Mesoproterozoic rocks of Namibia and their plate tectonic setting

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Abstract

Two main Mesoproterozoic provinces occur in southern Namibia: (1) The high-grade Namaqua Metamorphic Complex (NMC) composed of a supracrustal sedimentary succession and interpreted as a passive margin sequence in the west of the Kalahari craton; (2) The Sinclair Group and its northeastern correlatives, composed of two main magmatic and metamorphic units, reflecting northeast-directed subduction, which started before 1.37 Ga and lasted until about 1.1 Ga. These two units were tectonically juxtaposed during the 1.1–1.03 Ga Namaqua orogenic event.

The Kairab–Kumbis Metamorphic Complex comprises metasedimentary and metavolcanic rocks intruded by the 1.37 Ga arc-related Aunis tonalite. The mafic volcanic rocks from this complex have geochemical features of island arc calcalkaline basalts; they were emplaced and metamorphosed along an active margin before 1.37 Ga. The 1.2–1.1 Ga low-grade unmetamorphosed volcanic and immature sedimentary rocks of the Sinclair Group and its northwestern equivalents rest disconformably on the Kairab–Kumbis Complex. They occur in fault-bounded depocenters defining a regional arc-shape structure up to 100 km-wide and with a minimum length of 2000 km. The plate tectonic setting of this arc is best constrained by the composition of volcanic rocks from the ~1.2 Ga Barby Formation and coeval granitoids; they comprise high-K calcalkaline rocks suggesting emplacement in an active continental margin setting. The final stage of this continental arc evolution is recorded in the <1.1 Ga tholeiites of the Opdam Formation. High Ti-content and flat REE-patterns in the tholeiites suggests an extensional event, whereas high Th/Ta and La/Nb ratios, low Ce/Pb values and negative anomalies for Nb–Ta suggest a subduction-related setting for the mantle source from which the mafic magmas were derived. Docking of continents led to the slab detachment, allowing interaction between the asthenospheric mantle and the mantle wedge enriched during the subduction process. The magmatic underplating related to this event induced the genesis of large-scale batholitic granitoid bodies in the NMC and a ~1.1–1.0 Ga high-grade LP/HT metamorphism, with mineral assemblages indicating an anti-clockwise P–T–t path.

Keywords: Namaqua Belt; Sinclair Group; Subduction; Docking; Lithosphere delamination; Mesoproterozoic; Namibia

1. Introduction

The Mesoproterozoic was a time of major crust generation within a global accretionary event commonly linked to the assembly of the Rodinia Supercontinent (Hoffman, 1991; Moores, 1991; Weil et al., 1996). This accretionary event is documented in a worldwide net of orogenic belts (e.g. the Grenvillian belt in northern America and the Kibaran orogenic system in Africa), leading to the final assembly of Rodinia at ≈1 Ga. In Namibia, two main Mesoproterozoic provinces are distinguished (Fig. 1): (1) the Namaqua Metamorphic Complex (NMC) located southwest of the Hauchab-Excelsior-Lord Hill Shear Zone, and (2) the Sinclair Group basin to the northeast of this structure. Additional Mesoproterozoic complexes identified in Namibia include basement "inliers" within the Neoproterozoic Damara/Kaoko belts (Kröner et al., 1991, 2001; Seth et al., 1998; Steven et al., 2000), and the

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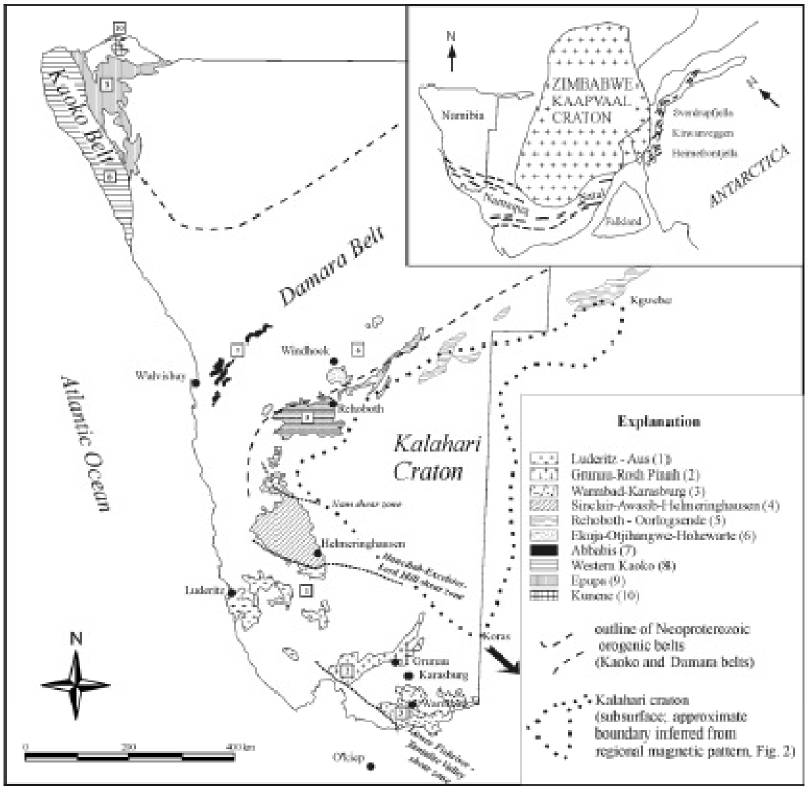


Fig. 1. Distribution of geological units affected by Mesoproterozoic events in Namibia. Inset shows the reconstruction of the relative positions of southern Africa and East Antarctica in Gondwana and location of the Namaqua-Natal belt (Jacobs et al., 1993).

"Kunene Anorthosite Complex", which straddles the border between Angola and Namibia and represents one of the largest AMCG-type (Anorthosite–Mangerite–Charnockite–Granite) intrusions in the world (Silva, 1990; Ashwal and Twist, 1994).

The medium- to high grade NMC of southern Namibia forms the northern extension of a metamorphic belt that can be traced from Namibia through South Africa and into Eastern Antarctica (Fig. 1 inset, Fig. 2; Hartnady et al., 1985; Martin and Hartnady, 1986; Jacobs et al., 1993). It is composed of various ortho- and paragneisses, whose geographical distribution, age and tectonic setting are inadequately documented in Namibia. The link between the NMC and the rest of the Kibaran orogenic system in Africa has never been properly established.

The low-grade to unmetamorphosed Sinclair Group supracrustal assemblage (Fig. 1) is exposed to the northeast of the NMC. It is part of a succession of volcanic and sedimentary basins aligned along the western and northwestern margins of the enigmatic Rehoboth Subprovince in the west of the Kalahari Craton, geophysically distinct from the Zimbabwe–Kaapvaal Craton, though hidden underneath younger sediments (Figs. 1 and 2; Clifford, 1970; Borg, 1988; Kampunzu et al., 1998) and includes the Kgwebe Formation of Botswana, the Mesoproterozoic Sinclair–Nauzerus Groups of Namibia and the Koras Group of South Africa. Bimodal volcanic sequences and minor epicontinental sedimentary clastic rocks, which form the dominant basin fill, are mildly deformed. Batholithic granite bodies and subordinate gabbro and granodiorite were emplaced during the development of the Sinclair basin in Namibia (Watters, 1976; Borg, 1988; Schalk, 1988; Hoal, 1990; Ziegler and Stoessel, 1993). The tectonic setting during the development of the latter has always been a matter of controversy (Von Brunn, 1967; Watters, 1974, 1976; Kröner, 1977; Mason, 1981; Brown and Wilson, 1986; Borg, 1988; Hoal, 1990, 1993; Thomas et al., 1994a,b; Hoal and Heaman, 1995; Kampunzu et al., 1998), while the tectonic and stratigraphic relation between the NMC and the Sinclair

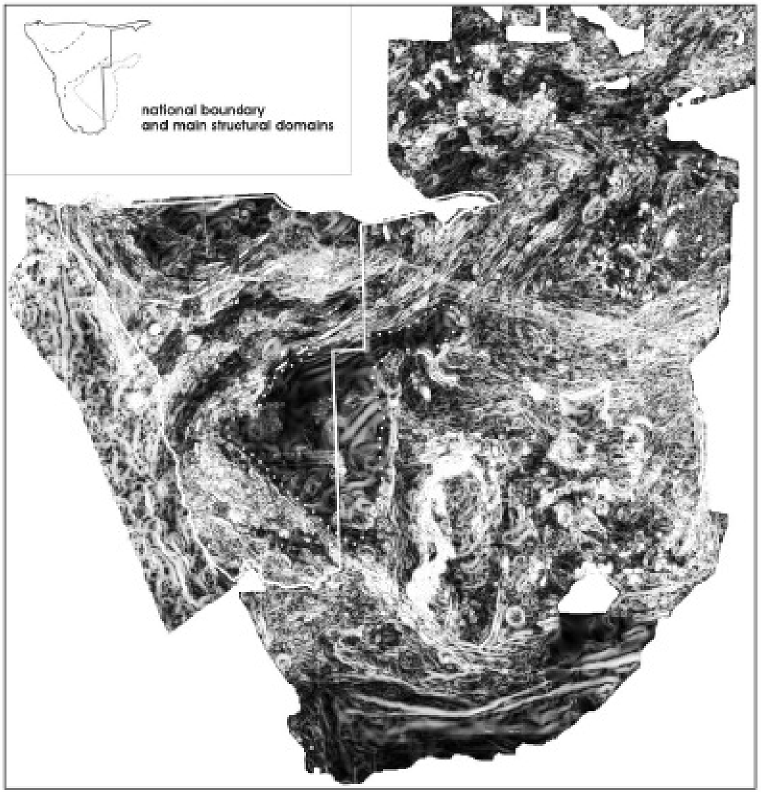


Fig. 2. Regional map of southern Africa; method: total horizontal derivative of total magnetic field upward continued (1000 m intervals; Wackerle, unpublished).

province has not attracted much attention (Watters, 1974; Blignault et al., 1983; Hoal, 1990, 1993; Thomas et al., 1994a,b; Colliston and Schoch, 2000). Equally the internal stratigraphy and timing of events within each of the two provinces has not yet been adequately constrained; this is especially true for the Rehoboth part of the Sinclair province, where Mesoproterozoic rock assemblages are partly covered by Neoproterozoic sedimentary sequences and overprinted by Neoproterozoic tectono-thermal events of the Damara orogeny in Namibia (Miller, 1983; Gareeb et al., 2002).

The aim of this paper is to give a review of the Mesoproterozoic rocks of Namibia by characterising their individual stratigraphic, geochronological, metamorphic and geochemical properties. Structural relationships between the different areas are discussed where appropriate and some attempt is made to place the current knowledge of these rocks in a plate tectonic frame.

2. Namaqua Metamorphic Complex (NMC)

In Namibia, the NMC covers ≈60,000 km^2, defining a ≈100 km-wide northwesterly trending strip from Luderitz on the Namibian coast southeast to the Orange River (Fig. 3). Within this strip, the NMC is exposed in three main areas (Luderitz-Aus; Grunau-Rosh Pinah and Warmbad-Karasburg) separated by Neoproterozoic/Phanerozoic cover. The Mesoproterozoic tectonostratigraphic units exposed in these areas form the northern Grunau terrane (Table 1; Colliston and Schoch, 1998). The Lower Fish River – Tantalite Valley shear zone (Fig. 3), a southeast-trending crustal-scale lineament, marks the contact along which the high-metamorphic grade Grunau terrane was thrust towards the southwest onto the lower-metamorphic grade Paleoproterozoic (2.0–1.8 Ga) Pofadder terrane (Colliston and Schoch, 1998), or Richtersveld Complex (Blignault et al., 1983). The latter comprises sedimentary

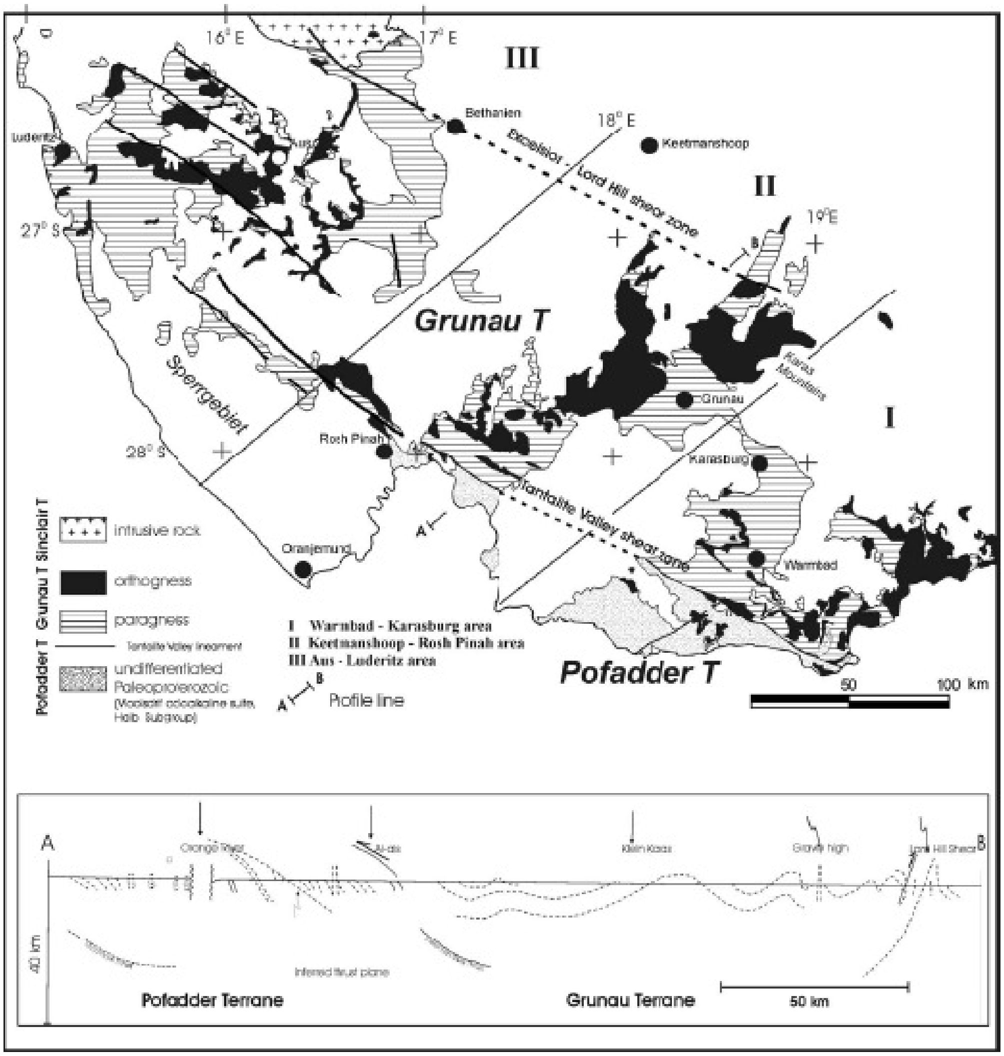
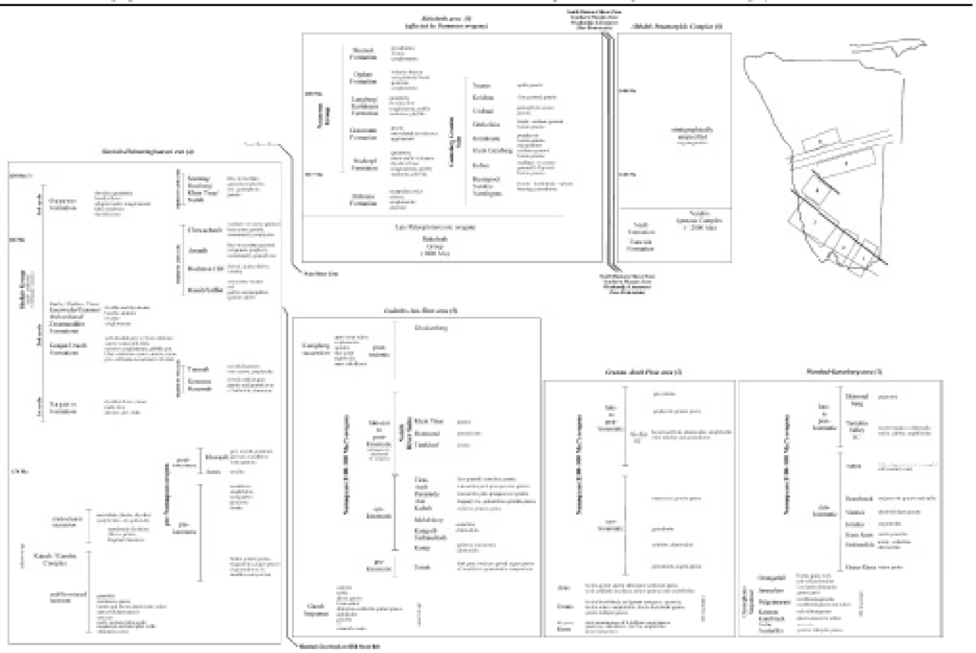


Fig 3. Simplified geological map of the Namaqua Metamorphic Complex (NMC) and distribution of supracrustal sedimentary and igneous rocks and major shear zones in Namibia. Inset shows a regional cross-section through the Namaqua Metamorphic Complex (NMC) documenting the tectonic style of this terrane in Namibia (Blignault et al., 1983).

and volcanic rocks of the Haib Subgroup intruded by tonalites to granodiorites of the Vioolsdrif Igneous Suite (Blignault et al., 1983; Reid and Barton, 1983). This tectonic boundary is less well defined to the east of the NMC, where Mesoproterozoic orogeny (locally known as the Namaquan) resulted in interfolding of units of the Grunau and Pofadder terranes. The Lower Fish River – Tantalite Valley shear zone is part of a regional thrust and fault system that includes the Hauchab-Excelsior-Lord Hill shear and the Nam shear zone to the north (Fig. 1; Vajner, 1974; Blignault et al., 1974; Hoal, 1990). Exhumation of granulites southwest of the Hauchab-Excelsior-Lord Hill shear zone indicates a major crustal discontinuity. This becomes less obvious towards the northwest, due to a

Table 1
Overview of the lithostratigraphic units of the NMC, SHA, Rehoboth terrane and the AMC and their relationship to the Namaquan and Damaran orogeny



Inset map shows schematically the spatial distribution of the areas and crustal scale boundaries inbetween; the boundaries are also indicated in the table as lines separating the individual areas.

general northwesterly decrease in metamorphic grade, which led to the juxtaposition of amphibolite facies rocks on both sides of the shear zone in the Awasib area (Fig. 4). The Grunau terrane is juxtaposed with the Sinclair–Helmeringhausen–Awasib (SHA) terrane along the Hauchab-Excelsior-Lord Hill shear zone, with the terrane boundary regarded as the front of the Namaqua orogen (Blignault et al., 1983).

The Namibian part of the NMC has been covered by regional mapping and preliminary structural and petrographic investigations (Greenman, 1966; Blignault, 1971, 1974a,b, 1975, 1977; Beukes, 1973; Moore, 1973; Blignault et al., 1974; Blignault, 1975; Jackson, 1976; McDaid, 1976, 1978; Toogood, 1976; Kartun, 1979; Geringer and Beukes, 1973; Schreuder and Genis, 1977; Beukes and Botha, 1978a,b,c,d,e,f). Few geochronological and geochemical analyses have been reported (Moore, 1975; Kartun, 1979; Moore, 1981; Barton, 1983; Burger and Coertze, 1975, 1978; Kartun and Moore, 1979; Reid and Barton, 1983). Available geochronological data (Table 2) suggest that the Namibian NMC includes Mesoproterozoic (1250–1100 Ma) rocks and a strongly reworked Paleoproterozoic (2000–1700 Ma) basement. The proportion and the areal distribution of these two components remain largely unknown. In the sections that follow, the terms pre-, syn- and post-kinematic refer to the main structural grain of the NMC, which is northwest–southeast trending.

2.1. Warmbad-Karasburg area

The rocks of the NMC exposed in this region define a $\approx$8000 km^2 diamond-shaped area largely covered by younger sediments (Fig. 3). The high-grade polycyclic metamorphic rocks of the Grunau terrane strike southeasterly in this region adjacent to South Africa. They overthrust low-grade Paleoproterozoic basement of the Pofadder terrane to the southwest along the Lower Fish River – Tantalite Valley shear belt. Towards the east, Namaqua deformation and metamorphism increase in both terranes, and complex interference folding is common on a regional scale. Primary interlayering between Paleo- and Mesoproterozoic rocks of these terranes has been proposed by Toogood (1976) and Blignault et al. (1983). However, further structural and geochronological work is required, and

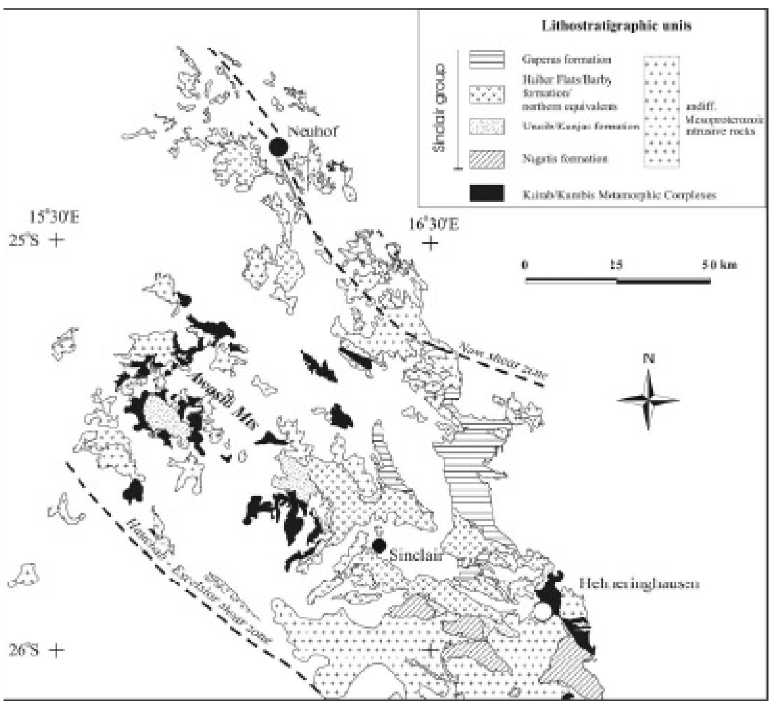


Fig. 4. Simplified geological map of the Sinclair–Awasib–Helmeringhausen (SAH) area (Namibia) showing the Mesoproterozoic rocks. Intrusive rocks are undifferentiated.

Table 2
Geochronological database for the NMC, SHA and Rehoboth areas, U–Pb zircon ages only

	Stratigraphic unit	Locality	ID	Lithology	Age	+	–	Method	Reference
Namaqua Metamorphic Complex									
Luderitz Aus	Glockenberg	Luderitz area	LC57	Granite	1022	nd		MG	Burger (1975)
			LC57+	Granite	1086				
			LC56+	Granite	1178				
			LC56	Granite	1194				
	Jamelspan F	Aus	PRU122	Biotite schist	1116				
	Kunguib	Luderitz area	Kunguib	Granodiorite	1952				
				Granodiorite	1978				
	Namaqualand	Hottentots Bay	KL6	Augengneiss	1080				Kröner (1976)
			KL5	Granite	1114				
	Tierkloof	nd	PRU115	Diorite	1080				Jackson (1976)
	Tschaukaib	Luderitz area	TK1+	Granite	1012				Burger (1975)
			TK1	Granite	1018				
	Tsirub	Aus	PRU114	Tonalitic augengneiss	1052				Kröner (1976)
	nd	Grabwasser261	B658	Granite	822				Burger (1974)
		Grasplatz	KL3	q-Monzonite gneiss	1082				Kröner (1976)
		Luderitz area	KL1	hbl-Gneiss	1755				
			KB1	Felsic volcanics	1131				Burger (1975)
			LC80+	Felsic volcanics	1132				
			LC80	Felsic volcanics	1184				
			LC33	Granite	1233				
			LC33+	Granite	1264				
			LC19	Augen gneiss	1742				
			LC19+	Augen gneiss	1752				
Warmbad-Karasburg		Altdoring 3	BB1	Paragneiss	1020				Burger (1974)
		Warmbad 145	GBW1	Granite	874	20	20		Burger and
		Eendorn 106	GBW2	Granite	1258	25	25		Coertze (1975)
	Beenbreek	Beenbreek	PRU105	Granite	1218	nd			Toogood (1976)
	Naros	Ondemaatje	PRU106	Granite	1042				
	Nautsis	Keimas 99	PRU112	Granite	952				
	Orangefall	Velloorsdrift 93	PRU103	Biotite gneiss	1142				
			PRU104	Biotite gneiss	1720				
	Stolzenfels	Stolzenfels 74	PRU117	Norite	1110				
	nd	Witputs 258	PRU150	Migmatitic gneiss	1861	26	26		Kröner et al. (1983)
	nd	Langklip	BJ1011	Charnockite	1100	14	14		Walraven et al. (1982)
Sinclair–Helmeringhausen–Awasib area									
	Aunis	Awasib	BH942	Tonalitic gneiss	1376.5	1.7	1.6	SG	Hoal and Heaman
	Awasib	Mountains	BH943	Granite	1216	1.8	1.5		(1995)
	Chowachasib		BH760	Granite	1216.7	4.7	3.5		
	Guperas F	Dabis 15	No 54	Granophyr	940	40	40	MG	Burger and Coertze (1973)
	Haisib	Awasib Mnts	BH658	qtz-syenite	1216.2	1.8	1.5	SG	Hoal and Heaman (1995)
	Hoemoed	Hoemoed 128	PRU113	Granodiorite	1078	nd		MG	Kröner (1976)
	Hoemoed	Excelsior 59	PRU115	Diorite	1080				
	nd	Sesriem 137	KES7	Rhyolite	1170	30	30		Burger and Coertze (1978)
	Nubib	Rooiberg 83	BW1592	Granite	1247	nd	nd		Watters (1974)
		Draaihoek 119	KES9	Granite	1302	20	20		Burger and Coertze (1975)
		Sesriem 137	KES6	Granite	1350	40	40		
		Helmeringhausen 12	No. 64	Granite	1360	50	50		Burger and Coertze (1973)
	Rooiberg	Rooiberg 83	1592b	Granite	1232	nd			Kröner (1976)
		Rooiberg 83	BW1592	Granite	1270				Burger (1974)
		Rooiberg 83	1592a	Granite	1270				Kröner (1976)
	nd	Kanaan 194	KES8	Granite	1236	25	25		Burger and Coertze (1975)
	Hammerstein	Neuhof	HAM01	Tonalite	1380	14	14	SHRIMP	Becker et al. (in preparation)
Rehoboth area									
	Kamasis	Olivantsvloer 453	KES237	Orthogneiss	1328	25	25	MG	Burger and Walraven (1978)
	Kamkam	Swartskaap 332	KES177	Orthogneiss	1406	35	35		Burger and Coertze (1978)

(continued on next page)

Table 2 (*continued*)

Stratigraphic unit	Locality	ID	Lithology	Age	+	–	Method	Reference
Biesiepoort	Biesiepoort 275	KES179	Granite	1110	30	30		
Gamsberg	Nauzerus 1	PH2	qtz-porphyr rhyolite	1010	30	30		Burger and
	Hohenheim 24	PH1	Granite	1078	30	30		Coertze (1975)
	Gamsberg 23	RNA05	Quartz–porphyry	1019	63	41		Nagel (2000)
	Hohenheim 24	PH1	Granite	1080	70	70		Hugo and Schalk (1974)
	Abendruhe 411	PH3	Granite	1089	30	30		Burger and Coertze (1975)
	Gamsberg 23	RNA06	Quartz–porphyry	1095	121	21		Nagel (2000)
	Gamsberg 23	KAW 2273	Granite	1112	26	26		Ziegler and Stoessel (1993)
	Corona 223	RNA15	Porphyritic granite	1116	54	39		Nagel (2000)
Gaoeib	Danigas 209	PH6	Granite	1110	30	30		Burger and Coertze (1975)
Kobos	Nauzerus 1	PH2	qtz-porphyr rhyolite	1010	70	70		Hugo and Schalk (1974)
	Kanaus S 336	KES16	Granite	1064	20	20		Burger and
	Nauchas 14	KES15	Granite	1178	20	20		Coertze (1975)
Koichas	Kangas 371	KES2	Granite	1104	20	20		
Rostock	Rostock South 414	P156	Granite	829	15	15		Pfurr et al. (1991)
	Greylingshof 107	P211	Granite	1084	8	8		
	Greylingshof 107	P217	Orthogneiss	1097	24	24		
	Greylingshof 107	P153	Rhyolite	1099	16	16		
	Rostock South 414	P073	Granite	1194	26	26		
	Rostock North 393	P084	Granite	1207	15	15		
Uitdraai	Ghoab Oos 381	PH8	Granophyr	910	70	70		Hugo and Schalk (1974)
	Ghoab Oos 381	PH8	Granophyr	912	50	50		Burger and Coertze (1975)
	Uitdraai Oos 296	NN	Granite	912	50	50		Hugo and Schalk (1974)
nd	Oorwinning 134	KES5	Granite	948	20	20		Burger and Coertze (1975)
	Oorwinning 134	KES13	Porphyry	976	20	20		Burger and Coertze (1978)
Skumok F	Lekkerwater 144	No 59	Quartz–porphyry	1110	75	75		Van Niekerk and Burger (1969)
Langberg F	Uitdraai Oos 296	PH7	Quartz–porphyry	1083	30	30		Burger and Coertze (1975)
	Rehoboth Station	LAN01	Rhyolite	1100	5	5	SHRIMP	Becker et al.
Kartatsaus F	Kartatsaus 293	KAR01	rhyolite	1090	15	15	SHRIMP	(in preparation)
		PH5	Quartz-feldspar–porphyry	1090	30	30	MG	Burger and Coertze (1975)
		PH4	Quartz-feldspar–porphyry	1090	70	70		Hugo and Schalk (1974)
Nuckopf F	Kuruneib 378	NN	Felsic volcanics	1112	75	75		Schalk (1970)
	Areb North 202	UH1030896-7	Rhyolite	1211	36	29		Hilken (1998)
	Rehoboth Townlands 302	KES174	Quartz–porphyry	1212	30	30		Burger and Coertze (1978)
		SP101	Quartz–porphyry	1216	11	11	SHRIMP	Becker et al. (in preparation)
	Nauzerus Wes 111	N88–124	Rhyolite	1216.4	10.1	10.1	SG	Schneider et al. (2004)
nd	Marienhof 577	KAW3041	Porphyry dyke	1210	7	7	MG	Stoessel and Ziegler (1989)

ID = Sample identification given by the author; nd = not determined, MG = TIMS multi grain, SG = TIMS single grain.

the individual paragneisses distinguished by Toogood (1976) are grouped, currently, into the pre-kinematic Onseepkans Sequence (Table 1). Quartzo-feldspatic gneiss, quartzite, quartz–muscovite schist, calc–silicate gneiss, hornblende agmatite, hornblende gneiss and schist, cordierite–sillimanite–garnet gneiss and biotite gneiss with calc–silicate boudins are the most common rock types. Subdivision of the Onseepkans Sequence into seven formations (Table 1) is based on the pre-dominance of one of these rock types in the respective unit. Intrusions are distinguished into syn-kinematic augengneiss (Grassriver type), norite and enderbite (Stolzenfels type; U–Pb zircon age of 1110 Ma (Toogood, 1976)), mafic granulite (Kumkum type) and amphibolite (Jericho type), alkali felspar granite (Nautsis type), megacrystic granite (Beenbreek type; U–Pb zircon age of 1218 Ma (Toogood, 1976)) and late syn-kinematic granite intermingled with country rock (Naros type). Late- to post-kinematic igneous complexes include layered mafic–ultramafic intrusions (Tantalite Valley Igneous Complex) and pegmatite (Skimmelberg type) (Tables 1 and 2; Kartun, 1979; Kartun and Moore, 1979; Moore et al., 1979).

2.2. Grunau-Rosh Pinah area

More or less continuous exposures of the NMC define a 30 km-wide and 100 km-long strip between Rosh-Pinah to the southwest and Grunau to the northeast (Fig. 3). Along the Lower Fish River – Tantalite Valley shear zone the low-grade Paleoproterozoic Pofadder Terrane in the southwest is overthrust from the northeast (Fig. 3) by the medium- to high-grade, strongly deformed Grunau terrane. Further to the northeast, the northwest-striking, steeply dipping Lord-Hill thrust separates the Grunau terrane from a gneissic complex of 'unknown age and origin' exposed northeast of the thrust (Blignault et al., 1983). Reconnaissance mapping of the Grunau terrane (Beukes, 1973; Blignault, 1977; Schreuder and Genis, 1977) resulted in an informal poorly constrained stratigraphy (Beukes, 1973), featuring three pre-kinematic supracrustal units, i.e. lower 'pink', intermediate 'grey' and upper 'black' gneiss, which, together, define the Garub Sequence around Aus. The 'pink' paragneisses forming the Houms River Formation are exposed in the northeastern and southwestern part of the Grunau terrane (Blignault, 1977; Schreuder and Genis, 1977). They are composed of (1) fine- to medium-grained K–feldspar–quartz paragneiss, with an arkosic precursor (Kröner, 1971; Beukes, 1973; Geringer, 1973) and (2) subordinate quartzite, calc-silicate, marble and amphibolite. Minor felsic metavolcanic rocks occur in the northeast (Schreuder and Genis, 1977). The 'grey' equigranular fine-grained gneiss of the Umeis Formation structurally overlies the Houms River Formation to the west. It contains metasedimentary rocks (biotite/hornblende- or garnet gneiss, quartzite and biotite schist) and metamorphic rocks with probable igneous precursors, which include amphibolite, biotite–hornblende gneiss and porphyritic quartz–feldspar gneiss (Kartun, 1979). A gradual transition between low-strain Paleoproterozoic Vioolsdrif plutonic rocks of the Pofadder terrane (Fig. 3) and this gneiss complex is exposed along the Lower Fish River thrust (Blignault, 1977). A U–Pb multigrain zircon age of 1861 ± 21 Ma was obtained from orthogneiss caught up along the thrust and interpreted as time of crystallization of the magmatic precursor (Kröner et al., 1983), but so far no Paleoproterozoic age was recorded further north of the shear zone. The 'black' gneisses of the Arus Formation consist of metapelites (biotite–garnet and/or sillimanite schist, gneiss and cordierite gneiss), minor quartzite and amphibolite.

Toogood (1976) questioned this stratigraphy which is, however, supported by (1) the regional structural pattern suggesting a synform with the Houms River Formation at the margins and the Arus Formation at the core and; (2) a similar stratigraphic succession within the Garub Sequence of the Luderitz-Aus area. Syn- to post-kinematic intrusions in the Grunau-Rosh Pinah area are coeval and petrologically similar to those exposed in the Aus area. Syn-kinematic intrusive bodies include enderbite and charnockite, granodiorite, augen gneiss and megacrystic garnet–biotite granitic gneiss. The late-kinematic Ai-Ais Igneous Complex represents a layered intrusion exposed over 70 km^2. It intrudes the Umeis Formation and the megacrystic granite gneiss (Blignault, 1977), and includes an early gabbroic phase (subsequently recognised as charnockite by Toogood, 1976), amphibolite to minor tonalite and granodiorite, and late-stage porphyritic granite gneiss. Pink granite intruding the Ai-Ais Igneous Complex is probably genetically unrelated to this layered complex. Small pyroxenite and serpentinite bodies occur in the Karas Mountains (Fig. 3), northeast of the Grunau-Rosh Pinah area. Late-kinematic intrusive bodies comprise small stocks of a granophyric leucogranite and related pegmatites.

2.3. Luderitz-Aus area

Two main supracrustal assemblages were defined north of Aus (Jackson, 1974; McDaid, 1975), whereas correlative supracrustal assemblages near Luderitz have not been differentiated (Greenman, 1966). The older Garub Sequence, which is part of the Grunau terrane (Fig. 3), is made of quartzite, marble, calcsilicate rock, biotite schist, sillimanite–cordierite–garnet gneiss, amphibolite, hypersthene–plagioclase–hornblende–biotite–quartz granulite, and minor ironstone and ultramafic rock (serpentinite and chlorite schist). Supracrustal metasedimentary assemblages are defined by a quartzite–calcsilicate–metapelite succession; gradational changes occur between these lithologies. The Garub Sequence is intruded by pre- to syn-kinematic batholiths, which locally induced the development of migmatites in the country rocks.

The younger Konipberg Formation (McDaid, 1976) is exposed locally and rests with a tectonically overprinted contact on the NMC. It comprises quartz–mica schist, conglomerate, quartzite, mica schist and amphibolite, with minor calcsilicate and pyroclastic rocks. In contrast to the Garub Sequence, it is intruded only by late-kinematic granites and pegmatites cross-cutting an earlier metamorphic fabric. No detailed study has been carried out on this unit and its stratigraphic position with respect to the NMC remains unresolved.

Three igneous suites are known in the NMC of the Luderitz-Aus area and available U–Pb zircon dates are shown in Table 2. Local names appear in brackets, while the regional correlation within the Grunau terrane is given in Table 1. A pre-kinematic dark-grey, medium-grained tonalitic or granodioritic augen gneiss (Tsirub type) postdates the deposition of the Garub Sequence. It is characterized by ovoid quartz–feldspar phenocrysts or feldspar megacrysts converted into phenoblasts/megablasts and a small amount of poikiloblastic pink garnet. The synkinematic igneous rocks include a minor ultramafic phase, which was followed by widespread mafic to felsic plutons, including charnockites, noritic (Konip type) to granodioritic (Jakkalskop, Kunguib, Tschauchaib types) intrusions, mesocratic schlieren-rich granite gneisses (Kubub type), leucocratic garnetiferous megacrystic K-feldspar granites (Aus type), mesocratic biotite granites and fine-grained granite gneisses (Pyramide, Anib, Tiras types). The late-syn- to post-kinematic intrusions range from diorite (Tierkloof type) through granodiorite (Hoemoed type) to granite (Glockenberg, Klein Tiras types) and are combined to define the Naisip River Igneous Suite (Jackson, 1976). There are no reliable geochronological data for these rocks, although the dates presented in Table 2 suggest that the main igneous events might have taken place between 1236 ± 25 Ma (Tierkloof granite, Burger and Coertze, 1975) and 1022 Ma. The Tierkloof granite intruded the Kumbis Complex; it is in turn tectonically overprinted by the Hauchab-Excelsior-Lord Hill shear zone. Therefore, the age datum of 1236 ± 25 Ma constrains a minimum age for an early metamorphic event and a maximum age for the shearing event.

3. Sinclair–Helmeringhausen–Awasib (SHA) terrane

3.1. Kairab–Kumbis Complex (Fig. 4)

A medium-grade metamorphic complex underlies the low-grade Sinclair group supracrustal package (Hoal, 1990). This pre-Sinclair complex consists of isolated occurrences of metamorphosed igneous and sedimentary rocks grouped into the eastern Kumbis Complex (Watters, 1976) and its probable western equivalent, the Kairab Complex (Hoal, 1989, 1990), which is intruded by the Aunis tonalite. The Aunis tonalite yielded a precise U–Pb single zircon age of $1376.5 + 1.7/-1.6$ Ma (Hoal and Heaman, 1995). The spatial distribution of these ≈1.37 Ga igneous complexes is unknown; however, recent SHRIMP zircon ages from the Hammerstein tonalite, intruding amphibolite and metasedimentary rocks of the Neuhof Formation 80 km to the north of the Aunis tonalite, yielded an igneous crystallisation age of 1380 ± 14 Ma (Becker et al., in preparation). This implies that both intrusions are part of a calcalkaline magmatic suite of regional extent and that the Neuhof Formation may form the northern continuation of the Kairab–Kumbis Complex (Becker et al., in preparation). Previously, similar ages were reported from various granitoids in the Rehoboth inlier (Table 2), but need to be confirmed by modern dating techniques.

As part of the SHA terrane, the Kairab–Kumbis Complex is separated from the Grunau terrane by the Hauchab-Excelsior-Lord Hill shear zone (Fig. 4). Medium-grade rocks are exposed on both sides of this tectonic boundary, and it appears that lithologies of the Kairab–Kumbis Complex partly match those of the Garub Sequence. Classified as undifferentiated basement, they include calcsilicate rocks, aluminous gneiss, biotite and biotite–amphibole schist, quartzite, quartz–feldspar gneiss, amphibolite, chlorite schist, layered biotite gneiss, serpentinite and biotite granite gneiss (Hoal, 1990).

In addition, an igneous complex exposed in this area comprises a metavolcanic succession of basalts and rhyolites, with subordinate volcaniclastic rocks. Plagioclase-phyric, high-alumina pillow basalts interbedded with porphyritic, locally flow-banded, rhyolites form a succession with a minimum thickness of ~4000 m. Intrusive mafic and felsic bodies are commonly associated with these volcanic rocks. Metagabbro, diorite and subordinate peridotitic intrusions related to this succession range from a few metres to several kilometres across. One large outcrop of gabbro associated with pillow basalts and a mafic dyke swarm in the Awasib Mountain area is interpreted as a dismembered ophiolite (Hoal, 1990).

Pre-Sinclair granitoids exposed in the Awasib area include unclassified pre-kinematic biotite granite gneiss, megacrystic augen gneiss of granodioritic to tonalitic composition, metafelsite, amphibolite, metagabbro, peridotite and diorite. Post-kinematic tonalite (Aunis type) and granite (Khorasib type) are spatially linked to the metavolcanic succession described earlier (Hoal, 1990).

The Kumbis Complex is intruded by coarse- to medium-grained, equigranular and porphyritic subalkaline granites and minor aplitic rocks (Tumuab type); the pluton is restricted to the area north of the Hauchab-Excelsior-Lord Hill shear zone where it also intrudes granitoid rocks of the Naisip River Suite dated at 1236 ± 25 Ma (Burger and Coertze, 1975).

3.2. Sinclair group

Previous work in the SHA area comprises regional mapping, structural analysis, petrography and geochemical analysis but very little microprobe and geochronological work (Martin, 1965; Von Brunn, 1964, 1967, 1969; Schalk, 1970; Watters, 1974, 1977; Watters, 1982; Harrison, 1979; Hoal, 1989, 1990; Burger and Coertze, 1973; Brown and Wilson, 1986; Hoal et al., 1986; Hoal and Heaman, 1995).

Mesoproterozoic low-grade to unmetamorphic volcanic and sedimentary rocks of this area comprise into the Sinclair Group. They rest disconformably on medium-grade rocks of the Kumbis and Kairab Complexes and are separated from the Grunau terrane to the south by the Hauchab-Excelsior-Lord Hill Lineament. To the north they are separated from the Rehoboth terrane by the NNW–SSE-striking Nam Shear Zone (Figs. 1 and 4). The western part of the area is characterized by rugged inselbergs rising steeply above the sand and scree-covered plains of the Namib Desert, while its eastern limit is defined by a sharp escarpment, which still further to the east is topped by a high plateau of Neoproterozoic and Mesozoic sedimentary rocks.

In the type area, the Sinclair Group includes three conformable volcano-sedimentary cycles deposited in north–south elongated basins (Table 1; Hoal, 1990). The first cycle is represented by the Nagatis Formation (<4000 m thick), which unconformably overlies grey gneisses and granodiorite of the Kumbis Complex. Rhyolitic flows and minor mafic lavas follow upon basal arkoses, grit and shale. They are intruded by coarse-grained granite and granodiorite (115 km^2 outcrop area), rich in mafic enclaves (Haremub, Kotzerus type) and the Tumuab granite of batholithic dimension, all of which predate the second cycle.

The second cycle encompasses the sedimentary Kunjas (Sinclair area) and Urusib (Awasib area) Formations (<2500 m thick), which are composed of two successions of braided river and lacustrine sedimentary deposits with occasional fluvial discharge (Harrison, 1979; Hoal, 1990); they are overlain by the volcaniclastic Barby and Haiber Flats Formations, respectively. Basal pebble conglomerate of local provenance (<120 m thick) or breccia, grit and arkose disconformably overly the Kumbis and Kairab Complexes and the Haremub granite. Where absent, they are replaced by arkosic grit or ferruginous shales locally containing a few thin limestone lenses. Poorly sorted arkosic wacke, siltstone, laminated grey shale and dark-brown to white quartzite define a fining-upward sequence above the conglomerate. Sedimentary structures include planar and trough cross-bedding, ripple marks, parallel lamination, desiccation cracks, scour surfaces, longitudinal bars and channel fill sequences.

In the Sinclair area, the Kunjas Formation (Fig. 4) is overlain by volcanic and volcaniclastic rocks of the Barby Formation (<8500 m thick). Basal felsic ash-flow tuff, stratified tuffite and volcaniclastic conglomerate are followed by massive calcalkaline volcanic and subvolcanic rocks (Table 1; Watters, 1974; Hoal, 1990). This probably includes also a volcanic pile of trachyandesite, trachybasalt and tholeiitic basalt (~520 m thick) of the Guperas Formation (Watters, 1974), which are separated from the lower calcalkaline volcanic rocks by a rhyolitic ash-tuff intercalated with a local conglomerate <80 m thick. Rhyolitic lava flows (>1000 m thick), which form the base of the Guperas Formation, are separated from volcanic rocks of the Barby Formation by clastic sedimentary rocks (<100 m thick). These rhyolitic rocks correlate with those of the Haiber Flats Formation in the Awasib area. Quartz–feldspar porphyries show contorted or laminar flow banding, vesicles and local lithophysae. Domes, plugs (up to 8 km^2 in diameter) and quartz-porphyry dykes of this magmatic event cut across all other lithologies.

The volcanic rocks exposed in the Awasib area were grouped into the Haiber Flats Formation (Hoal, 1990). Basaltic andesites and andesites are overlain by a large volume of rhyodacites and rhyolites. Coarse-grained agglomerates and dome-shaped rhyolitic plugs indicate preserved volcanic centres. This 2100 m thick succession of predominantly andesitic lavas exposed on the eastern flank of the Awasib Mountains was tilted to the east during the emplacement of the coeval subvolcanic Awasib granite (Hoal, 1990) dated by U–Pb single zircon method at 1216.4 ± 1.2 Ma (Hoal and Heaman, 1995). Additional intrusive coeval igneous complexes in the Awasib area include the Chowachasib granite, the Bushman Hill diorite and tonalite and the subalkaline Saffier and Haisib Suites, the latter of which consist of clinopyroxene–orthopyroxene bearing gabbro–monzogabbro, with minor olivine and quartz, and coarse-grained equigranular or porphyritic monzonite–syenites. Plagioclase- or clinopyroxene–phyric monzonite–syenite predominates, whereas olivine–phyric rocks are rare. The emplacement age of the intrusive rocks was determined by U–Pb single zircon method at $1216 + 4.7/-3.5$ Ma (Chowachasib) and $1216 + 1.8/-1.5$ Ma (Haisib; Hoal and Heaman, 1995).

Isolated occurrences of bimodal volcanic and associated sedimentary rocks (Keerweder, Welverdiend, Zwartmodder and Eensam Formations) in the northern part of the Sinclair area have been regarded as lateral equivalents of the Barby Formation (Schalk, 1988). They consist of a basal rhyolitic ash-flow, disconformably overlying the Neuhof Formation. Higher up in the stratigraphy occur alternating amygdaloidal basalt and rhyolite, interbedded with epiclastic conglomerates (Schalk, 1985).

This igneous event ends with the emplacement of granitic batholites including: fine- to medium-grained porphyritic and granophyric red granites in the northern (Nubib type) and southern (Rooiberg type) parts of the basin into which the Sinclair Group was deposited. However, present U–Pb multigrain zircon age data between 1247 and 1360 Ma determined for the Nubib intrusion (Burger and Coertze, 1973, 1975, 1978; Watters, 1974) argue against this stratigraphy.

The Guperas Formation (Fig. 4) disconformably overlies the Barby Formation and the Nubib granite and according to Hoal (1990) marks the third volcaniclastic cycle in the SHA terrane. It is preserved within two north-trending grabens (Watters, 1974). Lithic sandstone and pure cross-bedded white quartzite occur at the base of the succession. Higher up important thickness variations over short distances are common, and lithic sandstone or subgreywacke and conglomerate (0–3700 m thick), with minor shale and siltstone, are the dominant rock types. Cross-bedding is ubiquitious in the sandstones, while oscillation ripples and mudcracks are common in the fine-grained rocks. The clastic rocks are interlayered with and partly overlain by rhyolitic lavas, minor ash-flow tuffs (0–1000 m) and mafic amygdaloidal lavas (<30 m). The volcaniclastic rocks include three successive sheets of ignimbrite up to 35 m thick. The contacts of the basal and upper flows with the underlying sedimentary rocks are sharp, while gradual transitions occur at the base of the second flow and below the sedimentary rocks forming the top of the succession. Minor occurrences of tuff, lapilli-tuff and tuffite are known in the Guperas Formation (Watters, 1974). The Sonntag granite was emplaced towards the end of Guperas deposition; it is considered an equivalent of the Gamsberg granite in the Rehoboth terrane and the Chowachasib granites of the Sinclair–Awasib area (SACS, 1980; Hoal and Heaman, 1995).

4. Rehoboth terrane

A large pre-Neoproterozoic basement inlier is preserved in the Rehoboth area within the southern foreland of the Neoproterozoic Damara belt. The so-called Rehoboth Basement Inlier (RBI) covers more than 10 000 km^2. It is tectonically overlain by Damaran thrust sheets to the north, and unconformably covered by Neoproterozoic to Phanerozoic sedimentary rocks to the south; much of the area is covered by sand of the Kalahari and Namib deserts to the east and west, respectively.

Regional mapping of the Rehoboth terrane (Handley, 1964, 1965; DeWaal, 1966; Schalk, 1970, 1985, 1987, 1988) was followed up by various structural, geochemical and geochronological studies (Hugo and Schalk, 1974; Williams-Jones, 1984; Seifert, 1986; Borg and Maiden, 1986; Borg, 1988; Reid et al., 1988; Pfurr et al., 1991; Ziegler and Stoessel, 1993; Hilken, 1998; Nagel, 2000; Garoeb et al., 2002; Ledru et al., 2002; Schneider et al., 2004), many of which remain unpublished. The isolated Oorlogsende porphyry (Fig. 5), situated close to the Botswana border, was studied by Hegenberger and Burger (1985). It probably represents the eastern continuation of the Rehoboth terrane and provides a link with the Kgwebe volcanic complex of Botswana (Kampunzu et al., 1998).

The RBI is composed of a late Paleoproterozoic domain (1850–1720 Ma; Becker et al., 1998; Ledru et al., 2002) and of a Mesoproterozoic domain. A major unconformity occurs between these two domains; an east west tectonic fabric, formed during a low-pressure/high-temperature metamorphism (De Thierry, 1987), is restricted to the Paleoproterozoic domain. The reconstruction of the original stratigraphy in this area is hampered by the patchy distribution of outcrops, and by tectonic complications (e.g. folding and thrusting during the Damaran orogeny). Therefore, previous models for the evolution of the RBI remain very general (Borg and Maiden, 1986; Borg, 1988). A revised Mesoproterozoic stratigraphy of this area is proposed by Becker et al. (2005).

4.1. Western rehoboth

The classic subdivision of Mesoproterozoic rocks in the western Rehoboth Complex and the correlation with the Sinclair Group (SACS, 1980) is shown in Table 1. Basal sedimentary and rhyolitic rocks of the Nuckopf Formation

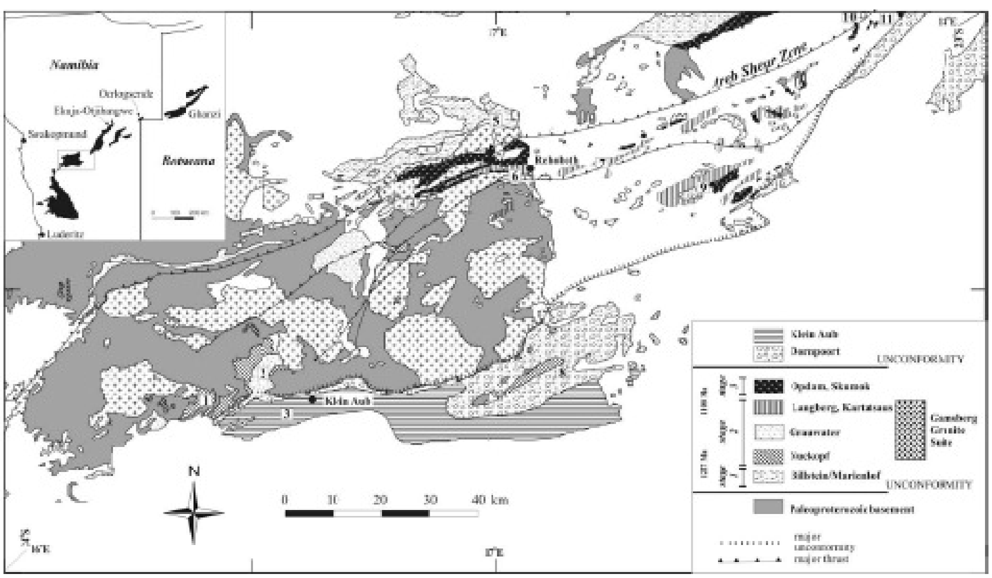


Fig. 5. Simplified geological map of the Rehoboth Basement Inlier. Inset shows the regional distribution of Mesoproterozoic basins and the position of the map. Numbers indicate type localities, which are referred to in the text: (1) Nauzerus (2) Grauwater (3) Klein Aub (4) Marienhof (5) Tsebris Mountains (6) Rehoboth (7) Langberg (8) Witkrans (9) Kartatsaus (10) Opdam (11) Tsumis (Becker et al., 2005).

are overlain by sandstones, conglomerates and minor bimodal volcanic rocks of the Grauwater Formation. Both units were subsequently intruded by the Gamsberg Granite Suite and are unconformably overlain by a thick (>5000 m) monotonous continental red-bed succession (Dornpoort and Klein Aub Formations) and a minor mafic volcanic succession in the lower portion of the Dornpoort Formation (SACS, 1980; Borg, 1988; Schalk, 1988). This classification is opposed by Hoffmann (1989), who regards the regional unconformity at the base of the Dornpoort Formation as the boundary between the Paleo- to Mesoproterozoic Rehoboth and the Neoproterozoic Damaran terranes.

4.2. Central and eastern Rehoboth

In the central and eastern Rehoboth area, Mesoproterozoic volcanic and sedimentary rocks are grouped into three stages (Figs. 5 and 6; Becker et al., 2005):

(1) The basal Billstein Formation includes quartzite and low-grade metapelites previously classified into the underlying Marienhof Formation; both units were previously linked to the Paleoproterozoic Rehoboth Sequence (SACS, 1980). However, the absence of pre-Damara deformation and regional metamorphism and the existence of a gradual transition to the overlying units support that the Billstein Formation marks the base of the Mesoproterozoic era in the RBI (Fig. 6). The onset of sedimentation is not geochronologically constrained, but a minimum age based on a multi-grain zircon date for porphyries intruding these sedimentary rocks is 1210 ± 7 Ma (Ziegler and Stoessel, 1993). Cross-bedded quartzite (<100 m thick) rests unconformably on the Paleoproterozoic basement and grades into interlayered metapsammites and monomict and polymict conglomerates (<500 m thick). Reddish manganese-rich metapelites (<1000 m thick) overlie these metapsammites, with minor intercalated mafic schist bands at the bottom. White quartzite occurs higher up in the succession in the form of lenses and layers (<20 m thick). However, the original position of this lithology remains uncertain as these lenses are highly sheared at their base.

(2) Volcaniclastic successions, dominated by rhyolitic ignimbrites and spanning more than 100 Ma, characterize the second unit (Fig. 6). The Nuckopf, Langberg and Kartsaus Formations and parts of the Grauwater Formation of Schalk (1988) are considered to be the remnants of this ignimbritic system, which is placed into the Nauzerus Group (Becker et al., 2005). The Nuckopf Formation comprises several isolated outcrops. The most complete lithostratigraphic section is exposed west of Klein Aub, (Fig. 5) where clastic sedimentary rocks (~400 m thick) rest unconformably on the Paleoproterozoic Piksteel granite. They consist of layered horizons of phyllitic shale, phyllite, mudstone, gritty sandstone (with cross-bedding) and an ill-sorted polymict conglomerate. Distal hyaloclastic to epiclastic rocks and two horizons of amygdaloidal basalts (<35 m thick) are intercalated with the sedimentary successions. Felsic volcanic rocks (<300 m thick) overlie the interbedded volcanic and sedimentary rocks and, towards the north, transgress unconformably over the Paleoproterozic basement. These volcanic rocks yielded a TIMS U–Pb single zircon date of 1225 ± 10 Ma (Schneider et al., 2004) and a SHRIMP U–Pb zircon date of 1226 ± 11 Ma (Becker et al., in preparation) taken to mark their crystal-

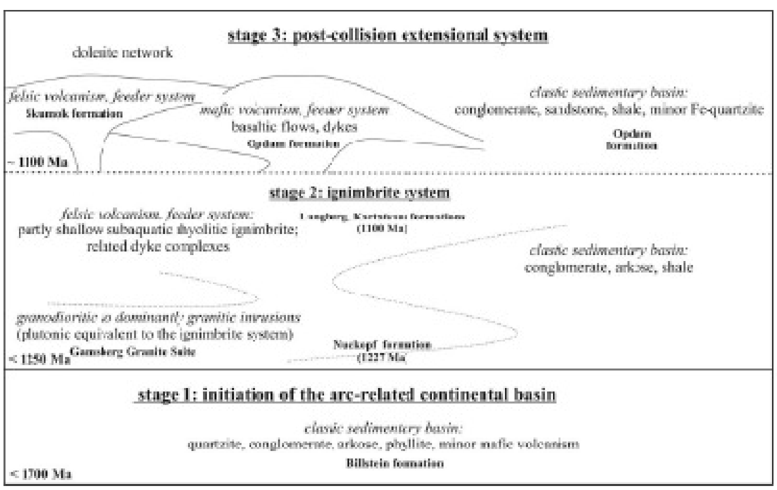


Fig. 6. Proposed model showing the three-stage evolution of the Mesoproterozoic sedimentatary and arc igneous assemblage in the Rehoboth area (Becker et al., 2005).

lization age. The lower portion of the volcanic pile includes abundant basement fragments, while higher up finely flow-banded quartz feldspar porphyry containing local lithophysae are dominant. Pyroclastic rocks ranging from tuff to agglomerate and pyroclastic breccia are subordinate (DeWaal, 1966). At the top of the succession, the volcanic rocks interfinger with sedimentary rocks of the Grauwater Formation. The basal agglomerate, interpreted as a talus breccia, grades into and is overlain by coarse- to fine-grained sedimentary rocks. Thin lenses of rhyolitic tuff, lapilli and agglomerate intercalated with the sedimentary successions prove the overlap between the extrusion of lavas and the deposition of sediments. Higher up in the stratigraphic succession, immature metapsammites with intercalated lenses of conglomerate and slate dominate. Sedimentary structures comprise normal bedding, cross-bedding and ripple marks. Basalts are absent at the type locality and rare elsewhere. Several sedimentary units devoid of felsic volcanic rocks on the northern flank of the RBI were emplaced into the Grauwater Formation on the strength of the lithological similarity of sandstone (Schalk, 1988). At the base of these units, conglomerates of variable lithological character often transgress the Paleoproterozoic basement, while higher up they are intercalated with the sandstone as channels fills. The correlation of these rocks with the Grauwater Formation implies that the sedimentation is distal in relation to the volcanic centres.

The Langberg Formation overlies quartzite, and metapelite of the Billstein Formation to the south of Rehoboth, while the Kartatsaus Formation occurs in an isolated syncline south of the Langberg Formation. High-resolution aeromagnetic data suggest that the Kartatsaus Formation represents a lateral correlative of the Langberg Formation to the west. This correlation is supported by SHRIMP U–Pb zircon crystallization ages of 1100 ± 5 Ma and 1090 ± 15 Ma for igneous rocks from these two units, respectively (Becker et al., in preparation). Ill-sorted clast-supported polymict conglomerates (<200 m thick) occur at the base of both units. Relatively thick metapsammite lenses intercalated locally with conglomeratic wedges of limited lateral extent mark this succession. The sedimentary rocks are overlain by rhyolitic flows (~500 m thick), intercalated with greyish pebbly psammite at their base; the volcanic rocks are succeeded by red quartzite and slate. Within the Langberg Formation, a homogeneous fine-grained volcanic assemblage, traced over >30 km along strike, grade to a coarse-grained quartz–feldspar porphyry to the west and to lithophysae, flow banded rocks and fine grained porphyries to the east. Local very fine-grained rhyolitic tuff marking an extremely strong explosive event was documented. The Kartatsaus Formation comprises silicified pyroclastic rocks, with abundant coarse angular basaltic and rhyolitic clasts. Contorted flow-banding, vesicular lava and lithophysae occur locally, while quartzite and metatuffite form distinct interbedded horizons (5–30 m thick).

Future geochronological analysis will probably lead to a refined stratigraphy of stage 2, distinguishing more volcaniclastic successions in that period. One U Pb multigrain zircon age of 1132 ± 75 Ma from rhyolite grouped presently into the Nuckopf Formation (Table 2) supports this assumption.

Batholitic granite complexes emplaced during and at the end of the second unit in central and eastern Rehoboth form the poorly defined Gamsberg Granite Suite (GGS), which itself is intruded by quartz–porphyry and dolerite dykes. U–Pb zircon multigrain dates are in the range 1000–1250 Ma (Table 2) and argue for a long-lasting episode of continuous plutonism, accompanying the volcanism of the Nauzerus Group. However, none of the ages is considered to be reliable and modern geochronological analysis is needed to establish within that period distinct magmatic suites, which could better document the continuity of magmatic activity than the more rarely preserved volcanic rocks. At the regional scale, the GGS was correlated with the late-stage Sonntag and Chowachasib granites of the Sinclair–Awasib area (SACS, 1980; Hoal and Heaman, 1995). The mineral and textural composition varies considerably in the GGS, the most common types being biotite-bearing granites with a medium-grained equigranular (Klein Gamsberg type), seriate (Kobos type) or porphyritic (Hornkrantz type) texture. Less frequent are biotite–hornblende–titanite granodiorites (Namibgrens, Biesiepoort and Narabis types). Intrusions of smaller size are probably younger, but this should be constrained using geochronological data. These small intrusions comprise reddish to greyish subalkaline granites with or without hornblende (Uitdraai type), pinkish aplite granite with minor mica (Neuras, Koichas types), and a light medium-grained biotite granite (Tierkolkies type). Rhyolitic and microgranitic enclaves are frequent within these granites.

(3) The third unit in central and eastern Rehoboth includes clastic sedimentary and mafic igneous rocks of the Opdam Formation (Figs. 5 and 6). It overlies the Langberg Formation in the Rehoboth area, while further east it transgresses the quartzites of the Billstein Formation. The basal clastic succession (10–150 m thick) comprises greyish quartzite, polymict conglomerate and intercalated slate. It is overlain conformably by basaltic amygdaloidal flows, flow breccias and rare pillow basalts (>1800 m thick), which are interbedded with pink to brown cross-bedded to ripple-marked quartzite, magnetite-bearing quartzite, conglomerate and pebbly quartzite, phyllite and calcareous phyllite. Volcanic breccias occur above the lava flows and often grade into quartzite layers (Williams-Jones, 1984). Pre-kinematic epidotite stockwerk occurs within the basaltic flows and breccias.

On the regional scale, the deposition of clastic rocks outlasted the basaltic volcanism, and resulted in a >1000 m thick sedimentary succession defining an upward fining sequence, from conglomerate to reddish pebble- and fragment-rich quartzite. Four horizons of rhyolitic volcanic rocks (<50 m thick) are intercalated within the quartzites at the top of this unit. They include quartz porphyry and pyroclasts, suggesting a resurgence of felsic volcanism (Skumok Formation). Numerous mafic dykes, ranging from coarse-grained gabbro to dolerite, form a dense network inferred to be the feeder dykes of the basaltic flows (Ledru et al., 2002). The final infill of the basin consists of planar-bedded immature sandstones.

5. Mesoproterozoic basement inliers within the Damaran (Pan-African) orogen

A number of basement inliers occur within the Pan-African Damara belt (Fig. 1). Little is known about these inliers; however, geochronological data indicate the existence of Mesoproterozoic tectonic blocks.

5.1. Ekuja–Otjihangwe Nappe Complex (EONC)

This poorly exposed inlier is situated 120 km northeast of Windhoek within the southern zone of the Damara belt (Fig. 1). It has attracted some interest because of its potential for copper, with reserves estimated at $6 \times 10^5 t$ (Steven et al., 2000). The main rock assemblage includes banded biotite paragneiss, migmatitic gneiss, augen gneiss and foliated amphibolite (Kasch, 1986, 1987; Steven et al., 2000). The precursors of the gneisses are assumed to be basaltic andesite, dacite, tonalite and rhyolite. SHRIMP U–Pb zircon dates between 1063 ± 9 and 1115 ± 13 Ma are taken to mark the igneous crystallization ages of these rocks (Steven et al., 2000).

The EONC was originally thought to belong to the poorly investigated Hohewarte Complex further south, which may represent its continuation. It is composed of quartzite, amphibolite, calcsilicate and feldspar gneiss, intruded by undated pre-Damaran porphyritic granite.

5.2. Abbabis Inlier

Isolated basement domes within the central zone of the Damara belt are grouped in the Abbabis Metamorphic Complex (Figs. 1 and 7; Smith, 1965; Jacob, 1974; Marlow, 1981; Sawyer, 1981; Brandt, 1987). In the east, a basal Tsawisis Formation consists mainly of metapelite, gneiss, meta-arkose, and subordinate marble, calc-silicate rocks and conglomerate (Brandt, 1987). It is overlain by metapelite and amphibolite forming the base of the Naob Formation, which are succeeded by quartzite, marble, calc-silicate and metavolcanic rocks. Granitoids of the Narubis igneous complex intruded these supracrustal rocks and form the widespread augen gneiss of the Abbabis Metamorphic Complex. Cross-cutting mafic dykes now converted into amphibolite are ubiquitous. In the west, medium- to coarse-grained quartzo-feldspathic gneiss, quartzite, micaceous quartzite and cordierite schist constitute a lower metasedimentary sequence (Sawyer, 1981). It is overlain, probably unconformably, by an upper metasedimentary sequence consisting of basal calc-silicate rocks and marble passing into a sequence of gneisses, schist and glassy quartzite. The eastern part of this domain contains a large body of coarse-grained augengneiss of uniform appearance.

SHRIMP U–Pb zircon dates of 2038 ± 5 Ma for augen gneisses from the Abbabis Metamorphic Complex in the Ida Dome area are interpreted as time of igneous crystalli-

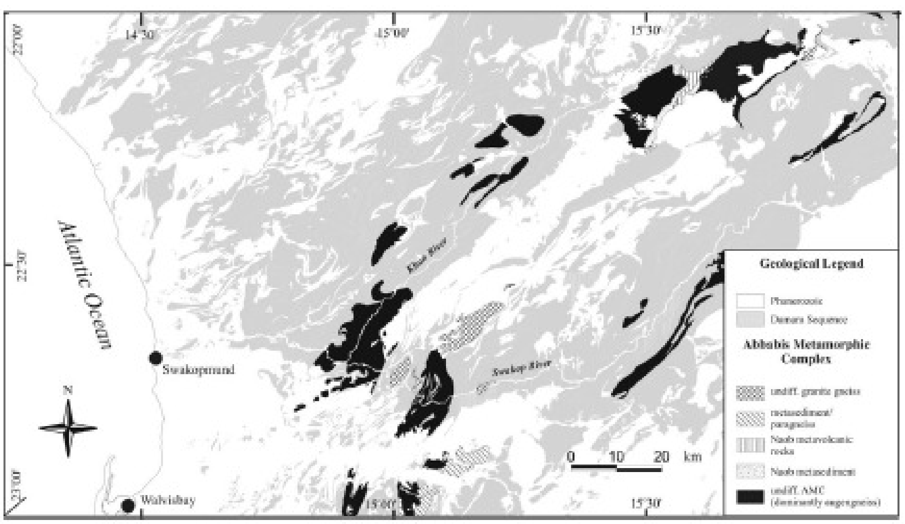


Fig. 7. Simplified geological map of the Abbabis Metamorphic Complex within the Neoproterozoic Damara orogen.

zation and constrain a minimum age for the supracrustal rocks in this complex (Tack and Bowden, 1999). A multi-stage history is indicated by the complex internal structures of the zircons with magmatic zoning, metamorphic overgrowths and various inclusions. The rims yielded a late Neoproterozoic age of 526 ± 17 Ma. Kröner et al. (1991) reported U–Pb single zircon dates between 1240 and 1040 Ma for augen gneisses near the Khan–Swakop confluence; they were interpreted as time of crystallisation of the igneous protolith. A xenocrystic grain yielding an age of 2014 ± 39 Ma suggests the ultimate source of this gneiss to be Paleoproterozoic basement. Nd-model ages of 1.0–1.2 Ga (Hawkesworth et al., 1981) determined on overlying Panafrican sediments imply that a Mesoproterozoic rock province was the dominant source for these rocks arguing for its regional distribution.

5.3. Central/Western Kaoko zones

Two major shear systems (Purros lineament and Sesfontein thrust) subdivide the Pan-African Kaoko orogenic belt of northwestern Namibia into three distinct terranes: the eastern (EKZ), central (CKZ) and western (WKZ) Kaoko zones (Figs. 1 and 8). All three terranes comprise Neoproterozoic metasedimentary rocks and granitic orthogneisses and a poorly documented Pre-Neoproterozoic basement (Miller, 1983; Goscombe et al., 2003a,b; Kröner, 2003). Greenschist facies rocks prevail in the EKZ, high-pressure/low-temperature (HP–LT) metamorphism marks the CKZ and low-pressure/high-temperature (LP–HT) metamorphism (up to granulite facies) characterizes the WKZ, dated by Sm–Nd whole rock-garnet method at 595 ± 13 and 573.2 ± 7.6 Ma (Goscombe et al., 2003a). Pan-African oblique thrusting with a sinistral component along the Purros and Sesfontein shear zones juxtaposed the EKZ, CKZ and WKZ. Pre-Neoproterozoic, migmatites and augen gneisses within the WKZ yielded U–Pb single zircon dates of 1701–1683 Ma and 1510–1490 Ma, respectively (Kröner, 2003). One Sm–Nd whole rock-garnet age of 1239 ± 69 Ma (Goscombe et al., 2003a,b) is regarded as a mixing age; compositional maps from the sample show two phases of garnet growth, the younger of which is interpreted to be Damaran in age.

5.4. Epupa Metamorphic Complex–Kunene Anorthosite Complex

The late Archean to Mesoproterozoic Epupa Metamorphic Complex (EMC) of northern Namibia is regarded as the southwestern margin of the Congo Craton (Figs. 1 and 9; Tegtmaier and Kröner, 1985; Brandt et al., 2003; Seth et al., 2003). The EMC comprises a generally poorly investigated sequence of medium- to high-grade para- and orthogneisses. Two metamorphic complexes were defined in the northeastern corner of the EMC (Brandt et al., 2003). The east–west trending Epembe terrane (≈50 km-long and 10 km-wide; Fig. 9) comprises mafic

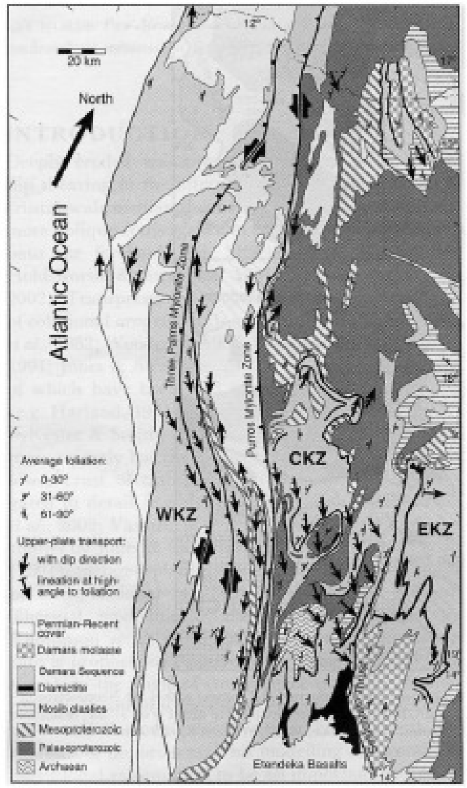


Fig. 8. Simplified geological map of parts of the Kaoko belt showing transport direction in a transpressional regime during the Pan-African orogeny (Goscombe et al., 2003b).

and felsic granulites and paragneisses. Peak metamorphic conditions under ultrahigh temperatures (950 °C), medium pressures and a clockwise P–T–t path were documented for these granulites (Brandt et al., 2003). U–Pb single zircon dates between 1520 and 1510 Ma for these rocks (Seth et al., 2003) probably represent the igneous crystallization ages of the protolith of the granulites, whereas Pb–Pb step-wise leaching TIMS garnet and retrograde sapphirine age data from 1490 to 1447 Ma are thought to mark the granulite facies metamorphic event (Seth et al., 2001, 2003). The Orue terrane (Fig. 9) comprises migmatitic ortho- and paragneisses evolved under upper amphibolite facies metamorphism (650–720 °C at 6–7 kb). The Otjitambi–Ehomba shear zone separates the Epembe and Orue terranes.

The Kunene Anorthosite Complex (KIC) is ∼350 km-long and ∼80 km-wide, and represents one of the largest anorthosite bodies in the world (Silva, 1990). The KIC

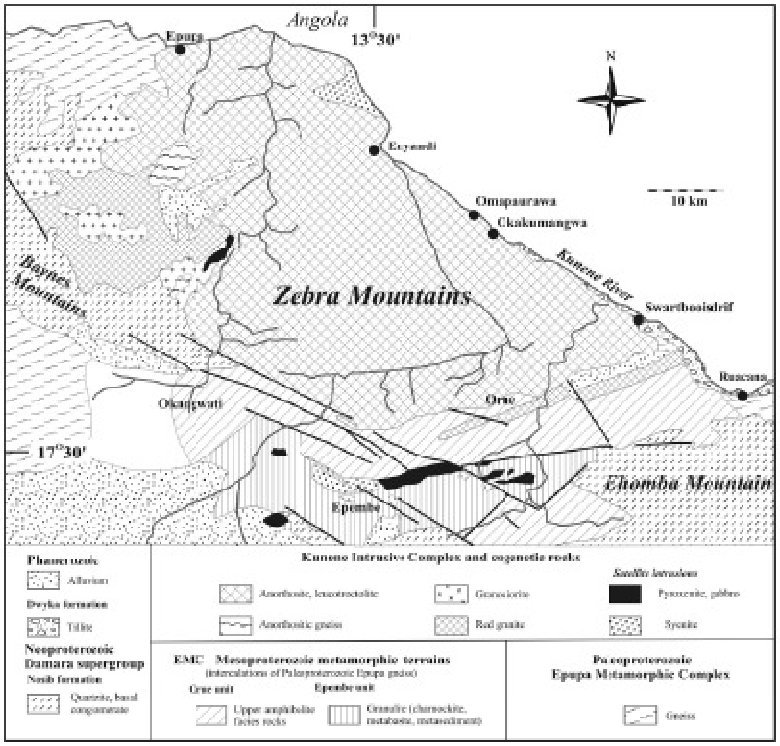


Fig. 9. Simplified geological map of parts of the Epupa Metamorphic Complex and the Kunene Intrusive Complex (Seth et al., 2003 modified after Menge, 1998).

straddles the boundary between Namibia and Angola and intrudes Paleoproterozoic rocks (Tegtmaier and Kröner, 1985). This intrusion is undeformed and unmetamorphosed to the north, particularly in Angola (Morais et al., 1998), but has been affected by Pan-African deformation and metamorphism in Namibia (Drüppel and Okrusch, 2000; Drüppel et al., 2000, 2001). The KIC is essentially composed of a layered sequence of anorthosite, leucotroctolite and troctolite, with subordinate noritic and gabbroic rocks (Köstlin, 1974; Silva, 1990; Von Seckendorff et al., 2000; Drüppel et al., 2001). A U–Pb single zircon date of a mangerite vein from the KIC in Angola constrains the emplacement of the KIC at 1370 ± 4 Ma (Mayer et al., 2000).

6. Metamorphism and structural development

6.1. Namaqua Metamorphic Complex in Namibia

The metamorphism and structural development of the Namaqua Metamorphic Complex has been dealt with in numerous studies (Blignault, 1974a,b, 1977; Beukes and Botha, 1975a,b; Jackson, 1976; Toogood, 1976; Kröner and Blignault, 1976; Kartun, 1979; Blignault et al., 1983; Hartnady et al., 1985; Van Aswegen et al., 1987; Cilliers and Beukes, 1988; Hoal, 1990; Van der Merwe, 1995; Colliston and Schoch, 1996, 1998, 2000; Visser, 1998). Comprehensive reviews of the tectono-metamorphic evolution of the NMC in Namibia were given by Blignault et al. (1983), Van Aswegen et al. (1987), Hoal (1990) and Colliston and Schoch (2000), and the main features recognized by them are summarized below. The structural grain of the NMC comprises northeast-plunging stretching lineations and long axes of mega- and mesoscopic sheath folds, with a foliation trending NW–SE, which is rotated in the plane of maximum shear. Southwest-directed transport in the NMC in Namibia occurred under ductile mid-crustal metamorphic P–T conditions (Blignault et al., 1983; Colliston and Schoch, 2000). Crustal-scale thrusts with transport >100 km delineate the boundary between the Pofadder and Grunau terranes. Structural maps of individual areas show a complex juxtaposition of tectonic blocks displaying distinct deformation histories. Metamorphic inversion along the Lower Fish River – Tantalite shear zone (Fig. 3) is shown by greenschist facies rocks to the south of the shear zone, structurally overlain by a granulite facies package to the north of the thrust. Both blocks are affected by late north–south trending open folds. The mineral assemblages in the Grunau terrane (Jackson, 1976; Toogood, 1976) grew under LP/HT metamorphic conditions.

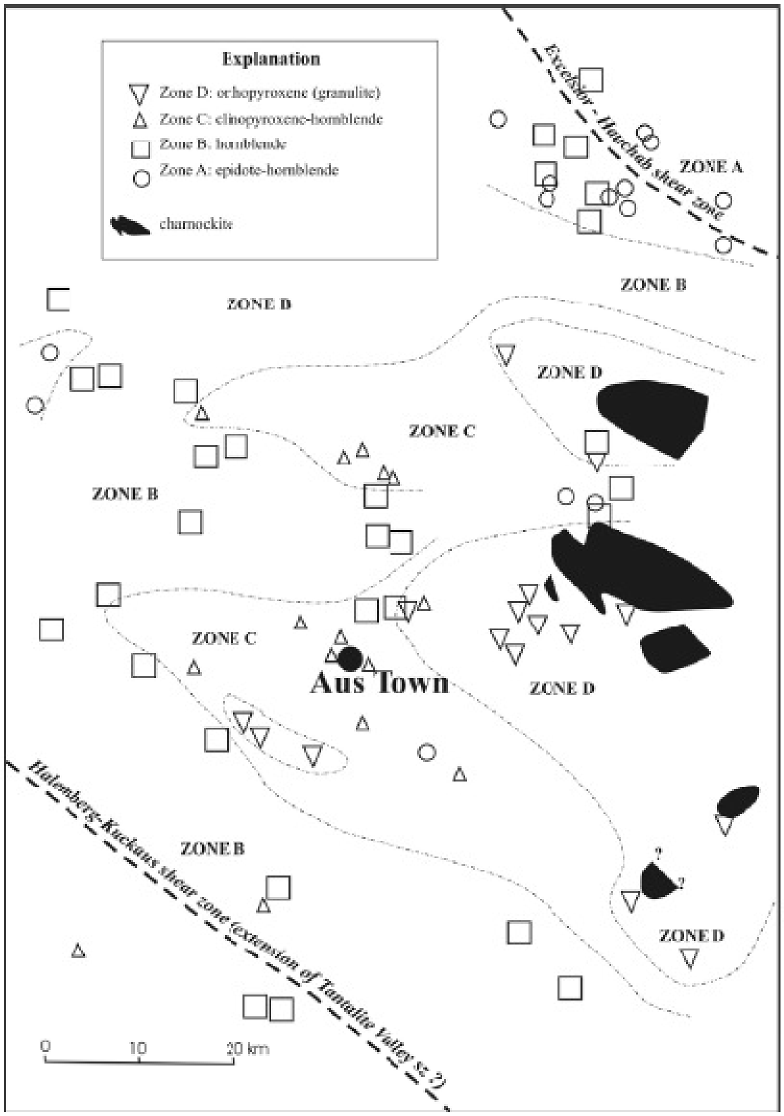


Fig. 10. Distribution of Mesoproterozoic metamorphic index minerals and isograds in the Namaqua Metamorphic Complex (NMC), Aus area, Namibia (Jackson, 1976).

In the Aus area (Figs. 2 and 10), there is a general increase in prograde metamorphism from amphibolite facies (625 °C) in the northwest to granulite facies (780–860 °C) in the southeast; pressure estimates for both facies are given at 6 ± 2 kb (Jackson, 1976). Similar T-peak conditions (800–860 °C) under higher pressures (8–9 kb) mark the Grunau terrane in the Warmbad-Karasburg area (Toogood, 1976). A second metamorphic event is recorded in that area in both the Grunau and Pofadder terranes (Toogood, 1976); migmatites of the Aus area (Jackson, 1976) possibly formed during M2. The widespread migmatisation in the Grunau terrane occurred under P/T-conditions of 5–7 kb and 650–700 °C. Polycyclic reactivation of major shear zones (Figs. 1 and 3) has been documented (e.g. Jackson, 1976; Toogood, 1976; Kartun, 1979; Jacobs et al., 1993; Colliston and Schoch, 2000) using superimposed shear sense indicators for the succession right-lateral–vertical movement (e.g. The Lower Fish River – Tantalite Valley shear zone), or both reverse and normal displacement (e.g. Hauchab-Excelsior-Lord Hill shear zone). Late-kinematic intrusions (e.g. Tantalite Valley Igneous Complex) are coeval with a regional metamorphic event under P/T-conditions of ~4–5.5 kb and ~580 °C.

6.2. Sinclair–Helmeringhausen–Awasib, Rehoboth areas

The oldest tectonic fabric within the Kairab Complex is a regional gneissic fabric or migmatitic layering, which exhibits D_1 deformation event related isoclinal folds. It is overprinted by tight to overturned D_2 folding. The mineral lineation, formed during the D_2 event, is dispersed because of its refolding during D_2, but a down-dip component dominates. Amphibolite facies metamorphic conditions prevailed during D_1 and D_2. A minimum age of 1370 Ma for these deformation events is constrained by the Aunis tonalite, which is not affected by these two deformation events. However, a regional NNW-trending fabric (S_3) and locally developed mineral lineations with a heterogeneous distribution post-dating the Aunis intrusion and predating the deposition of the Sinclair Group, were linked to transcurrent tectonics (Hoal, 1985, 1990).

The Sinclair Group and its northeastern equivalents are weakly folded and display low-grade greenschist or sub-greenschist metamorphic assemblages. The Awasib area is characterized by open folds and NW–SE-trending dextral and sinistral shear zones (Hoal, 1990). In contrast, the Nam Shear belt is characterized by a subvertical mylonitic foliation and subhorizontal stretching lineations. In the Rehoboth region, Neoproterozoic Pan-African structures (not described here) pre-dominate and have completely overprinted possible older structures (Ledru et al., 2002).

7. Geochemistry

There are no geochemical data on igneous rocks exposed in the NMC. All the data discussed below come from the SHA and Rehoboth terranes and for better comparison were recalculated to 100% LOI free. The quality of the data is variable and REE and several critical trace elements are available only for a small number of samples (Hoal, 1990; Becker et al., 2005).

7.1. Mafic rocks

This section covers the composition of metabasalts and gabbros marked by SiO_2 contents <53 wt.% wt. and Mg# ($Mg/Mg + Fe^{2+}$ with $FeO/Fe_2O_3 + FeO$ normalized at 0.8) in the range 0.4–0.7. The metabasalts of the Kairab Complex are characterized by the following composition: SiO_2: 46.5–50.3 wt.%, Mg#: 0.4–0.6, TiO_2: <0.5 wt.%, Al_2O_3: 17–21 wt.%, Rb: <20 ppm, Ba: <300 ppm, Sr: <300 ppm, Zr: <50 ppm, Y: <20 ppm and Ti/V: <10. The composition of these mafic rocks (Fig. 11) resembles that of island arc calcalkaline basalts (Hoal, 1990).

The basalts of the Barby and Haiber Flats Formations (Watters, 1974), are characterized by slightly higher values for SiO_2 (52–53 wt.%), Mg# (0.48–0.51), TiO_2 (0.7–0.9 wt.%), Al_2O_3 (<17 wt.%), K_2O (3.5–5.5 wt.%), Rb (>100 ppm), Ba (>1000 ppm), Sr (>500 ppm), Zr (>90 ppm), Y (>20 ppm) and Ti/V ~ 20. They represent high-K calcalkaline basalts (Fig. 11).

The basalts of the Opdam Formation (Williams-Jones, 1984; Becker et al., 2005) are tholeiitic and marked by the following geochemical characteristics: SiO_2: 46–52 wt.%, Mg#: 0.40–0.75, TiO_2: 0.7–2 wt.%, Al_2O_3: ⩽17 wt.%, K_2O: 0.1–1.3 wt.%, Rb: <20 ppm, Ba: <300 ppm, Sr: <400 ppm, Zr: <100 ppm, Y: 15–30 ppm, Ti/V: >20, Th/Ta: >30, La/Nb: 1.5–10 and Ce/Pb: 2–5. Their chondrite-normalized REE patterns are flat (Fig. 12), with an overall enrichment factor of ~10 for most elements. Primordial mantle normalized spider diagrams display Nb and Ta negative anomalies (Fig. 12). These mafic rocks show affinities with late orogenic continental tholeiites (e.g. Kapenda et al., 1998), with most immobile elements showing the influence of subduction-style mantle enrichment at their source (cf. high Th/Ta and La/Nb vs. low Ce/Pb values).

7.2. Intermediate and felsic rocks

The Aunis tonalite of the Kairab Complex represents a low-K tonalitic suite ($Na_2O/K_2O > 1$), with a few samples showing a slight trondhjemitic affinity (Fig. 13). They represent arc-granitoids (Fig. 13). The Haiber Flat and Barby Formation volcanic rocks define a high-K calcalkaline suite. The Awasib and Chowachasib granites in the AHS terrane represent intrusive equivalents of these calcalkaline volcanic rocks with which they share geochemical characteristics. The same holds true for the 1227 Ma Nuckopf and 1100 Ma Langberg Formation volcanic rocks in the Rehoboth terrane, whose intrusive equivalents are grouped into the poorly defined Gamsberg Granite Suite, which consists of metaluminous to slightly peraluminous I-type calcalkaline granitoids (Fig. 13).

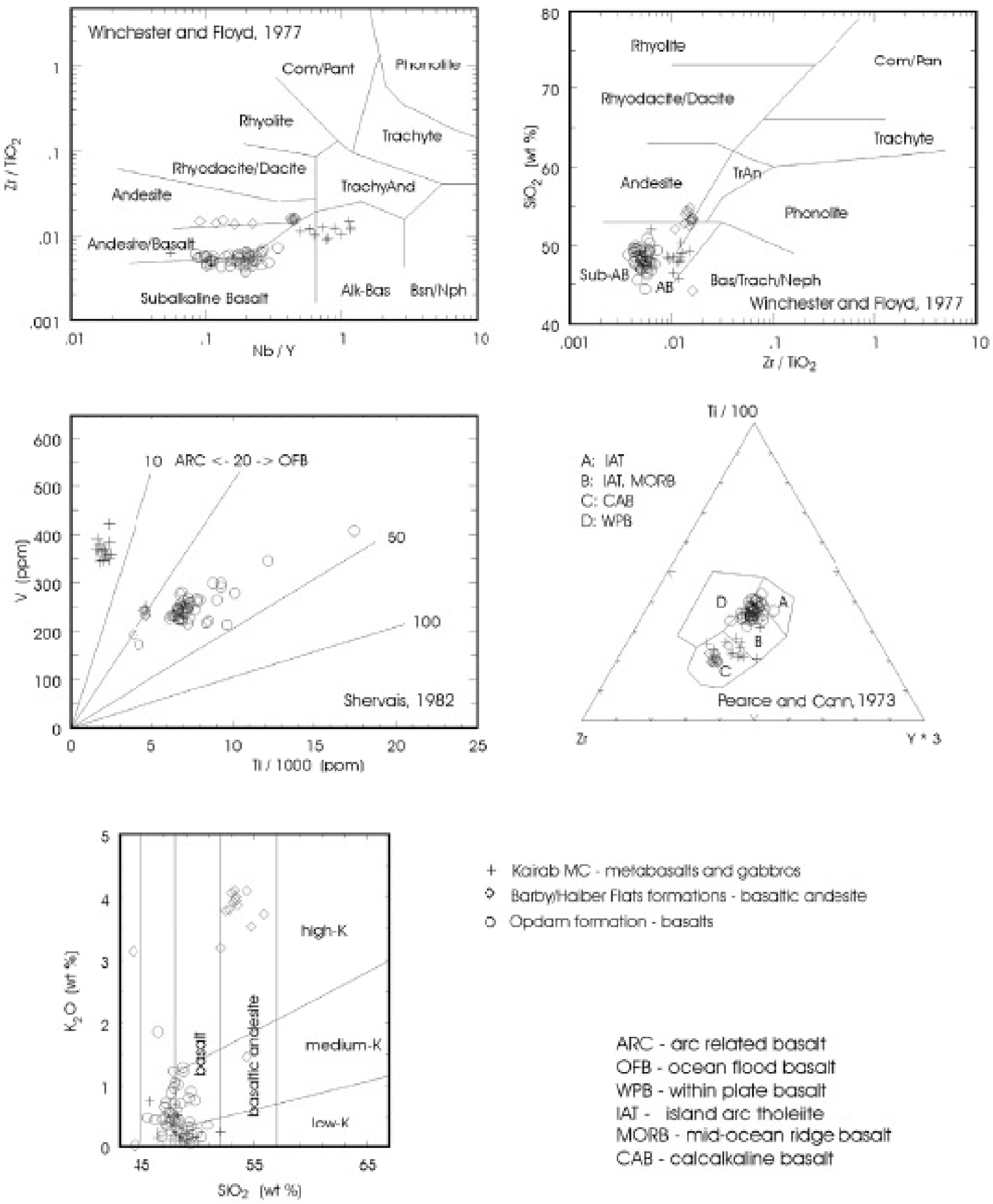


Fig. 11. Geochemical characteristics of Mesoproterozoic mafic igneous rocks in Namibia (data from Watters, 1974; Hoal, 1990; Ziegler and Stoessel, 1993; Becker et al., 2005). See also Pearce and Cann (1973), Winchester and Floyd (1977), Shervais (1982).

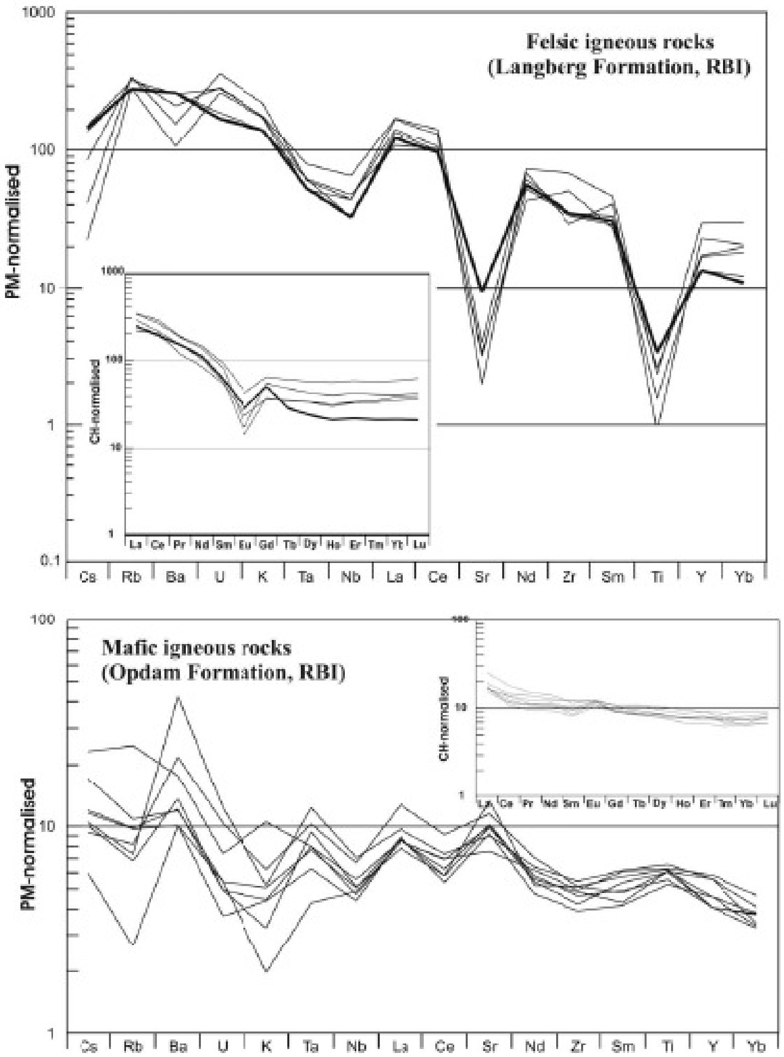


Fig. 12. Representative chondrite (CH)-normalised rare-earth-element (REE) patterns and primordial mantle (PM)-normalised multi-element diagrams for Mesoproterozoic mafic (Opdam Formation) and felsic (Langberg Formation) igneous rocks in Namibia (data from Becker et al., 2005; normalizing values of Sun and McDonough, 1989).

Dacites and rhyolite of the Langberg Formation are medium- to high-K rocks (Na_2O/K_2O: 0.3–0.7) and straddle the fields of volcanic arc and within-plate felsic rocks in tectonic setting discrimination diagrams (Fig. 13). Important to note is that these lavas overlap the field of post-collisional felsic igneous rocks (Pearce, 1996). Chondrite-normalized REE patterns are moderately steep (Fig. 12), with extremely variable normalized values in the range 4–200. Primordial mantle normalized spider diagrams display negative anomalies for Sr–Nb–Ta–Ti, with a general increase of normalized values from moderately to highly incompatible elements.

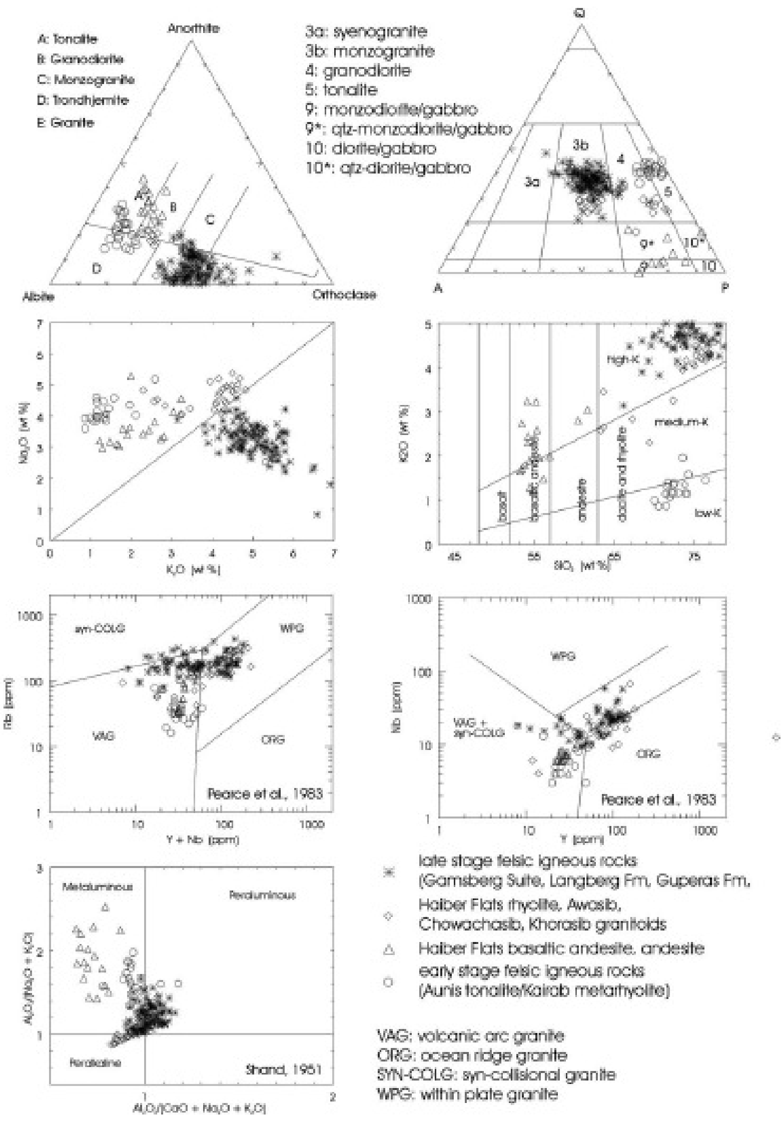


Fig. 13. Geochemical characteristics of Mesoproterozoic felsic igneous rocks in Namibia (data from Watters, 1974; Hoal, 1990; Ziegler and Stoessel, 1993; Becker et al., 2005). See also Shand (1951) and Pearce et al. (1983).

8. Discussion

8.1. NMC–SHA, Rehoboth terranes, EONC, Abbabis inlier

The main findings for discussion are listed below:

(1) The NMC comprises a metasedimentary assemblage of unknown deposition age, which is composed of quartzite, calc–silicate and metapelitic rocks. Colliston and Schoch (2000) suggest a Paleoproterozoic fore-arc setting during the deposition of this

assemblage, which was related to the Poffadder terrane. However, its composition suggests a shelf marking a passive margin sequence (Jackson, 1976), which may have been deposited in early Mesoproterozoic times before the onset of subduction.

(2) The SHA terrane encompasses the Kairab/Kumbis Complex, which was affected by amphibolite-grade metamorphism prior to being intruded by the 1.37 Ga Aunis tonalite. It includes high-alumina/low-potassium pillow basalts, showing affinities with island arc basalts. Hoal (1990) correlated the Kairab Complex with the Garub sequence of the NMC on the basis of lithological similarities. However, carbonates and calcsilicate rocks are absent in the Kairab Complex, whereas Mesoproterozoic metaluminous granitoids older than 1.2 Ga are unknown in the NMC.

(3) 1.23–1.1 Ga low-grade to unmetamorphosed volcanic and sedimentary rocks of the Sinclair Group and its northwestern equivalents rest disconformably on the Kairab Complex. These supracrustal sedimentary and volcanic rocks occur in fault-bounded depocenters, which define a regional arc-shape structure >100 km-wide and >2000 km-long (Figs. 1 and 2). The ≈1.2 Ga volcanic Barby Formation and coeval granitoids comprise high-K calcalkaline mafic, intermediate and felsic rocks. Sedimentary rocks comprise immature and minor mature siliciclastic rocks, which are interlayered with pyroclastic rocks. The 1.1 Ga rhyolites of the Langberg and Kartatsaus Formations are intruded by syenogranites that are thought to have A-type characteristics, although a final interpretation awaits detailed geochemistry.

(4) Mafic rocks characterize the overlying Opdam Formation. High Ti-content and flat REE-patterns in these tholeiites suggest their emplacement in an extensional setting, whereas high Th/Ta and La/Nb ratios, low Ce/Pb values and negative anomalies for Nb–Ta in the PM-normalized spiderdiagrams mark a subduction-style enrichment of the mantle source of these mafic magmas. These data require the involvement of the asthenosphere and a lithospheric mantle enriched during a subduction event during the genesis of these mafic igneous rocks.

(5) Late-Mesoproterozoic (1.2–1.05 Ga) meta-igneous rocks from inliers within the Damaran belt (Namibia) and its lateral correlative, the Ghanzi-Chobe belt (Botswana), document the wide distribution of these rocks to the north and northeast of the AHS and Rehoboth terranes. Nd-model ages of 1.0–1.2 Ga (Hawkesworth et al., 1981), determined on overlying Panafrican sediments, suggest that the Mesoproterozoic terrane may underlie the entire southern Damara belt, the Khomas trough (Kröner et al., 1991).

(6) The main transport direction during the 1.1–1.03 Ga Namaqua orogeny is towards the southwest, with a juxtaposition of terranes along crustal-scale shear zones, resulting in a metamorphic inversion along the Lower Fish River – Tantalite Valley shear zone, where granulite facies rocks structurally overlie greenschist facies rocks. Serpentinite occurrences located along the Hauchab-Excelsior-Lord Hill shear zone, decorate the boundary between the NMC and SHA-terranes.

(7) LP/HT granulites of the NMC crystallized under the same P/T conditions as their correlatives exposed in the O'kiep area of South Africa (Fig. 1), where they are precisely dated at 1060–1030 Ma. This metamorphic event is characterized by an anti-clockwise P–T–t path (Waters, 1989, 1990; Robb et al., 1999).

(8) The timing of magmatic activity in the NMC is poorly constrained. Igneous rocks comprise ultramafic, mafic, intermediate and felsic members.

8.2. Geotectonic interpretation

The onset of the Wilson-cycle rifting-drifting stage is at present not constrained within the NMC, although the paragneisses of the Garub sequence could represent a passive margin succession.

A period of northeast-directed subduction, including island-arc accretion, is recorded in the Kairab Complex (Hoal, 1990; Fig. 14), and presumably started before 1370 Ma, which is the age of the oldest subduction-related igneous body. It is known that subduction-related magmas are generated only when the developing slab reaches a depth of ~100 km, i.e. the subduction starts several million years before the generation of arc magmas. The regional Mesoproterozoic convergence vector (Fig. 15) deduced from the arc-shape of the Mesoproterozoic magmatic arc in Namibia was broadly east–west in the present-day coordinates (Watters, 1976; Hoal, 1993). Tectonic truncation as possible cause for this shape can be excluded by the arrangement of gravity anomalies and the regional magnetic pattern enveloping the Kalahari Craton (Kleywegt, 1967; Fig. 2, Wackerle, unpubl.). The ⩾1.37 Ga amphibolite facies metamorphism, migmatisation and isoclinal folding in the Kairab–Kumbis Complex is linked to the arc accretion.

The high-K calcalkaline volcanism in the Sinclair area at ≈1216 Ma requires subduction beneath a continental crust. This continental crust section (overriding plate) is part of the Kalahari craton, while the NMC represents the leading edge of the western craton. The Hauchab-Excelsior-Lord Hill shear zone marks the boundary between the two converging plates. This is in contrast to a model, which interprets the Lower Fish River – Tantalite Valley shear zone further south as the main suture (Hoal, 1990). At ≈1.1 Ga outpouring of continental tholeiites originating from a lithospheric mantle section, enriched during subduction, mark a major change in tectonic processes, i.e. a switch from subduction to asthenosphere uprise, triggering partial melting of a lithospheric mantle section enriched

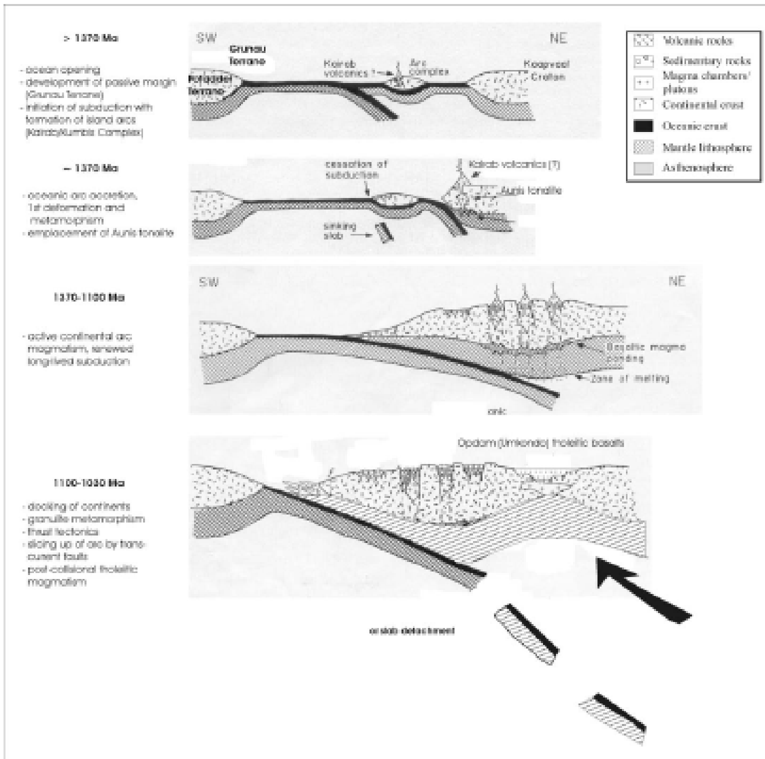


Fig. 14. Geological model for the evolution of the Mesoproterozoic terranes in Namibia encompassing the NMC, SAH and Rehoboth domains (modified after Hoal, 1990).

during the subduction stage. We believe that docking of continents led to the slab detachment, allowing the interaction between the asthenospheric mantle and the mantle wedge enriched during the subduction process (Kampunzu et al., 1998). The magmatic underplating related to this event induced an anti-clockwise P–T–t path with a high-grade LP/HT metamorphism recorded in the NMC granulite facies rocks. Furthermore the Umkondo large igneous province in the Kalahari craton (Hanson et al., this volume) is related to the same event. Late arc-splitting and subsequent back-arc spreading could also induce this upwelling of asthenospheric mantle leading to partial melting of the lithosphere to generate tholeitic volcanism. However, this interpretation is not supported by the fact that this tholeiitic magmatism is widespread outside the arc system and extends well into the Kaapvaal craton. The intracontinental rift model (Borg, 1988) has been discussed and argued against by Hoal (1990); the sedimentology and geometry of the basins, the apparent bimodal character of the volcanic rocks and their younging ages towards the northeast, and geophysical patterns (i.e. gravity anomalies) cited in favour of such a model have been shown to be wrong or, alternatively, do not constrain this setting unequivocally.

8.3. Central/Western Kaoko zones, Epupa Complex, Kunene Anorthosite Complex

The geotectonic setting during the emplacement of Mesoproterozoic igneous rocks in northern Namibia and Angola is not yet well constrained. The AMCG–Kunene Complex is coeval to the arc-magmatism in the AHS-terrane. However, the Kunene Complex intrudes the Paleoproterozoic basement of the Congo craton and has been taken for a pristine extensional igneous complex (Morais et al., 1998), although AMCG magmatism commonly marks post-collisional settings. Granulites exposed in the Epupa Complex display a clockwise P–T–t path and yielded Pb–Pb stepwise leaching TIMS peak-metamorphic garnet and retrograde sapphirine ages of ≈1490–1447 Ma

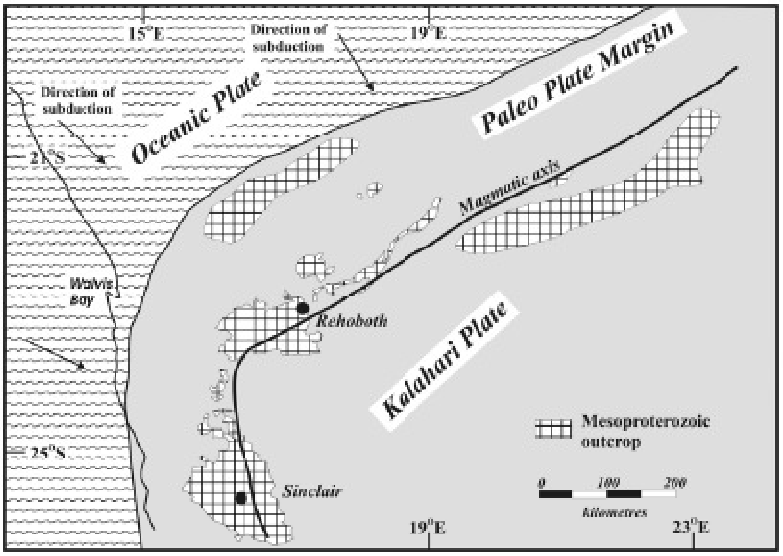


Fig. 15. Interpretative position (present-day coordinates) of the Mesoproterozoic magmatic arc and convergence direction in Namibia, compatible with the data discussed in the paper (modified after Watters, 1976).

(Seth et al., 2001, 2003; Brandt et al., 2003). This is close to the age of migmatised tonalitic rocks from the WKZ, which yielded U–Pb single zircon dates of 1510–1490 Ma (Kröner, 2002). Therefore, a plate-convergence process is likely. The absence of Mesoproterozoic arc igneous rocks older than 1.4 Ga in the AHS and Rehoboth terranes suggest that the convergence between the Congo and Kalahari cratons postdate 1.49 Ga.

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