ophiolite in the western United States. Besides this distinction, both ophiolite types are considered to contain all the members of an ophiolite sequence as defined in the Penrose conference (Anonymous, 1972). These are from bottom to top an ultramafic complex, a gabbroic complex, a mafic sheeted dyke complex and a mafic, commonly pillowed volcanic complex. Chromite layers, felsic igneous rocks and overlying sediments are associated with this rock assemblage.

The purpose of the present study is to present a synoptic view of the ophiolites of the Nubian shield including new field and isotopic data (Table 1 and 2). These ophiolites are then interpreted using the scheme of Moores (1982) and the geodynamic implications are reviewed.

OPHIOLITES AND SUTURE ZONES

The major ophiolite occurrences of the Nubian shield are described from north to south, i.e. from the Eastern Desert of Egypt through the Red Sea Hills and the central part of the Sudan to Ethiopia and Kenya (Fig. 1). This compilation is not complete, but most of the typical and well studied ophiolites are included. Not all mafic-ultramafic assemblages discussed in this study are proven remnants of

Table 1 - Sm-Nd analyses

Sample	Rock type	Sm (ppm)	Nd (ppm)	^{147}Sm/ ^{144}Nd	^{143}Nd/ ^{144}Nd	2 σ	εNd_t	Age (Ma)	Model age (Ga)
Gerf Terrane									
GG-40	Black Shale	1.62	8.47	0.1158	0.511830	24	-8.0	750	2.05
GG-41	Greywacke	1.99	11.54	0.1041	0.511660	18	-10.2	750	2.06
GG-57	Amphibolite	0.939	4.43	0.1281	0.512659	18	7.0	750	0.88
Abu Swayel									
EG-844	Metasediment	7.43	36.01	0.1247	0.511835	19	-8.9	732	2.24
EG-845	Metasediment	9.22	49.85	0.1118	0.511816	14	-8.1	732	1.99
EG-850	Amphibolite	2.42	10.53	0.1391	0.512663	23	5.9	732	1.00
EG-851	Amphibolite	8.22	40.2	0.1236	0.512021	19	-5.2	732	1.90
Onib Ophiolite									
SD-121	Ultramafic	0.073	0.097	0.4541	0.514363	53	7.0	810	0.77
ONH-43/3	Pillow Basalt	4.03	15.76	0.1545	0.512764	19	6.8	810	0.99
SD-189	Dyke	2.40	9.50	0.1530	0.512754	27	6.8	810	0.99
ONH-69/2	Pillow Basalt	2.67	8.84	0.1824	0.512948	13	7.5	810	0.98
Gabgaba Terrane									
AM-39	Basalt	6.26	26.24	0.1442	0.512689	17	7.0	887	1.01
PS-96	Greywacke	2.70	10.17	0.1562	0.512658	33	5.0	887	1.30
Nuba Mountains									
4071	Arkose	0.545	0.144	0.2290	0.512981	10	3.4	800	§-1.72
4081	Gabbro	3.79	11.62	0.1970	0.513006	26	7.2	800	1.32
4113	Chlorite Schist	2.88	11.44	0.1523	0.512565	13	3.1	800	1.45

§ Note that samples with flat or depleted REE-pattern can yield meaningless model ages.
Chondritic ratios used are ^{147}Sm/^{144}Nd=0.1967, ^{143}Nd/^{144}Nd=0.512638.
Model age is based on a linear mantle depletion with εNd_t=0 at t=4.5 Ga and εNd_t=10 at t t=0.

Table 2 - Rb-Sr analyses

Sample	Rock type	Rb (ppm)	Sr (ppm)	^{87}Rb/ ^{86}Sr	^{87}Sr/ ^{86}Sr	2σ	Age (Ma)	Sr_i
Abu Swayel								
EG-844	Metasediment	66.4	174	1.0934	0.715963	16	732	0.704538
EG-845	Metasediment	60.2	141	1.2192	0.717447	15	732	0.704708
EG-850	Amphibolite	6.0	544	0.0324	0.704001	23	732	0.703662
EG-851	Amphibolite	28.2	272	0.2952	0.709949	17	732	0.706865
Onib Ophiolite								
SD-121	Ultramafic	< 1 §	6 §		0.702995	14	810	0.702995
ONH-43/3	Pillow Basalt	3 §	179 §	0.0484	0.704664	15	810	0.704104
SD-189	Dyke	4 §	272 §	0.0425	0.702980	14	810	0.702488
OND-14/3	Gabbro	< 1 §	159 §		0.702756	22	810	0.702756
Gabgaba Terrane								
AM-39	Basalt	36.4	422	0.2496	0.706115	31	887	0.702951
PS-96	Greywacke	30.9	315	0.2862	0.705880	26	887	0.702253
*				0.2862	0.705867	27	887	0.702240
*				0.2862	0.705842	16	887	0.702215
Nuba Mountains								
4071	Arkose	0.46	20.8	0.0642	0.704819	19	800	0.704085
4081	Gabbro	1.07	69.5	0.0445	0.703538	18	800	0.703029
4113	Chlorite Schist	0.24	657	0.0010	0.702892	25	800	0.702880

* Duplicate analysis
§ XRF analysis

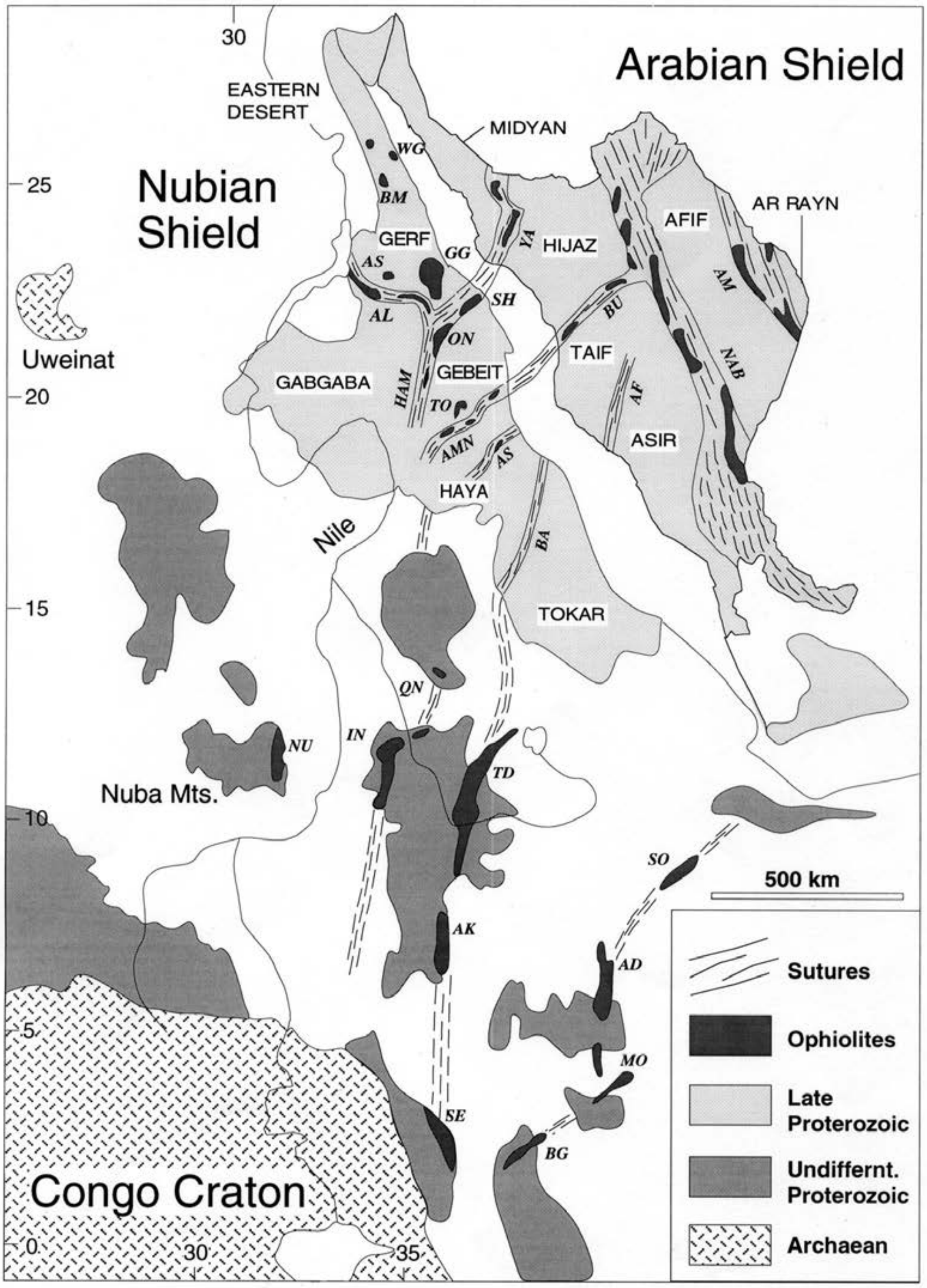

Fig. 1 - Geological sketch map of NE Africa and Arabia showing major terranes and suture zones, modified and compiled after Stoeser and Camp (1985), Vail (1985) and Kröner et al. (1987). Terranes are named in capitals. Abbreviations for the suture zones and ophiolites are (in italics) AD: Adola, AK: Akobo, AF: Afif, AL: Allaqi, AM: Al Amar, AMN: Amur-Nakasib, AS: Abu Swayel, ASH: Wadi Ashat, BA: Baraka, BM: Barramayia, BG: Baragoi, BU: Bir Umq, GG: Gebel Gerf, HAM: Hamisana, IN: Ingessana, MO: Moyale, NAB: Nabitah, NU: Nuba Mountains, ON:Onib, QN: Qala al Nahal, SE: Sekker, SO: Soka, SH: Sol Hamed, TD: Tulu Dimtu, TO: Jebel Tohado, WG: Wadi Ghadir, YA: Yanbu.

oceanic crust. Some comprise a complete and partly coherent sequence. Others are highly dismembered or incomplete or embedded in shear zones and strongly deformed. But even there internally intact fragments are still preserved and support an interpretation of such rocks as parts of an ophiolite sequence. The identification of the lithological units is also hampered by greenschist to amphibolite facies metamorphism that affected all ophiolitic and arc related rocks of the Nubian shield. The major rock types and field relations of the mafic-ultramafic assemblages of the Nubian shield that are considered as ophiolites are described in the following.

Wadi Ghadir

The northernmost ophiolites of the Nubian shield occur in the Eastern Desert of Egypt. One outstanding example is well exposed in the Wadi Ghadir area southwest of Marsa Alam. This ophiolite, although slightly dismembered, includes most rocks of an ophiolite sequence (El Bayoumi, 1980; 1984). An internally coherent section comprising layered gabbro, isotropic gabbro with rare plagiogranite, sheeted dykes, pillow lavas and cherts is ca. 2 km thick. The peridotites occur as fragments of variable sizes and reach dimensions of up to hundreds of metres within a mafic-ultramafic tectonic mélange. It has all ophiolite lithologies except the basement onto which the rocks were obducted and the post-emplacement deposits. The ophiolite units are intruded by calc-alkaline plutonic rocks and are overlain by calc-alkaline arc volcanics.

Kröner et al. (1992) dated a Wadi Ghadir plagiogranite by single zircon evaporation at 746±19 Ma. This is coeval with the 740-780 Ma old ophiolites from the northern Arabian shield (Pallister et al., 1987) and the northern Red Sea Hills (Kröner et al., 1992).

The nature of the footwall is still controversial, since there is no direct evidence for an older continental basement beneath the Eastern Desert (Kröner et al., 1988). The low Sr initial ratios of 0.702 - 0.704 for most of the Eastern Desert rocks (e.g. Stern and Hedge, 1985) support an ensimatic environment. There is, however, strong indirect evidence for the close proximity of a craton, as indicated by Pb isotopes (Stacey and Stoeser, 1983; Sultan et al., 1992) and by the ca. 1500 Ma old detrital zircons (Kröner et al., 1988; Wust et al., 1987).

The structural position of the ophiolites is more or less horizontal as suggested by the almost horizontal layering of the gabbro, the orientation of the pillow basalts and the vertical to subvertical orientation of the sheeted dykes. Ries et al. (1983) considered the Wadi Ghadir as part of a large shallow dipping nappe resting on a basement with continental affinities.

Gebel Gerf

The largest ophiolite complex of the ANS is the Gebel Gerf massif (Fig. 1). It belongs to a large nappe of an ultramafic mélange with minor fragments of gabbro and basalt in the north and east, and a massive layered gabbro, containing chromite lenses in the southwest (Zimmer et al., 1995). The associated Gebel Heiani and Gebel Hagar Zarga basalts contain pillow lavas and dykes. The basal tectonic contact of the Gerf ultramafic rocks is exposed as a thrust over a series of volcano-sedimentary rocks including turbiditic layers and black shales in the east and staurolite-bearing schists in the southeast. These indicate a mature sedimentary precursor. In places, the ultramafic mélange is overlain by a coarse-grained, poorly sorted post-obduction conglomerate that contains ophiolite-derived fragments such as gabbros, basalts and ultramafics.

The Gerf gabbroic section is geochemically similar to back-arc or subduction-related oceanic crust, whereas the pillows and dykes of Gebels Heiani and Hagar Zarga display geochemical features of modern N-MORB (Zimmer et al., 1995). The pillow lavas and dykes show LREE depletion, and the isotopic data also point to a depleted source with Sr^i of 0.7021 to 0.7032 and eNd^t of +6.5 to +8.8 (Zimmer et al., 1995).

A single zircon Pb/Pb evaporation age of 741±21 Ma from a plagiogranite (Kröner et al., 1992) and a Sm/Nd mineral age of 771±53 Ma (Zimmer et al., 1995) date the time of ophiolite formation.

The Sm/Nd isotope data of this study on the metasediments immediately east of and structurally below the ultramafic mélange, namely on a black shale, a greywacke and an amphibolite constrain the isotopic nature of the footwall rocks of the Gerf Nappe (Table 1). The amphibolite yielded an eNd^t of +7.0 inferring mantle derivation. The metasediments are continent-derived with eNd^t of -8.0 and -10.2. This suggests that a N-MORB-type ophiolite nappe was thrust onto a continental margin. The continent itself could not yet be identified.

Abu Swayel

In the Abu Swayel area, ca. 100 km east of Lake Nasser, small, highly sheared serpentinite bodies intruded by a gabbro-diorite suite are the remnants of a presumed ophiolite nappe. Structurally below and in fault contact with the serpentinite are metavolcanics and staurolite-bearing metasediments indicating a mature sedimentary precursor. Kröner et al. (1992) dated zircons from a gabbro intruding the serpentinite at 729±19 Ma and a diorite at 736±11 Ma which provide a minimum ages for ophiolite formation.

In this study, Sr and Nd isotopes were analyzed from metavolcanics and metasedimentary rocks that are associated with the ophiolitic rocks of the Abu Swayel section (Table 1 and 2). The continental origin of the metasediments is underlined by the isotopic data with eNd^t ranging from -8.1 to -8.9. One amphibolite with eNd^t of +5.9 has an ensimatic isotopic signature. The close association of crustal and juvenile material is well in line with the assumption of a continental margin onto which the Abu Swayel nappe was thrust.

Wadi Allaqi

The Wadi Allaqi belt forms a ca. 20 km wide and 200 km long shear zone between Lake Nasser and the Gebel Gerf ophiolite. A pervasive vertically-dipping foliation affected all the rock types of the belt, including metasediments, metavolcanics, granitoids and ophiolitic rocks. The latter comprise isotropic and layered gabbros, some minor pillow lavas and large, partially serpentinized ultramafic bodies containing chromite lenses. Volcanic rocks of the Allaqi belt are geochemically characterized as arc-type volcanics (Mankel et al., 1987). A Rb/Sr isotope age of 769±20 Ma with a Sr^i of 0.70216±16 (MSWD=5.4) indicate that these volcanics belong to a juvenile island arc (Mankel et al., 1987). Single zircon ages of 770±9 Ma and 3017±3 Ma were obtained for an intermediate to felsic dyke cutting a serpentinite lense in the westernmost part of the Wadi Allaqi belt where it enters Lake Nasser (Kröner et al., 1992). The older zircons are believed to be inherited from the continental margin of the Nile craton. This agrees well with Pb isotopic data for the Aswan granite (Stacey and Stoeser, 1983; Sultan et al., 1992) that indicate continental affinities. The tectonic position of the Allaqi belt is thus a composite one. To the south the Gabgaba and Gebeit terranes consist of volcanic arc assemblages, whereas to the north the Gerf terrane which has partly continental characteristics and to the west The Allaqi projects towards the ancient continental margin of the Nile craton.

Sol Hamed

The Sol Hamed ophiolite is the northeastern extension of the Onib suture and can be traced to the Yanbu suture in Arabia (Hussein, 1977; Fitches et al., 1983). It forms an up to 10 km wide and 20 km long northeast trending mafic-ultramafic assemblage. The sequence includes ultramafics, partly layered gabbros, and pillow lavas with pillow breccias and cherts. A small sheeted dyke complex occurs between the gabbros and the pillow lavas. Though the contacts between the individual units are tectonic the ophiolitic nature of the complex is undisputed.

The ophiolite is unconformably overlain to the southeast by a volcano-sedimentary series. The basal conglomerate of this series contains pebbles of the ophiolite indicating its post-obduction age. In the vicinity of the Sol Hamed ophiolite no continental basement has been recognized. The Gebeit terrane to the south is an intraoceanic island arc terrane (Reischmann and Kröner, 1994), and the portion of the Gerf terrane immediately to the north of it consists of immature sediments and arc-type volcanic and plutonic rocks. The arc-type volcanic rocks overlying the Sol Hamed ophiolite yielded a Rb/Sr whole-rock errorchron age of 712±58 Ma (Fitches et al., 1983). The Sr^i of 0.70234 of these volcanics is low, underlining their ensimatic origin.

Wadi Onib

The Wadi Onib ophiolite is superbly exposed in the Wadis Onib and Sudi (Hussein et al., 1984; Hussein, 1999). Members of the ophiolite form segments several km in size along a suture zone connecting the Hamisanah suture with the Sol Hamed ophiolite. The whole complex dips moderately to steeply to the east. In the Sudi section an internally coherent sequence consists of 2 km thick serpentinized peridotite, a 2.5 km thick transition zone with cumulates of spinel-bearing wherlites and chromite lenses, 3 km thick layered gabbro and 2 km isotropic gabbro on top. The sheeted dykes, the pillow lavas and some small lenses of banded cherts are exposed as tectonic slices in tectonic contact with ultramafics in the Onib section. The whole complex is considered to be a complete, although dismembered, ophiolite.

Isotopic data of selected rocks are given in Table 1 and 2. The Sm/Nd data of the mafic and ultramafic rocks define an isochron with an age of 810±55 Ma (Fig. 2). This age matches the Pb/Pb zircon age of 808±14 Ma of a plagiogranite (Kröner et al., 1992) and thus is interpreted as the igneous formation age. The eNd^t of +7.0 strongly suggests an intraoceanic origin without significant crustal contamination. Sr initial ratios of 0.7028-0.7041 also indicate a depleted source, even for those samples in which Rb could not be measured due to low concentration, and the measured $^{87}Sr/^{86}Sr$ ratios could therefore not be age-corrected.

The Onib ophiolite was intruded by granitoids with ages of 713±12 Ma and 714±5 Ma (Kröner et al., 1992) which marks the minimum age of obduction. Post-tectonic igneous activity is documented by the 646±10 Ma old granite from the Wadi Onib section (Kröner et al., 1992) and the 649±18 Ma old Asoteriba volcanics from the Gebeit terrane and 671±16 Ma old younger granites (Cavanagh, 1974).

The ophiolite is associated with volcanics that are chemically classified as island arc type (Reischmann, 1986). These are interlayered with immature clastic sediments and limestones and intruded by arc-related calc-alkaline batholithic granitoids. Geological and geochemical investigations indicate that the Wadi Onib ophiolite is a high-Ti type ophiolite, probably of back-arc basin origin (Hussein, 1999).

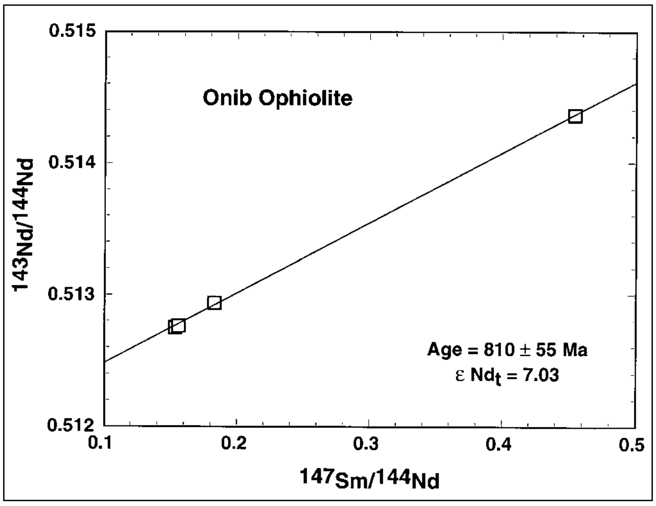

Fig. 2 - Sm/Nd isochron diagram of the Wadi Onib ophiolite.

Hamisanah

The 30 km wide and 300 km long Hamisanah shear zone is a prominent north-south trending tectonic element in the Red Sea Hills separating the Gebeit terrane from the Gabgaba terrane (Fig.1). Within the zone, arc-derived volcanics and sediments containing fragmented ophiolitic bodies have been strongly deformed by a vertically dipping foliation. The presumed ophiolitic remnants occur as slices or lensoid bodies within the shear zone over a length of 150 km. The ultramafic rocks are often completely serpentinized and/or carbonatized to form ophiocalcites. A complete ophiolite sequence could not be identified within the Hamisana shear zone. However, a possible dismembered ophiolite occurs approximately 20 km west of the Onib ophiolite. It contains chromite-bearing ultramafics, layered and isotropic gabbros and basic volcanics and is therefore taken as a presumed ophiolite.

Sm/Nd analysis of a basaltic sample from the Gabgaba terrane, AM-39, which was taken ca. 20 km west of the southernmost exposures of the Hamisanah zone, yields an eNd^t of +7.0 (Table 1), which indicates an intraoceanic nature of this part of the Gabgaba terrane, similar to the Gebeit terrane.

Jebel Tohado

The Tohado ophiolite complex (Figs. 1 and 3) forms a 400 m thick nappe remnant north of the Amur-Nakasib suture zone resting atop a series of greywackes, turbidites, and minor marbles. The thrust plane beneath Jebel Tohado that is mainly flat slightly steepens towards the E margins. The major part of the Jebel Tohado massif is composed of ultramafics that are strongly deformed and serpentinized or carbonatized. Gabbros and basalts are also present. The basalts are mainly massive without pillow structures. A small basaltic dyke complex is situated at the northern margin of the massif.

The Tohado ophiolite is interpreted to have been obducted onto the intraoceanic island arc crust of the Gebeit terrane (Reischmann and Kröner, 1994). The isotopic composition of a greywacke from the footwall of the nappe, approximately 2 km SW of Jebel Tohado, yielded an eNd^t of +5.0 and a Sr^i of 0.7033 (PS-96, Table 1 and 2, assuming an age of 887±20 Ma, Reischmann et al., 1992) indicating no significant continental influence. The Tohado complex is thus interpretated as a remnant of an ophiolite nappe in a flat tectonic position above an arc terrane.

Amur-Nakasib

The Amur-Nakasib suture forms an s-shaped SW-NE trending belt between the Gebeit terrane and the Haya terrane (Fig.1). It can be projected across the Red Sea as a continuation of the Bir Umq suture. Various parts of an ophiolitic assemblage are known from this belt. Serpentinized ultramafics occur frequently as slices within the steeply dipping, highly sheared rocks of the belt. Gabbros also form lenses in numerous places, but pillow basalts are only found in the Wadi Amur area and in the Arbaat-Nakasib area north of Port Sudan (Embleton et al., 1984; Hussein, 1999). Although highly dismembered, the mafic-ultramafic assemblage is considered as part of an ophiolite sequence. The neighbouring volcanics have the chemical compostion of modern arc tholeiites and the associated sediments, mainly conglomerates, greywackes and subordinate limestones, are also typical for an arc environment. Since the Gebeit terrane and the Haya terrane have been identified as island arc terranes (Reischmann et al., 1992; Reischmann and Kröner, 1994) the Amur-Nakasib belt is interpreted as an ophiolite containing inter-arc suture zone.

Wadi Ashat

During this study, a vertically dipping ENE-WSW trending shear zone was discovered in the Wadi Ashat area. This zone can be traced in a WSW direction for at least 100 km. Little is known about the Wadi Ashat shear zone because it is not easily accessible from the Red Sea coastal plain. This up to 5 km wide zone is parallel to the Amur-Nakasib suture and is also affected by the later Oko shear zone that caused the s-shaped form. The rocks within the Ashat shear zone are volcanics and granitoids of island arc affinity (Kröner et al., 1991). The age of this shear zone is only constrained by the age of the neighbouring terranes. Arc type granitoid rocks to the north have been dated at 871±10 Ma and 856±10 Ma (Pb/Pb single zircon ages, Kröner et al., 1991) and 854±9 Ma (U/Pb zircon age, Reischmann et al., 1992). The onset of post-tectonic igneous activity can be deduced from tholeiitic dykes of 660±19 Ma (K/Ar whole-rock age, Vail and Hughes, 1977), and from the 671±18Ma volcanics of the Homogar group (Rb/Sr whole-rock age, Klemenic, 1985).

Within the shear zone there are small slices of serpentinized pyroxenite with still preserved primary minerals. The Ashat shear zone is here tentatively considered to be an intrer-arc suture zone separating the northern from the southern Haya terrane. Hussein (1999) reports an ophiolite occurrence with ultramafics and layered gabbros in the Abu Samar region which is the continuation of the Ashat shear zone to the SW.

Baraka

The Baraka lineament is a dominant tectonic element in the southern Red Sea Hills, separating the Tokar and Haya terranes. It is a zone of intense, vertically dipping, N-S trending shearing that was interpreted as a suture zone (Vail, 1983; Kröner et al., 1987). However, complete ophiolites have not been found. Nevertheless, some ultramafic rocks have been reported by mapping teams of the Geological and Mineral Resources Department (GMRD) of the Sudan 150 km SSE of the Haya railway junction (I.M. Hussein, pers. commun., 1993). These ultramafic rocks are exposed in the region of uppermost Khor Baraka and may indicate the possible presence of ophiolites therein. The terranes on both sides of the Baraka lineament consist of arc-related igneous assemblages and minor immature sediments. Both terranes have the same age of ca. 850 Ma (Kröner et al., 1991). Although the origin of the Baraka lineament and the possible ophiolitic nature of the enclosed ultramafics remains unresolved it is here retained as the boundary between the Tokar and Haya terranes.

Nuba Mountains

The Nuba Mountains are located in the Kordofan province in the central Sudan. There is no continuous outcrop from the ANS to the Nuba Mountains, and interpretations derived from the Nuba Mountains relating to the ANS are speculative. Ophiolitic rocks are known from the vicinity of the town of El Biteira, about 250 km west of the River Nile (Hirdes and Brinkmann,1985; Brinkmann, 1986; Abdelsalam and Dawoud, 1991; Brinkmann et al., 1994). These ophiolites mainly consist of fine-grained serpentinite, serpentinized pyroxenites, talc schists and rare lenses of chromite (Hirdes and Brinkmann, 1985). The ultramafics are tectonically intercalated with layered and isotropic gabbros and greenschist to amphibolite facies metabasalts. However, I could not identify any dyke or pillow structures within these metabasalts. The highly deformed ophiolitic rocks are thrust over a metasedimentary sequence of schists, quartzites and limestones. The foliation in the ophiolitic rocks and the thrusts dip gently to the NW.

The ophiolitic character of the mafic-ultramafic assemblage could be confirmed by geochemical studies of Hirdes and Brinkmann (1985), Brinkmann et al. (1994), and Abdelsalam and Dawoud (1991), providing convincing evidence that they are part of ancient oceanic crust thrust onto the Nile craton and its platform sediments.

The samples analyzed in this study (Table 1 and 2) comprise a gabbro from the ophiolite, a chlorite schist and an arkosic sediment from below the ophiolite. The gabbro sample has an eNd^t of +7.2 indicating an ensimatic origin. The arkose and the chlorite schist on the other hand have lower eNd^t of approximately +3 which indicates a minor continental influence. The Sr^i is 0.7030 for the gabbro and 0.7029 for the chlorite schist, but slightly higher for the arkose with 0.7048. Together with the mid-Proterozoic Nd model ages of the basement (Harris et al.,1984) these data show a close association of ensimatic and ensialic chararcterisics which is typical for a continental margin environment.

Ingessana

The Ingessana ophiolite is situated in the Ingessana hills in the Blue Nile Province (Price, 1984). The ophiolitic rocks comprise serpentinized dunites and harzburgites that contain chromite and asbestos deposits, pyroxenites, layered and isotropic gabbros, and dolerite dykes that intrude the gabbros. This internally coherent sequence forms the lower part of an ophiolite sequence. It rests on greenschist facies metasediments separated by a basal tectonic contact. In the SW of the ultramafics there is a series of amphibolite facies gneisses. These gneisses are considered to be a deformed batholithic granite. Abdel Magid (1982, cited in Price, 1984) reported a Rb/Sr whole-rock age of 802 Ma (cited without error) for this gneiss complex which is the oldest known age from that area. The ophiolite and its country rocks were intruded at ca. 540 Ma by the post-tectonic Bau granite (K/Ar age, cited without error, Vail, 1976). Based on the geochemical data and the overall immature character of the clastic sediments, Price (1984) suggested that the ophiolite formed in an arc environment. To the north its possible extension may be the Qala al Nahal ophiolite, which is also known as a chromite deposit.

Akobo

The Akobo belt in western Ethiopia includes the ophio-

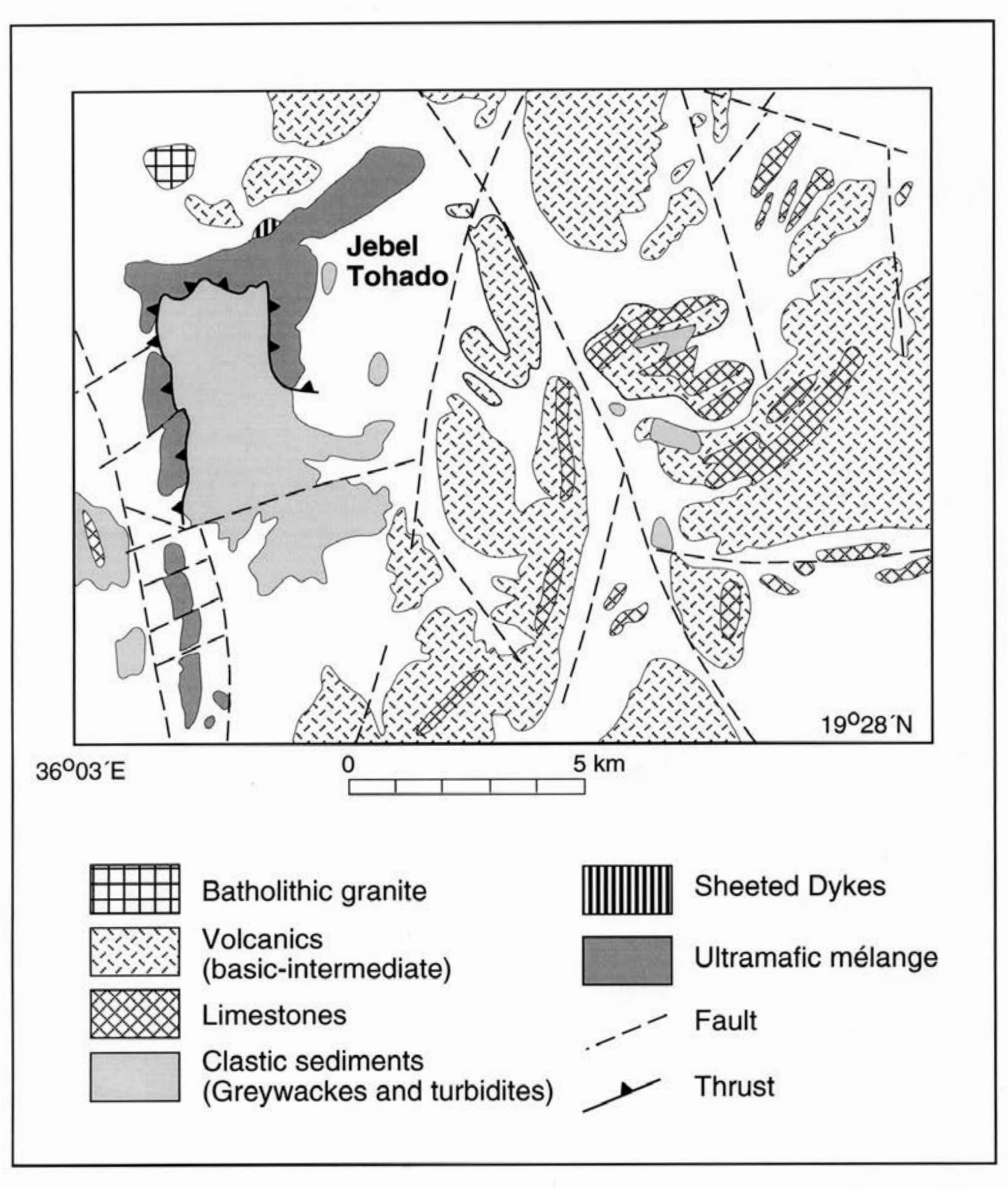

Fig. 3 - Geological map of the Jebel Tohado area, modified after Hussein (1999). The teeth at the trust line are inside the nappe.

lite exposed in the Akobo area and its northern extension in the Tulu Dimtu area. Further connections to the south are possibly exposed in NW Kenya. The ophiolites of the Akobo belt have been described by Kazmin (1976), Kazmin et al. (1979). They occur along a steeply dipping shear zone and are associated with a variety of metasediments and metavolcanics. The ophiolitic rocks comprise ultramafics, gabbros, and metabasalts. The ultramafics form several km long lensoid bodies and are mainly altered to serpentine and talc schists, but less altered dunites and peridotites are also preserved. Minor gabbros and amphibolites are associated with the ultramafics. Greenschist facies metabasalts were reported as well (Kazmin et al., 1979), but their relationship to the ophiolite sequence is not clear. Geochemical investigations indicate a mid-ocean ridge or island-arc character for the ophiolite (Warden et al., 1982).

Adola

The Adola belt comprises of the Adola and Kenticha ophiolites in southern Ethiopia. The Adola ophiolite forms lensoid bodies in steep tectonic contact with metasediments and gneisses. Ultramafics and gabbros are documented as well as pillow basalts (Kazmin, 1976; Kazmin et al., 1978), but an internally intact ophiolite sequence is not known from this highly dismembered complex. Possible extensions of the Adola belt and its contained ophiolites occur in the Moyale region in southern Ethiopia, the Baragoi area in Kenya, the Soka area in eastern Ethiopia and the Abdul Qadr in NW Somalia (Fig. 1). The Adola and Kenticha ophiolites are similar to inter-arc ophiolites. The Moyale ophiolite was assumed to be emplaced onto a continental margin (Berhe, 1990). However, single zircon dating revealed mainly late Proterozoic ages, and isotopic data show that this region is part of the Nubian shield (Teklay et al., 1998). Age determinations in eastern Ethiopia and NW Somalia (Kröner and Sassi, 1996) also yielded prevailing Pan-African ages. Inherited zircons of ages indicate the proximity of an old craton probably located somewhere to the east.

Sekker

This ophiolite comprises the mafic-ultramafic rocks of the Sekker area in NW Kenya (Vearncombe, 1983; Price, 1984; Berhe, 1990). Although this ophiolite is tectonically dismembered, most parts of an ophiolite sequence are preserved. Pyroxenites and serpentinites with some chromite lenses form the ultramafic part. Gabbros are layered as well as isotropic. Metabasalt occur as pillows and as dykes within the gabbros. The chemical composition is akin to a subduction related or enriched MORB-type ophiolite (Price, 1984). These rock types form a complex imbricate structure, tectonically emplaced on gneisses, quartzites, limestones, and metapelites. These belong to the margin of the Archaean Congo craton that is exposed in the west. The ophiolite occurs in a moderately dipping structural position above the Samor gneisses which are considered to be the foreland of the Congo craton. Nd isotope analyses of the Marich granite by Harris et al. (1984) also indicate the older cratonic character of the crust onto which the ophiolite was obducted.

DISCUSSION AND CONCLUSIONS

In this section the major features of the ophiolites and suture zones which were described above are summarized (Table 3) and a general classification of the ophiolites based on Moores (1982) is presented (Fig. 4). Finally, the discussion will focus on the implication of the ophiolites for the evolution of distinct arc terranes in particular and of the ANS in general.

It has to be noted that rocks belonging to metamorphic complexes (Moores, 1982) within the ophiolite sequences of the ANS are difficult to classify because of their general greenschist to amphibolite facies regional metamorphism that can even reach granulite facies in the realm of the Mozambique belt. Further, it should be pointed out that there is a bias of the data base and, consequently, of the interpretation towards the northern Nubian Shield which is much better studied than the region to the south.

Classification of the Nubian Ophiolites

Eastern Desert

The igneous rocks in the Eastern Desert of Egypt have low Sr^i. Using the data of Stern and Hedge (1985) the Sr^i range from 0.7019 to 0.7023 for ages >700 Ma, from 0.7025 to 0.7033 for 600 - 700 Ma and from 0.7028 to 0.7042 for the post-tectonic rocks that are < 600 Ma. These data indicate a slight increase of the Sr^i with time but nevertheless are indicative of an ensimatic origin for these rocks. In contrast to this, the sediments analysed in this study were derived from continental material which is evident from high Sr^i (Table 2) and distinctly negative eNd^t (Table 1). The nature of the basement onto which the ophiolite nappes of the Eastern Desert were thrust is therefore not well constrained (Kröner et al., 1987; 1988). The new isotopic data for the sediments do not necessarily indicate a continuous pre-Pan-African basement but provide further evidence for the proximity of continental crust. Two scenarios appear to be plausible: either the region was an active continental margin with highly irregular shape, embayments and promontories or it incorporated continental fragments. A decision on this subject cannot be made on the basis of the available data. But whatever the size and form of the former continent was, the isotope analyses of the sediments provide evidence that the basement of the ophiolite nappes contains pre-Pan-

Table 3 - List of Nubian Ophiolites

Ophiolite	Location	Structural	Geochemical	Mineralisation	Age (Ma)	Age References
		Class.	Class.		(Formation)	
Wadi Ghadir	Eastern Desert	Tethyan	BAB	Cr	746±19	Kröner et al. (1992)
Gebel Gerf	Red Sea Hills	Tethyan	MOR/IA	Cr	741±21	Kröner et al. (1992)
Abu Swayel	Eastern Desert	Tethyan	IA	S	736±11	Kröner et al. (1992)
Wadi Allaqi	Eastern Desert	?	IA	Cr	770±9	Kröner et al. (1992)
Sol Hamed	Red Sea Hills	Cordilleran	IA	Cr	712±58	Fitches et al. (1983)
Onib	Red Sea Hills	Cordilleran	BAB	Cr	808±14	Kröner et al. (1992)
Hamisanah	Red Sea Hills	Cordilleran	IA	Cr, S	887±20	Reischmann et al. (1992)
Tohado	Red Sea Hills	Cordilleran	IA	-	887±20	Reischmann et al. (1992)
Amur-Nakasib	Red Sea Hills	Cordilleran	IA	S	887±20	Reischmann et al. (1992)
Ashat	Red Sea Hills	Cordilleran	IA	Cr	871±10	Kröner et al. (1991)
Nuba Mts.	Kordofan	Tethyan	IA	Cr, Mg, S	ca. 750	assumed
Ingessana	Blue Nile	Cordilleran	IA	Cr	>802	Abdel Magid (1982)
Akobo	W Ethiopia	Cordilleran	IA/MOR	-	ca. 750	assumed
Adola	S Ethiopia	Cordilleran	IA	-	ca. 750	assumed
Sekker	NW Kenya	Tethyan	IA	Cr	ca. 750	assumed

Legend: IA =Island arc, MOR= Mid ocean ridge, BAB= Back arc basin, Mineralisations: Cr=Chromite, S= Sulfides (mainly Fe; Cu, Zn in traces), Mg=Magnesite

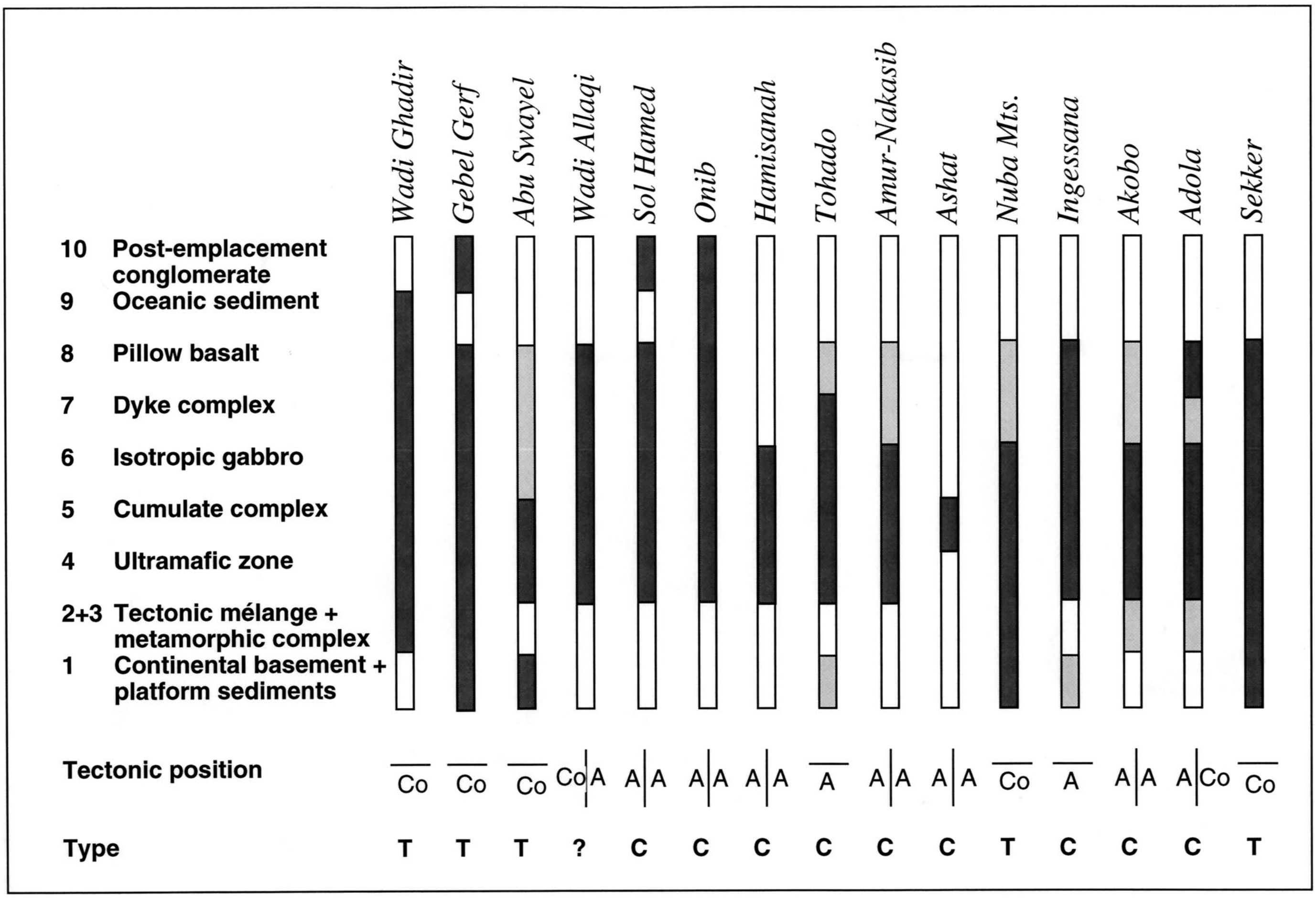

Fig. 4 - Compilation of the Nubian ophiolites based on the scheme of Moores (1982). Dark columns: proven, shaded columns: inferred, white columns: unknown. Tectonic position is mainly flat (horizontal line) or steep (vertical line) between arc terranes (A) or continental crust (Co). T indicates Tethyan-type, C indicates Cordilleran-type.

African continental material. According to the scheme of Moores (1982) the Eastern Desert ophiolites are therefore grouped as **Tethyan-type**. These are characterized by a horizontal structural position possibly over continental crust of the Nile craton.

Red Sea Hills

The ophiolites of the RSH occur in suture zones. The Allaqi belt is bound by an arc terrane to the south and probably a crustal terrane to the north. All other sutures separate juvenile arc terranes. These intra-arc ophiolites are characterized by a steep structural orientation with the exception of the Jebel Tohado, which is an ophiolite nappe with flat tectonic contacts to the underlying arc terrane. According to the scheme of Moores (1982) all intra-arc ophiolites of the Red Sea Hills are classified as **Cordilleran-type,** whereas the classification of the Allaqi belt remains uncertain.

Central Sudan

The Ingessana ophiolite lies on island arc crust and is therefore tentatively classified as **Cordilleran-type**. The ensimatic origin of the neighbouring gneisses is also supported by Nd isotopic data on the granulites of the Jebel Moya region to the NW which belong to the ensimatic area (Stern and Dawoud, 1991). The Bayuda desert in the north also belongs to the ensimatic domain of the Red Sea Hills because the ages from that region are late Proterozoic and the Sr^i are low (Ries et al., 1985). The Nuba Mountains ophiolite west of the Nile provide a good example of a **Tethyan-type** ophiolite, thrust in a flat structural position onto the Nile craton.

Ethiopia

The ophiolite occurrences in western Ethiopia are classified as **Cordilleran-type**. The terranes bordering the ophiolites to the west display arc characteristics. The other terranes to the east might be considered as pre-Pan-African because they consist of gneisses and migmatites and may therefore be part of an unknown continent to the east which is indicated by inherited pre-Pan-African zircons (Kröner and Sassi, 1996). The high degree of metamorphism does not necessarily prove an older age but might simply reflect the influence of the Mozambique belt metamorphism which has been documented from the central Sudan (Kröner et al., 1989; Stern and Dawoud, 1991). Since the pre-Pan-African basement is not clearly constrained, these ophiolites cannot unequivocally be classified. Tentatively the Akobo ophiolite is classified as **Cordilleran-type** in a steep structural position between arc terranes. The ophiolites in southern and eastern Ethiopia are possibly associated with continental crust but a classification is questionable.

Kenya

Ophiolites from the Sekker region in western Kenya are classified as **Tethyan-type**. The structural position is flat although it steepens towards the east in the highly sheared area affected by the Mozambique belt deformation. The Sekker ophiolite was obducted onto the foreland of the Archaean Congo craton to the west. The continental character of the basement is constrained by the isotopic data. The other ophiolites from the Baragoi area are related to the Sekker ophiolite and also match this classification.

Summary of the ophiolite classification

A two-fold classification of the ophiolites which can be applied to most of the ophiolites of the Nubian shield is proposed:

a) Ophiolites in steep structural position between arc terranes are mainly classified as Cordilleran-type.

b) Ophiolites in flat structural position resting as nappes on a continental margin or microcontinental basement are mainly classified as Tethyan-type.

Exceptions to this simplifying scheme are the Jebel Tohado and the Allaqi and Ingessana ophiolite.

The reasons for the different structural positions of this two-fold classification may be explained by the physical state of the neighbouring terranes. The young crust of the arc terranes will be more easily deformed than the old rigid crust of the Nile craton. Post-obduction deformation in arc terranes will preferentially reactivate old intra-arc sutures. Ophiolites in these zones will be tectonized and stretched out, forming elongate bodies. Old consolidated, cratonic crust will behave as a rigid block, and ophiolites resting as nappe complexes on such type of crust will not be affected by later deformation. This hypothesis provides a simple model that can easily be tested.

Life time of the magmatic arcs

Understanding of the origin of the ophiolites is impossible without knowledge about the neighbouring arc terranes. On the other hand the ophiolites themselves can provide important time markers for the arc evolution. Fig. 5 is a compilation of the evolutionary stages of arc terranes in the northern Nubian shield. These were selected because the age relation of the igneous phases are reasonably well established, whereas in the southern part of the Nubian shield age information is too scarce, and ophiolite formation has not been dated. The terranes selected are the Gerf terrane and the Eastern Desert, where T-type ophiolites are resting on a basement of presumed continental character. The others are from the Red Sea Hills, where C-type ophiolites occur along suture zones separating arc terranes.

In the terranes, where the beginning of the igneous activity is not well known, ophiolites provide a minimum age for the arc formation. Except the basalts from the Gerf Complex, all ophiolites - based on geochemical studies - are arc related, and are either back arc type or of supra subduction zone type. The formation of such ophiolites requires a simultaneously active arc and is therefore taken to constrain the minimum age of arc formation.

The oldest ages of the **Tokar** terrane are documented by

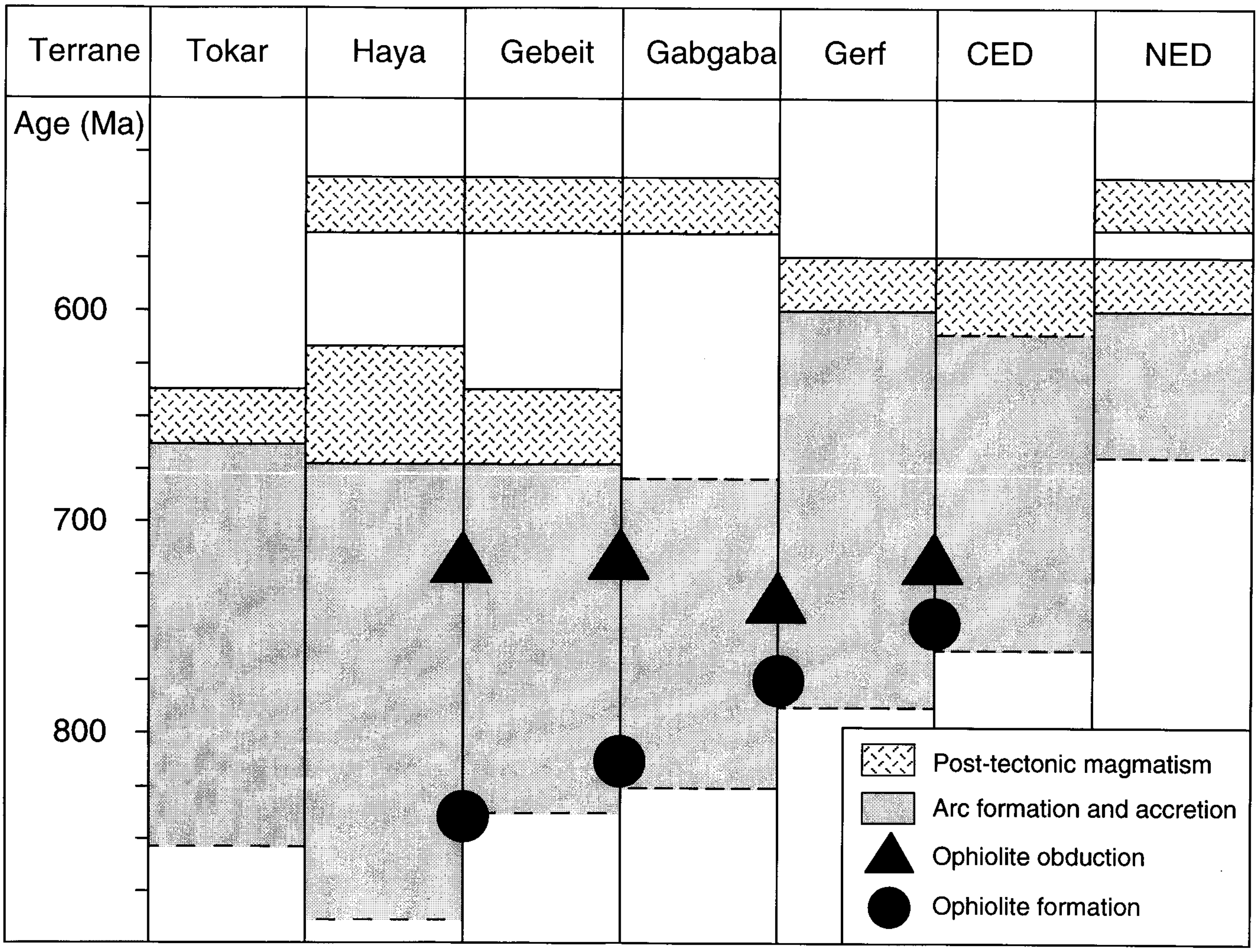

Fig. 5 - Schematic illustration of the arc evolution and ophiolite obduction for the northern Nubian shield. Broken lines indicate minimum or maximum values. CED: Central Eastern Desert, NED: Northern Eastern Desert. Detailed ages and errors are given in the text.

the 854±18 Ma and 840±16 Ma old rhyolites and the 827±33 Ma old gneisses (Kröner et al., 1991). A post-tectonic granite intrusion at 652±14 Ma marks the end of the arc evolution (Kröner et al., 1991).

The evolution of the **Haya** terrane started with the 887±20 Ma old rhyolites. Granites from this terrane were dated at 871±10 Ma and 856±10 Ma (Kröner et al., 1991), 855 Ma (cited without error, Vail, 1976), and 854±9 Ma (Reischmann et al., 1992). Post-tectonic volcanics erupted at 671±18 Ma (Klemenic, 1985) and undeformed dikes intruded at 660±19 Ma and 618±18 Ma (Vail and Hughes, 1977). A later stage of magmatism occurred at 539±12 Ma (Vail, 1976).

The oldest rocks of the **Gebeit** terrane known so far are the 832±26 Ma old arc volcanics (Reischmann and Kröner, 1994). The arc evolution comprises 723±6 Ma old magmatic complexes (Klemenic 1985) and orogenic granites with 686±18 Ma (Cavanagh, 1974). Post-tectonic younger granites intruded at 671±16 Ma (Cavanagh, 1974) as well as 646±10 Ma (Kröner et al., 1992). Volcanics of this magmatic phase were dated at 649±18 Ma (Cavanagh, 1974). The latest stage of the pan-African evolution took place at 552-542 Ma (cited without error, Vail, 1976).

The **Gabgaba** terrane is of similar age as the neighbouring Gebeit terrane, which can be derieved from the 808±14 Ma age of the Onib ophiolite (Kröner et al., 1992). Arc type rocks have an age of 769±20 Ma (Mankl et al., 1987). The end of the arc evolution is not directly dated but correlations of post-tectonic rocks with the Gebeit terrane indicate an end of the arc evolution at ca. 675 Ma and a late magmatic stage at ca. 550 Ma.

The **Gerf** terrane has a minimum age of 770±9 Ma (Kröner et al.,1992). Arc related magmas formed at 741±5 Ma, 691±5 Ma and 663±29 Ma (Stern et al., 1989). The end of the arc evolution is given by the 594±4 Ma old Assuan granite (Stern and Hedge, 1985).

The **Central Eastern Desert** must be at least as old as the 746±19 Ma old Ghadir ophiolite (Kröner et al., 1992). The post-tectonic igneous assemblage of the Dokhan volcanics is dated with 607±8 Ma (Stern and Hedge, 1985). This period is also known from a granite of 578±15 Ma (Sultan et al., 1990).

The **Northern Eastern Desert** is the youngest terrane with a 666 Ma old granite as the oldest known rock unit (Stern and Hedge, 1985). Post-tectonic phases are well dated at 600-575 Ma and 560-540 Ma (Stern and Hedge, 1985).

A first interpretation of this compilation concerns the life time, i.e. the time span of igneous activity of discrete arc segments. The longest activity of ca. 200 Ma is inferred for the Haya and Tokar terrane. This must be perceived as a maximum timespan, because the possibility of a hiatus has to be considered. The Gebeit terrane was active for about 150 Ma, and this time span of activity is well constrained. For the northern terranes in the Eastern Desert there is only limited information on the time of the igneous activity. Therefore, a 100 Ma life time must be considered as a minimum value. On average, a time span of 150 Ma for the late Proterozoic Nubian arcs seems to be realistic.

Timing of ophiolite formation and obduction

As already shown by Pallister et al. (1988) and Kröner et al. (1992) the ophiolites of the ANS belong to the oldest rocks of the arc assemblages. This is obvious in the Gebeit terrane where the oldest arc volcanics are 832±26 Ma (Reischmann and Kröner, 1994) and the oldest ophiolite at Wadi Onib is 808±14 Ma (Kröner et al., 1992). In the Tokar and Haya terrane ophiolites have not yet been dated, but the age can be deduced from the assumption that the Amur-Nakasib suture correlates with the Bir Umq suture of the Arabian shield. Ophiolites from there formed between 840 and 870 Ma ago (Pallister et al., 1988) and the same age range can be adopted for the Sudanese part of the suture. The ophiolites farther north towards the Eastern Desert are 770±9 Ma, 741±21 Ma and 746±19 Ma (Kröner et al., 1992). In general, it appears that the ophiolites in the northern Nubian shield are becoming younger farther to the north, and so are the arc volcanics.

The time when ophiolites were obducted provides a further age constraint indicating the closure of the related ocean and thus cessation of the major arc activity. In the arc terranes of the Red Sea Hills, the obduction age is >712±58 Ma for the Sol Hamed ophiolite (Rb/Sr whole-rock age, Fitches et al., 1983) and >736±11 Ma for the Allaqi belt (Pb/Pb single zircon age, Kröner et al., 1992). For the Bir Umq-Nakasib suture Stoeser and Camp (1985) suggested 715 Ma as the closure age, and the Hamisanah suture is also considered to have closed before 715 Ma (Stern et al., 1989). During this time a major accretion event took place in the central ANS, where the arc terranes were juxtaposed to each other and the C-type ophiolites as fragments of the ocean basins were trapped in sutures. This event is not visible in the T-Type ophiolites from the northern part of the shield. However, this accretion period is also known from two granulite terrains in the central Sudan, namely the 721±12 Ma old granulites from Sabaloka (Kröner et al., 1987) and the 739±2 Ma and 742±2 Ma old granulites from Jebel Moya (Stern and Dawoud, 1991). The high grade metamorphism of these areas, which belong to the northernmost extension of the Mozambique belt, is assumed to be caused by crustal thickening through collision. This indicates that, at about 700 Ma, all oceans between the Red Sea Hills arc terranes and the Nile craton had closed and the arcs had already been accreted to the ancient continent.

Evolutionary stages of the Northern Nubian shield

The evolution of the northern Nubian shield started in the arc terranes of the Red Sea Hills between 887±20 Ma and 832±26 Ma. The oldest ophiolites, closely associated in time and space with these arc type rocks, had been classified as C-type. In the Eastern Desert the evolution of a continental magmatic arc started about 100 Ma later. Similarly, the ophiolites of the Eastern Desert, which are classified as T-type, become younger towards the north. Obviously, the simple classification scheme of Moores (1982) reflects the differences in age and structural position of the ophiolites of the Nubian shield. Formation and obduction of ophiolites was restricted to the early evolutionary stages of the shield. These stages were related to the subduction of oceans of unknown dimensions (Reischmann et al., 1992).

Later on, closure of the oceans was followed by mainly felsic, post-tectonic igneous activity between 671±18 Ma (Klemenic, 1985) and 618±18 Ma (Vail and Hughes, 1977) in the Red Sea Hills arc terranes. This can be interpreted as a consequence of crustal thickening followed by partial melting in the thermally equilibrated middle or lower crust. The time gap between collision and the onset of this igneous phase is in agreement with the model of Patiño-Douce et al. (1992) who suggested a time span of about 50 Ma between the maximum of crustal thickening and crustal melting.

Another post-tectonic igneous stage is known from the Northern Eastern Desert and Sinai at 600-575 Ma (Stern and Hedge, 1985; Stern and Voegeli, 1987; Sultan et al., 1990; Beyth and Reischmann, 1997). Finally, important post-tectonic magmatism occurred at 560-540 Ma (Stern and Hedge, 1985). This mainly alkaline magmatism is known from the Northern Eastern Desert as well as from the Red Sea Hills arc terranes in the south and marks the last major igneous pulse in the Nubian shield at the end of the Pan-African evolution.

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APPENDIX

Analytical methods

Some of the Rb/Sr ratios were analyzed on a XRF following Pankhurst and O´Nions (1973). Sm, Nd, all remaining Rb and Sr concentrations, and the isotopic composition of Nd and Sr were analyzed on a Finnigan MAT261 mass spectrometer. The method is described in White and Patchett (1984). Isochron regression followed York (1969). Errors of the ages are quoted on a two-sigma level. Standard isotope values achieved during the course of this study are: ^{143}Nd/^{144}Nd=0.511837±18 for La Jolla normalized to ^{144}Nd/^{146}Nd=0.7219 and ^{87}Sr/^{88}Sr=0.710249±16 for NBS 987 normalized to ^{86}Sr/^{88}Sr=0.1194. eNd^t was calculated using ^{143}Nd/^{144}Nd=0.512638 and ^{147}Sm/^{144}Nd=0.1967 as chondritic values. Depleted mantle model ages assume a linear evolution from eNd^t=0 at 4.5 Ga to eNd^t=10 at present.