COMMUNICATIONS OF THE GEOLOGICAL SURVEY OF NAMIBIA

VOLUME 17
2016

MINISTRY OF MINES AND ENERGY

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VOLUME 17
2016

Obtainable from the Geological Survey
Private Bag 13297, Windhoek, Namibia

ISSN 1026-2954

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VOLUME 17
2016

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PAPERS

Cover Image : The cascade tufa at Otjikondavirongo, Kaokoland, Northern Namibia, showing the basal part of the tufa cliff, the actively accreting bryophyte curtain, the algae-filled pool at its base and the cave behind the curtain. The Herero place name signifies "Place beyond Places," with the sense of "The Outback", "The Back of Beyond" or "The Middle of Nowhere".

The Hakos-Rostock Nappe Complex: a case study of alpine-type thrust tectonics within the Southern Margin Zone of the Damara Orogen, Namibia

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Abstract: The Southern Margin Zone (SMZ) of the Pan-African Damara Orogen in Namibia constitutes an alpino-type fold and thrust belt in which Neoproterozoic Damara rocks and the Meso- to Palaeozoic basement were tectonically stacked by the collision of the Kalahari and Congo cratons.

The Hakos recumbent fold, with more than 100 km of hinge line, is the most spectacular structure of the SMZ; it forms part of a nappe pile that can be subdivided into the lower parautochthonous to allochthonous Rostock Nappe Complex (RNC) and the upper, allochthonous Hakos Nappe Complex (HNC). Detached Basement and Damara rocks are stacked in the basal RNC whereas the HNC is composed of Neoproterozoic passive margin sediments previously subdivided into the older Kudis and younger Vaalgras Subgroups of the Swakop Group. However, structural analysis and geological mapping demonstrate the presence of large scale of thrusts that delimit individual units. Therefore, due to the absence of absolute time markers the current stratigraphic models remain poorly constrained.

Three stages of deformation D1-D3 have been identified and are attributed to regional change in transport direction and tectonic style. A predominant shallow-dipping, bedding-parallel S_{0X}-foliation developed under amphibolite facies metamorphism of Barrovian type but may be diachronous in time. In the basal nappes it is related to D1 and carries a sub horizontal NE-SW stretching lineation L1. Overall top-to-the-NE-directed transport with a marked right-lateral component is suggested by asymmetric pressure shadows around L1 stretched clasts and sc-fabrics in granitic orthogneiss.

D2 marks a drastic change in transport direction towards the SE. Large-scale folds evolved together with shallow NW dipping thrusts. Three phases of D_2-folding vary from early isoclinal through tight to late open folds and are related to the development of distinct regional and local schistosities $S_{2,1}$ to $S_{2,3}$ which generally obscure older S-fabrics. A NE plunging stretching lineation is usually developed parallel to the fold axis. SE-directed D_2-thrusting occurs all over the study area but is focused at the contacts between individual nappes with displacement attaining crustal-scale. Here, stretching lineations and fold axes turn into a NW-plunging down-dip orientation.

Late-stage east-west compression D3 resulted in the local development of strike slip cleavage and gentle folding around N-S trending axes resulting in dome and basin interference structures that are visible at the scale of the geological map.

At the orogenic scale the D1 event records NE-directed oblique thrusting of the Congo Craton over the Kalahari Craton. The first phase of nappe emplacement which took place under a right-lateral transpression regime, however, is largely obscured by subsequent deformation. The change in D2 transport towards the southeast is attributed to progressive shortening (or blocking of the D1 thrust zones) leading to the escape of rocks towards the SE, perpendicular to D1. Complementary movements are recorded in the Gariep Belt and suggest overall movement of the Kalahari Craton towards the west; the plate motion post-dates the main phase of tectono-metamorphism in the Kaoko Belt that resulted from collision of the Congo Craton with the South American one. It is therefore assumed that the Kalahari Craton remained as an independent plate before the Pan-African Damara event, separated by a wide ocean from both the Congo and South American cratons.

Key Words: Kalahari; Damara Orogen; Southern Margin Zone; Nappe Complex; Stratigraphy.

To cite this paper: Becker, T. & Ledru, P. 2016. The Hakos-Rostock Nappe Complex; a case study of alpine-type thrust tectonics within the Southern Margin Zone of the Damara Orogen, Namibia. *Communications of the Geological Survey of Namibia*, **17**, 17-50.

Submitted in 2015

In memory of Hans Ahrendt, Franz Boehme and Klaus Weber

Introduction

The Neoproterozoic Damara Orogen (DO) in Namibia (Fig. 1) is considered to be an orogenic belt that was created by the collision of the Kalahari and the Congo cratons (Martin & Eder, 1983; Miller, 1983; Miller *et al*. 2009). Since the pioneer work of Martin & Walliser (1974) the DO has been compared to the Variscan Belt in Europe that is similarly characterised by a foreland progressively involved in the collision, a margin zone with large overthrust structures synchronous with Barrovian type metamorphism, and a central zone marked by migmatisation and granitisation of the continental crust.

A pertinent problem in the reconstructing of past orogens is to establish the link between observed microstructures and metamorphic assemblages, mesoscopic to megascopic structures such as thrust zones or folds belt and the overall geodynamics, i.e. the kinematics of the main plates involved in the collision. This problem is highlighted in the case of the DO where even the definition of the main tectonic units remains a matter of debate. The subdivision into various zones is based on distinct rock assemblages and/or structural elements within each domain; the boundaries being either of a tectonic nature and/or marked by the sudden change in sedimentary and/or igneous facies (Fig. 1, Martin, 1965; SACS, 1980; Miller, 1983; Hoffmann, 1983, 1989).

However, while Hartnady (1978), Weber & Ahrendt (1983), Coward (1983), Miller (1983), Pfurr *et al*. (1987) and Pfurr (1990) characterise the Southern Margin Zone (SMZ) as alpine-type thrusting, De Waal (1966), SACS (1980), Weber (1994), Vietor (1999), and Walter (1999) denied major displacements thus implying that reconstruction of the former stratigraphy is possible.

Here, we give a detailed description of the variation in lithofacies and the complex structural record preserved in the Hakos area of the SMZ where excellent outcrops allow the analysis of the rocks at all scales and both along strike and in cross-section. This serves not only to demonstrate the relation between locally observed structures and processes at plate-tectonic scale but also to show the limits in the reconstruction of the stratigraphy in highly deformed Precambrian domains. The study is based on two years of mapping of four 1: 50,000 maps complemented by thin section petrography, microstructural analysis and the review of published and unpublished literature on the geology of the study area.

The Southern Margin of the Damara Orogen

The ca 200 km wide zone marks the structural transition from the weakly deformed Southern Foreland (SF), with the Naukluft Nappe Complex (NNC) therein, through the South Damara Shear Zone (Coward, 1983, Garoeb *et al*. 2002) into the highly deformed Southern Margin Zone (SMZ) (Fig. 2).

The oldest Palaeoproterozic (~ 2.2 -1.8 Ga) and Mesoproterozoic (~1.3-1.06 Ga) rocks are exposed in several inliers both within the SMZ and SF (Table 1). In the SMZ, they are grouped into the allochthonous Hohewarte Metamorphic Complex whereas the southern and western occurrences have been combined into the autochthonous Rehoboth Basement Inlier (RBI). The detailed geological description of these pre-Damara units is given by Becker & Schalk (2008 a, 2008b).

The overlying Neoproterozoic rocks have been combined into the Damara Supergroup. The regional correlation of its stratigraphy in the Damara and Gariep orogenic belts is based on similarities in sedimentary successions and a number of marker units related to global glaciations (Hoffmann, 1989; Table 1).

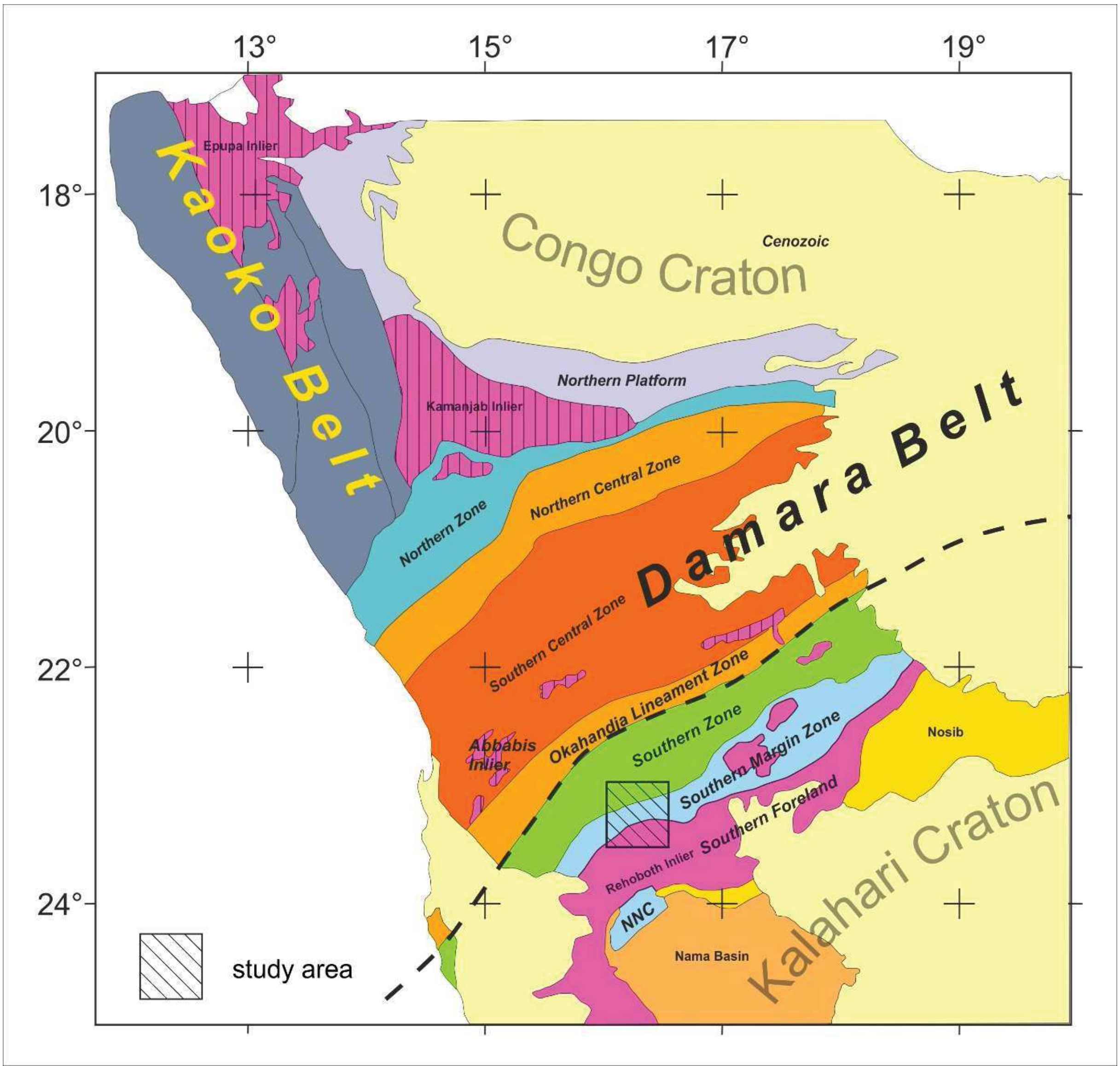

Figure 1. Simplified tectono-metamorphic map of Namibia illustrating the position of the Kalahari and Congo cratons with the Damara Mobile Belt in-between (supposed boundary marked by stippled line). Tectono-metamorphic subunits of the Kaoko belt are merged.

Early Neoproterozoic sedimentation, possibly related to the break-up of Rodinia, started with continental red bed sediments of the Tsumis and Nosib groups, the latter of which have been mapped both in the SF and SMZ. The overlying marine sediments, recording the transition of the sedimentation environment into a passive continental margin, have been subdivided into the allochthonous Hakos Group and the autochthonous Witvlei Group. They comprise glaciogenic diamictite, carbonate rocks, and various types of meta-turbidite and mass flows documenting two stages of global Sturtian (~720 Ma) and Marinoan (~630 Ma) glaciation followed by marine transgression. Late Neoproterozoic to Cambrian shallow marine sediments deposited during the Pan-African collision and uplift are preserved only in the SF where they constitute the Nama Group. These sediments mark the final molasse stage of the Damara cycle.

Hakos Terrane

	Group	Subgroup	Formation
Neoproterozoic	Hakos	Vaalgras	Gomab
			Haris
			Mahonda
			Melrose / Samara
			Naos
		Kudis	Hakosberg / Auas
			Blaukrans / Waldburg
	Nosib		Duruchaus
			Kamtsas

Swakop Terrane

Group	Subgroup	Formation
Swakop	Khomas	Kuiseb

Rehoboth Basement Inlier

	Group	Formation	Intrusive rocks
Mesoproterozoic			Gamsberg Suite
Palaeoproterozoic	Rehoboth	Elim Gaub Valley	Piksteel Suite Alberta IC Weener Suite

Table 1. Simplified stratigraphic model of the study area (after Hoffmann, 1989).

Description of the tectono-metamorphic domains

The SF is structurally characterised by the regional weak deformation with gentle to open folds and a discrete network of anastomosing shear zones marked by highly deformed mylonite. The metamorphic grade varies from non-metamorphic to very low grade conditions.

The NNC forms an isolated tectonic klippe within the SF. The zircon age of 549 +/- 1 Ma determined on tuff from the base of the underlying, autochthonous Nama Group (Grotzinger *et al*. 1995) constrains the maximum age of tectonic transport. It coincides with K-Ar mica cooling ages determined elsewhere in the central and southern DO (Ahrendt *et al*. 1977) suggesting that late gravity sliding during uplift of the orogen may have been an important mechanism for the emplacement of the nappes (Korn & Martin, 1959).

To the north, the Southern Foreland is juxtaposed with the SMZ along the South Damara Shear Zone. The tectonic style and metamorphism changes drastically across this structure and argue for its interpretation as a crustal-scale basal thrust or decollement (Coward, 1983; Garoeb *et al.* 2002). Parautochthonous to allochthonous rocks of the SMZ have been transported along this thrust to the south and south-east.

Rocks of the SMZ have been highly deformed during several stages of the Damara Orogeny and display one or more schistosities. Refolded low-angle thrust and fold nappes form first order structures and involve both pre-Damara rocks and the Damara Supergroup (DeWaal, 1966; Groote-Biedlingmaier, 1974; Bickle & Coward, 1977; Hartnady, 1978; Sawyer, 1981; Coward, 1983; Hoffmann, 1983; Miller, 1983; Pfurr *et al*. 1987; Pfurr, 1990;

Vietor, 1999). The nappes have been related at the orogenic scale to the SE-directed late phase (K3) of compression (Coward, 1983). The clock-wise metamorphic P/T path reached during M1 Barrovian type peak conditions of 588°C and 10 kb corresponding to a low geothermal gradient of 19.1°C/km (Kasch, 1983). A second distinct thermal event or metamorphism outlasting deformation is indicated by the post-tectonic growth of kyanite, garnet, hornblende and biotite under M2 conditions of 535°C and 7.3 kb, corresponding to a slightly elevated geothermal gradient of 23.8 °C/km.

Based on both local and regional studies various efforts have been made in establishing the main nappes and reconstructing the deformation history of the SMZ. Hoffmann (1983, 1989) subdivided the SMZ into a north-eastern thrust belt that is constituted by three lower nappes of basement rocks and the upper Naos Nappe with a largely preserved stratigraphy. In the north-eastern part of the study area these nappes are in direct contact with underlying autochthonous Palaeo- and Mesoproterozoic basement rocks. According to the author, a parautochthonous Samara-Dagbrek zone is sandwiched further south between these domains and is made of highly deformed Neoproterozoic sediments.

In the south-western extremity of the SMZ, the so-called Rostock area, basement and Damara rocks have been grouped into three nappes that are marked by vertical younging (Pfurr *et al*. 1987; Pfurr, 1990). Five deformation events (D1-D5) have been related to three phases of compression (K1-K3). NE-directed thrusting during K1 produced the regional mylonitic foliation and large-scale isoclinal folds with axes dipping gently to the NW or SE. SE-directed thrusting and imbrications of K2 are believed to document a fundamental change in tectonic transport. D2 of that stage is marked by large-scale closed to isoclinal folds with axes dipping to the NE that during D3 were homo-axially refolded into open folds. K3 has been related to the uplift of the orogen; it is characterised by NW verging back folding producing kink banding, open folds (D4) and regional up doming (D5).

In contrast, Vietor (1999) concluded, after mapping the eastern part of the Hakos structure and the Weissenfels syncline, that E to SE-directed closure of the Khomas trough at about 540 Ma resulted in oblique collision of the Kalahari and Congo cratons characterised by a sinistral component. In his model, the basement relief presents an important factor responsible for the local variation in tectonic movements. Folds with both NE and NW plunging axes were related to only one phase of progressive deformation. Their orientation was controlled by the angle between the direction of thrusting and the geometry of the ramps as well as by the strain rate: In shear zones, linear structures were rotated from NE towards the SW parallel to the direction of tectonic transport. Late stage open ENE-verging large-scale folds and a crenulation cleavage steeply dipping to the west were related to E-W compression.

Results

Revision and new geological mapping of the Gamsberg-Hakos area allows the definition of five major SW-NE trending tectono-stratigraphic domains comprising Palaeo- to Neoproterozoic rocks (Fig. 2).

Firstly, the autochthonous to parautochthonous pre-Damara Basement is exposed in the south-eastern part of the area and consists of volcaniclastic rocks of the Gaub Valley Formation overlain by the greenstone-type Elim Formation (Ledru *et al*. 2004; Becker & Schalk, 2008a). Intrusive suites emplaced into these units range from ultramafic to mafic (Alberta Layered Complex, DeWaal, 1966), tonalitic to granodioritic (Weener Suite; Becker & Schalk, 2008a), granodioritic (Piksteel Suite; Ziegler & Stoessel, 1993) to granitic (Gamsberg Suite, SACS, 1980; Becker & Schalk, 2008b). All rocks are marked by the penetrative Pan-African N-NW dipping schistosity. Palaeoproterozoic tectono-metamorphic fabrics that are documented in the SF are therefore completely overprinted and obscured by the Damara event. SE-directed reverse thrusting along a number of moderate to steep NW dipping shear zones resulted in the imbricate stacking of the basement. The most important of these shear zones is the Areb Mylonite Zone that occurs in the extreme SE of the study area and forms part of the Southern Damaran Shear

Zone (Garoeb *et al*. 2002). It has been interpreted as the floor thrust (decollement) defining the boundary between the SMZ and the SF (Coward, 1983). This domain is not described in the following sections.

Secondly, in the Samara-Dagbreek zone, marine Neoproterozoic rocks overlie the Pre-Damara basement with a generally strongly sheared contact. In less sheared domains, and possibly protected from deformation by a palaeorelief the basal Neoproterozoic stratigraphy appears to be locally preserved.

Thirdly, the Rostock Nappe Complex (RNC) is exposed in the central and western part of the study area. It is comprised of Paleoproterozoic and Mesoproterozoic intrusive and volcani-clastic rocks (Hill, 1975; Pfurr, 1990; Pfurr *et al*. 1991) that are tectonically stacked with basal Damaran rocks. The latter consist of fluviatile quartzite and conglomerate (Nosib Group). The structural position of the unit is interpreted to be parautochthonous to allochthonous since stratigraphic units that constitute the RNC have been recognised in the autochthonous domain, too. However, considerable transports are constrained by the subhorizontal planar fabrics as well as the tectonic stacking of Damaran and Pre-Damaran rocks.

Fourthly, The Hakos Nappe Complex (HNC) overlies the RNC and forms the structurally highest unit of the SMZ. It is constituted entirely of metamorphic Neoproterozoic rocks that vary from carbonate rocks, graphite schist, sandstone and basalts (Kudis Subgroup) to diamictite, carbonate, metapelite, quartzite and schist (Vaalgras Subgroup).

Fifthly, to the NW, the HNC is overthrust by rocks of the Khomas Group consisting of quartz-mica schist intercalated with lenses and layers of calcsilicate, minor graphite schist and amphibolite. The contact also marks the boundary between the Southern Margin Zone and the Southern Zone of the DO, and may represent the start of a suture. This hypothesis is supported by the presence of intensely deformed ultramafic rock slivers (ophiolite?) that have been mapped in a NE-SW strip about 5 km north-west of that contact defining the Schlesien line (Hartnady, 1978; Barnes & Sawyer, 1980) or Uis Pass lineament (Hoffmann, 1983). Massive amphibolite and metagabbro of the Khomas group in the extreme northwest of the study area are distinguished as Matchless Member of 0.5 to 3 km thickness. They show geochemical affinity with MORB (Schmidt & Wedepohl, 1983) and are related spatially and genetically with a number of VMS deposits supporting the interpretation as an ophiolite sequence. Again, this domain is not described in detail.

The main structures that have been mapped within the study area and which are referred to in the text, comprise (Fig. 2):

1) Five crustal scale shear zones that from NW to SE are called the Horosib, Berghof, Blaukrantz, Naos and Corona thrusts

2) The Hakos Anticline which is a D2 recumbent fold of about 10 km amplitude with limbs of more than 100 km hinge length,

Intensely folded and thrust rocks in the front of the Hakos structure form a stack of four major synclines marked by an overturned NW and normal SE limb. From NW to SE these are the Klein Chausib, Rooisand, Aris and Djab synclines.

The petrography of the individual rock units together with the description of the contacts is given in the following paragraphs from the base to the top of the tectonic pile. The description of tectono-metamorphic fabrics and the results of structural analysis are presented in the subsequent paragraphs. The geological map and a synthetic NW-SE cross-section of the area are shown in Figure 3 and 4, whereas Figure 5 illustrates the composition of the nappes and their contacts with the adjacent under- and overlying units. The “making of” the geology of the study area is presented in the electronic appendix 1, which allows visualising separately all geological units described thereafter.

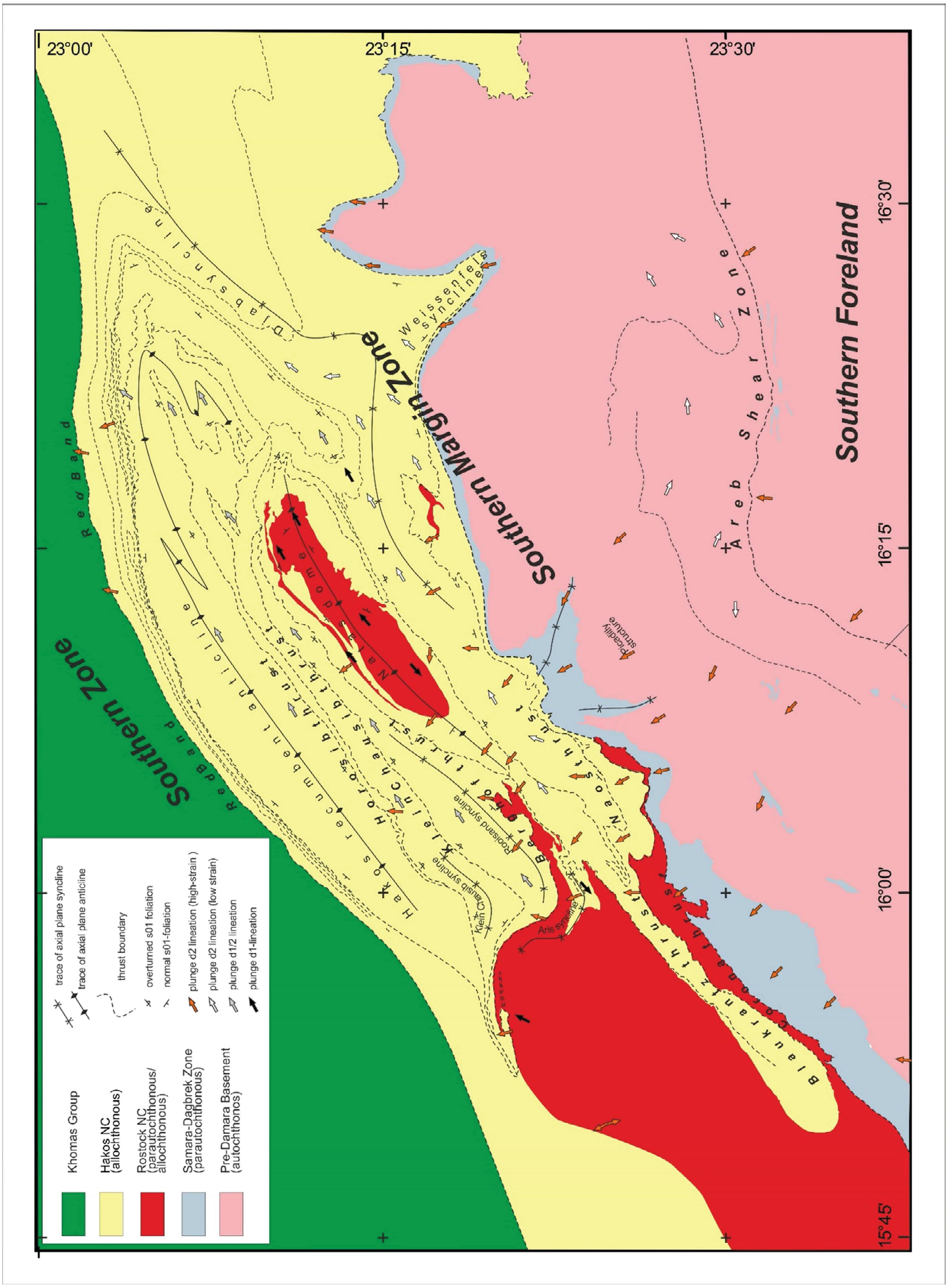

Figure 2. Structural map of the study area illustrating the main folds and thrusts referred to in the text.

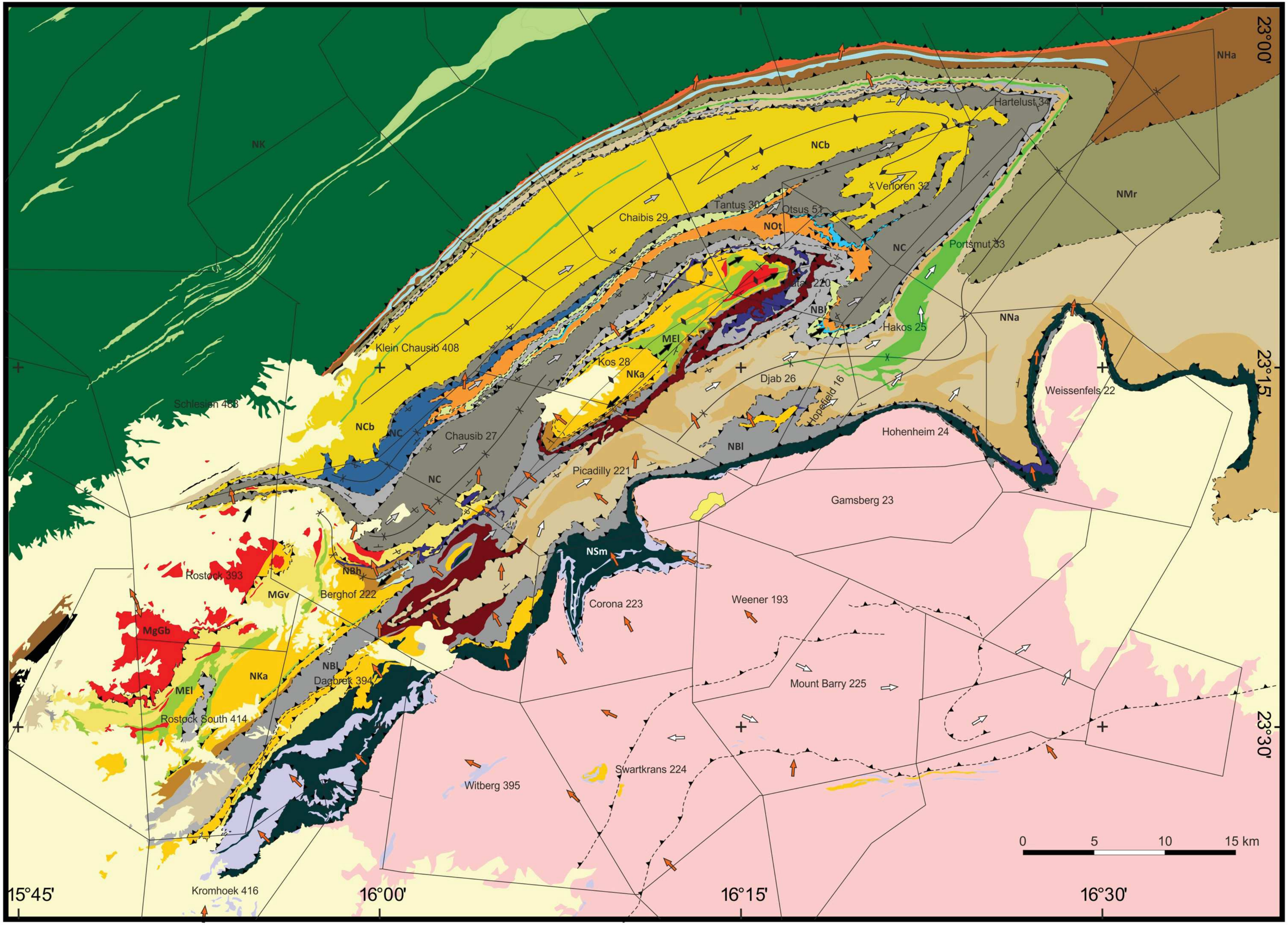

Figure 3. Geological map of the Hakos area

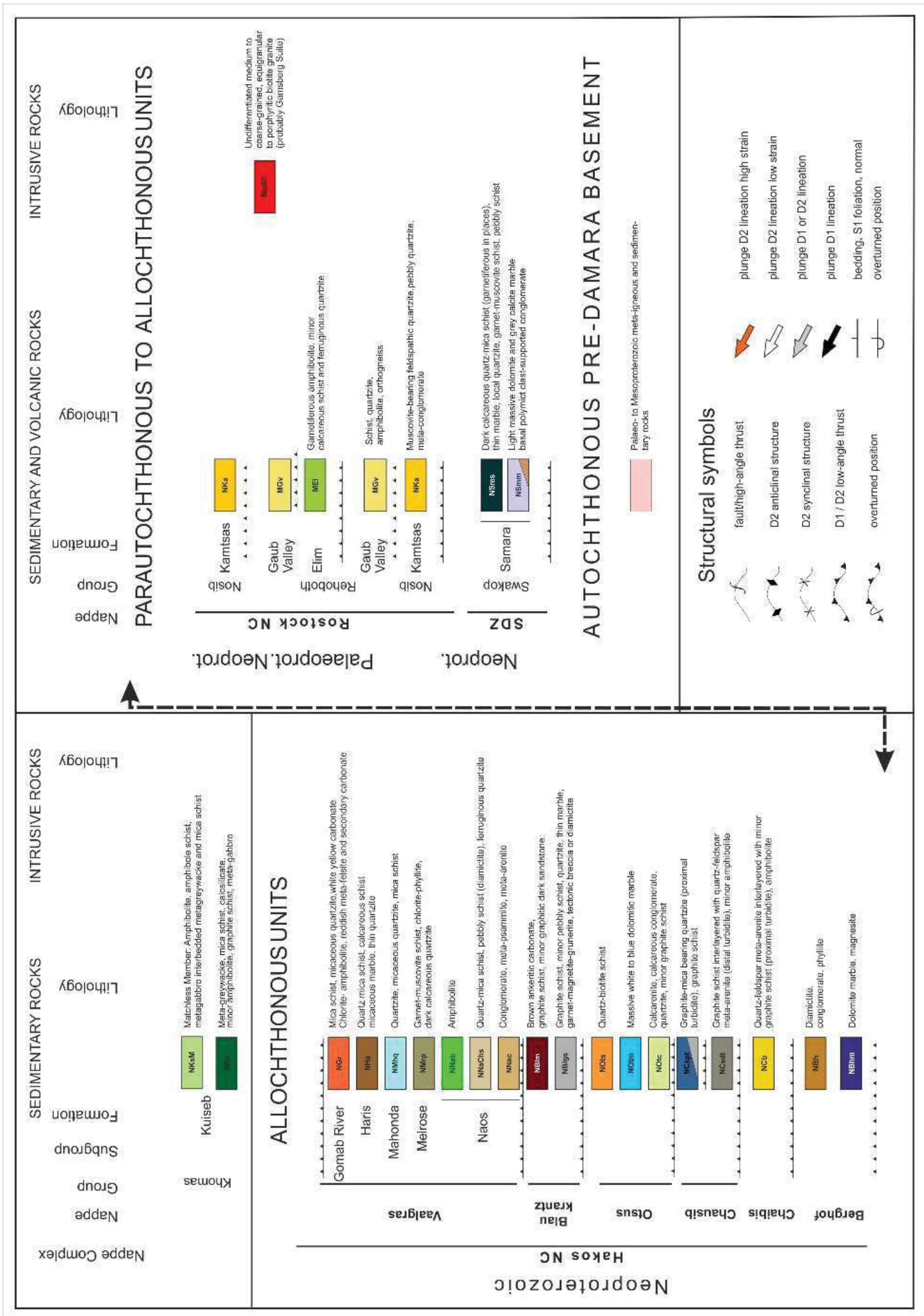

Figure 3. Geological map legend of the Hakos study area.

Samara-Dagbreek Zone (SDZ)

This domain consists essentially of isoclinally folded and thrust imbricated platform carbonates and metapelites of the Samara Formation of the Kudis Subgroup (Hoffmann, 1983). North of the Gamsberg, the SDZ forms a narrow zone only a few 100 metres wide and further east is completely obscured below higher nappes that here are in contact with the parautochthonous foreland. The domain widens SW of the Gamsberg to up to 10 km (Figs 2, 3) and 15 km SSW of the study area carbonate rocks are described in sedimentary contact with underlying Palaeoproterozoic basement (Nagel, 1995). Shearing of the unit therefore appears to become less important from NE to SW arguing for interpretation of the SDZ as a parautochthonous domain (Hoffmann, 1983). This is in contrast to Martin (1983) who considered the SDZ as the root zone of the Naukluft Nappe Complex.

The Samara Formation has been interpreted as a shallow marine platform sequence (Martin, 1983) that records transgression after global Neoproterozoic glaciation (Hoffmann, 1989). The least deformed succession occurs in the study area on Corona 224 in the hinge of the Corona syncline. A basal polymict conglomerate, generally only a few metres thick, is supported by clasts of local provenance; laterally it changes into gritty quartzite or is sheared off. At the western flank of the Gamsberg, a local fanglomerate is made exclusively of Gamsberg Granite with blocks up to house-size. Carbonate rocks overlie this conglomerate and basement rocks along a commonly strongly sheared contact which in places turns into an ultramylonite; they vary from finely laminated blue micritic dolomite to massive white dolomite marble containing minor proportions of quartz, tremolite, talc and mica. These are overlain on Kromhoek 416 along a sharp contact by a sequence of finely-bedded micaceous calcarenite. Blue micritic dolomite up to 1 m thick forms the top of the carbonate sequence and is overlain by poorly layered calcareous garnetiferous phyllite and schist that carries sporadic lone clasts up to 1m size.

The structural record differs from the overlying nappes by the presence of a penetrative NW-plunging D2 down-dip lineation which elsewhere is limited to domains of high shear. Vietor (1999) attributed the contrast in tectonic style to the closeness of the SDZ and the basement. High D2-strain rates near the contact resulted in the rotation of the lineations into the direction of transport.

Rostock Nappe Complex (RNC)

The RNC was first recognised by Hill (1975) in the type locality on Farm Rostock. The so-called Rostock inlier was later described in detail by Pfurr *et al*. (1987) who distinguished three nappes. The new geological map demonstrates the continuation of the complex towards the NE on farms Kos 28, Tantus 30 and Natas 220 where it forms a second tectonic inlier, the Natas Dome of 20 km length and up to 5 km width (Figs 2, 3). A number of smaller tectonic windows within the HNC link these major occurrences and demonstrate the continuation of the RNC below the HNC.

The observed contacts between the RNC and the adjacent rock units are illustrated in Figure 5: The RNC generally underlies the HNC and both units have been thrust over the SDZ. This is demonstrated on Dagbrek 394 and Berghof 222 where, along the SE limb of the Djab syncline the RNC structurally overlies the SDZ. Highly-sheared slivers of the RNC, from less than 100 m to several km in length have been mapped along this contact up to the Weissenfels Syncline and document the importance of the thrust boundary. Late D2-thrusting, postdating the tectonic emplacement of the HNC is demonstrated on Berghof 222 and Kos 28 where rocks of the RNC have been locally emplaced over the Blaukrans Nappe of the HNC.

The RNC consists of Palaeo- and Mesoproterozoic intrusive and volcano-clastic rocks that are tectonically stacked with basal Damaran rocks (Hill, 1975; Pfurr *et al*. 1987, 1991; Pfurr, 1990). In the Rostock Inlier, the oldest supercrustal rocks have been grouped into the Rostock Formation; they comprise sericite-quartz-feldspar schist with varying proportions of garnet and biotite that alternates with garnet amphibolite and minor proportions of metaconglomerate. The series has been interpreted as a bimodal volcano-clastic succession whose deposition age has been dated between 1747-1793 Ma (Nagel *et al*. 1996). The similarity in rock facies and age argues for correlation with the Gaub Valley Formation of the autochthonous domain. In the Natas Dome these rocks are limited to a small strip on the NW flank where they are imbricated with rocks of the HNC.

A distinct succession up to 200 m thickness in the upper part of the Rostock Formation is chiefly comprised of garnet-bearing amphibolite, amphibole schist with minor albite-biotite schist, garnet-feldspar gneiss, garnet-tremolite-tourmaline calcsilicate, ferruginous and magnetite quartzite (or chert), amphibole-chlorite-mica schist, garnet-chlorite phyllite as well as thin intercalations of brown dolomite. This so-called black band (Pfurr *et al*. 1991) is now grouped, because of similarity in rock types, into the Elim Formation. In the Natas Dome, the unit decreases over 15 km along strike from more than 100 m in the north-eastern corner of the dome to a small boudinaged ultramylonitic horizon less than 10 cm thick that is stacked between Neoproterozoic rocks, hence, documenting involvement of the basement in crustal-scale transports.

Intrusive rocks in both inliers include gneissic gabbro, tonalite and hornblende-bearing granodiorite of the Late Palaeoproterozoic (~ 1770 Ma) Weener Suite (Becker and Schalk, 2008a) whereas gneissic granite forms part of the Mesoproterozoic (1207 - 1084 Ma) Gamsberg Suite (Pfurr, 1990; Pfurr *et al*. 1991; Becker & Schalk, 2008b). Medium- to coarse-grained augen-gneiss in the Natas Dome consists of biotite-plagioclase-K-feldspar-quartz with minor secondary muscovite, chlorite and calcite. In zones of high strain the orthogneiss turns into a fine- to medium grained mylonitic L-tectonite.

The granite intruded amphibolite of the Elim Formation close to the contact altered into biotite schist and intruded by veins of Mo-W-Cu-Au mineralised pegmatite. The scheelite of this pegmatite has been exploited in the Natas mine (Reuning, 1925).

The Meso- and Palaeoproterozoic rocks are overlain by and locally imbricated with the Neoproterozoic Nosib Group, which consists of quartzite and local polymict, clast- to matrix-supported conglomerate that in the NE corner of the Natas Dome is marked by normal grading. Here the clast composition is dominated by granite, up to 50 cm in diameter whereas quartzite, vein quartz and amphibolite are subordinate. Laterally, the conglomerate grades into gritty sandstone. Sedimentary structures observed in the quartzite include locally observed large-scale low-angle cross-bedding arguing for aeolian deposition and in other places channel fills of conglomerate suggesting a fluviatile depositional environment. Sericite-bearing meta-arenite, schist and conglomerate in the south-eastern corner of the Natas Dome are now included into the Nosib Group. These rocks have been formerly attributed to the Gaub Valley Formation. However, the lack of amphibolite dykes and felsic intrusions that elsewhere are characteristic for the Gaub Valley Formation argue for the younger age of these sediments.

Hakos Nappe Complex (HNC)

The Hakos Nappe Complex has been subdivided into six major nappes marked by distinct rock compositions; these are from the base to the top the Berghof, Chaibis, Chausib, Otsus, Blaukrantz and Naos nappes (Figs 2, 3). The boundaries between adjacent nappes are generally characterised by one or more of the following features:

1) presence of major shear zones marked variously by the development of SC-fabrics, progressive rotation of fold axes into the direction of transport, mylonitic fabrics, occurrence of boudinaged layers of magnesite and talc, development of a tectonic melange, ramp structures, quartz-kyanite veins

2) change in sedimentary facies

3) stratigraphic inversion

4) sudden change in structural style.

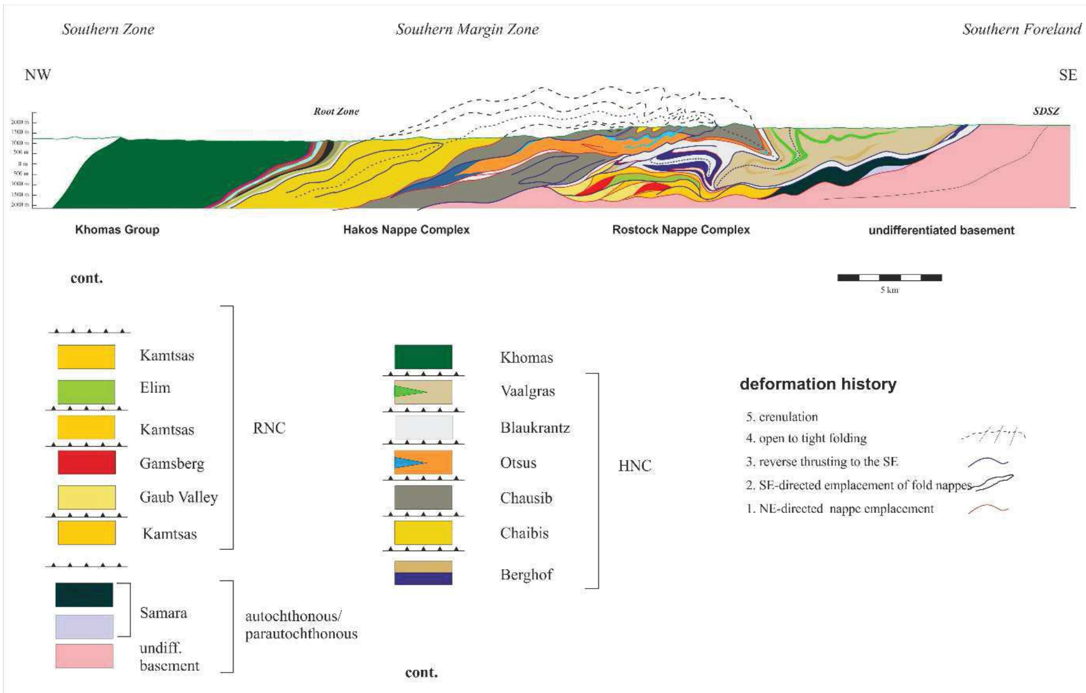

Figure 4. Synthetic cross-section and tectonic pile as derived from geological mapping.

Berghof Nappe

The Berghof Nappe remains a poorly known stratigraphic or tectonic unit whose facies have been first described by Miller (2008) as the Berghof Formation. The main rock types comprise carbonate rocks, conglomerate, phyllite and diamictite (Walter, 1999; Miller, 2008).

The facies are correlated with the Chuos Formation elsewhere implying glaciogenic sedimentation within the worldwide Sturtian Event. In contrast, we classify the unit as an allochthonous nappe of the HNC because of the highly sheared contact with underlying rocks marked by the presence of tectonic slivers of marble, schist and conglomerate and, in the Natas Dome by crystallisation of kyanite in the shear zone.

In the type locality on Berghof 222, the unit is exposed in the core of the Aris Syncline and tectonically stacked between underlying Nosib Group in the south, and overlying rocks of the HNC in the east being juxtaposed by the Blaukrans Thrust (Figs 2, 3, 4). In the north, the Berghof Thrust limits the Berghof Nappe from overlying Palaeo- and Mesoproterozoic rocks. 3 km north of that contact, subhorizontal overturned and highly sheared sediments of the Berghof Nappe form a tectonic window below Palaeoproterozoic basement thus demonstrating the crustal scale of the Berghof Thrust. Discontinuous layers and lenses of the Berghof Formation have been mapped around the Natas Dome, many of them only a few metres thick and always in thrust contact with adjacent rocks from both the RNC and the HNC. Syn- to post-tectonic kyanite-muscovite-quartz veins in the thrusts attest to hp/lt metamorphism during shearing. On the south-eastern flank of the Natas Dome, rocks classified into the Berghof Nappe form the core of a recumbent SE-plunging anticline (Figs 2, 3, 5) and are enveloped by rocks of the HNC.

The description of the reference section in the type locality Berghof 222 is given by Walter (1999). Here, the Berghof Nappe is folded into the SE-facing Aris Syncline, which is bounded on both sides by southward-directed thrusts (Fig. 2, 3). Lithological and structural asymmetries, related to structural complexities occur across the structure. On the northern, overturned limb, the Berghof Nappe is in tectonic contact with overlying Palaeoproterozoic basement and starts with a horizon, up to 10 m thick of clast-supported, non-graded polymict conglomerate. The elongated clasts of quartzite, gneiss and amphibolite are supported by the schistose quartz-feldspar-biotite matrix.

On the southern limb, the Berghof Formation overlies the Nosib Group along a sheared contact. Here, the base is marked by the sudden appearance of tectonic slivers and boudins of white marble, up to 1 m thick. The basal successions are overlain on both limbs by up to 50 m thick calcareous pyrite-rich phyllite. However, silicified brown porous dolomite above the phyllite has been mapped only at the northern limb and is probably related to hydrothermal brines moving along another thrust. Chert, talc, pyrite altered to hematite, as well as calcite, dolomite and quartz all occur. Crystals of all minerals up to 20 cm in size are found in ubiquitous vugs and veins.

These strongly altered rocks are overlain by light to dark grey, fine to very coarse-grained dolomitic marble that forms layers and boudins from 20 cm to 30 m thick. Small slivers of the dolomite have been mapped along strike for more than 40 km up to Natas 220. Here, they reach considerable thickness and are stacked between rocks of the HNC that in the vicinity of the contact turn into ultramylonite.

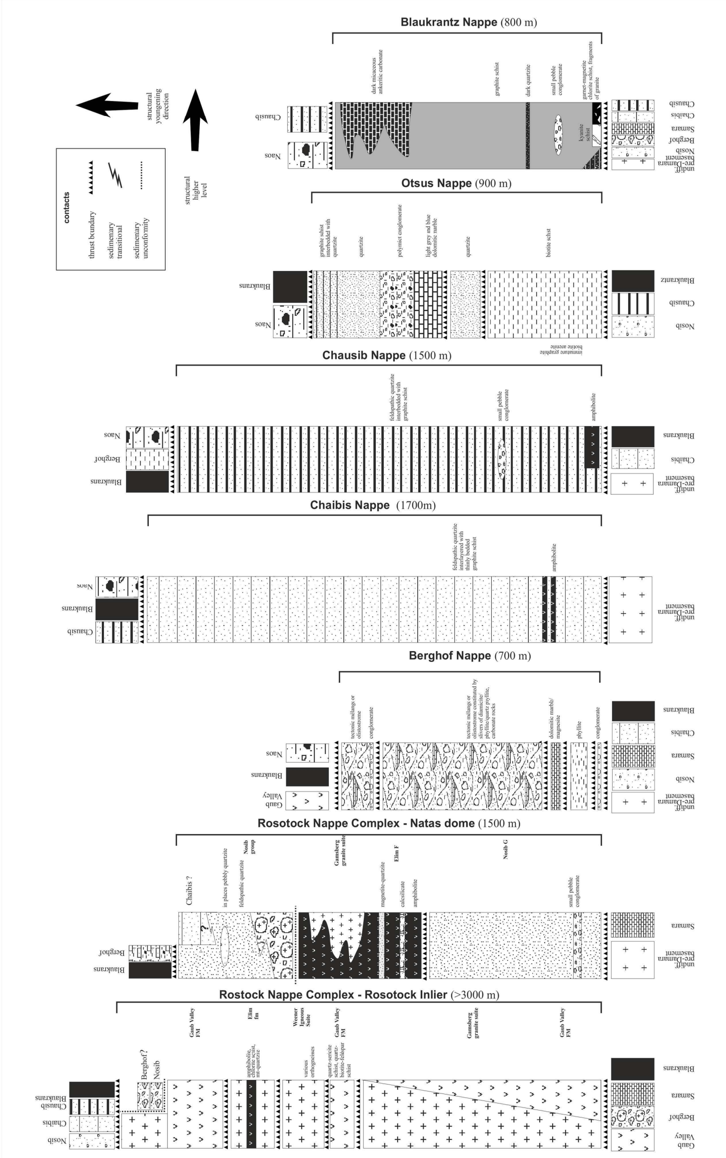

Figure 5. Composition and contact relationships of the Rostock and Hakos nappe complexes. Not to scale. The maximum tectonic thickness is given in brackets. The Naos Nappe is not shown because of its simple structural relationships to the underlying units.

The dolomitic marble is overlain by polymict conglomerate up to 20 m thick; the rock is characterised by clasts of quartzite, orthogneiss and soft pebbles, up to 1 m size and the finely-laminated sandy quartz-feldspar-mica matrix. The interpretation of the rock as glaciogenic diamictite (Walter, 1999) is doubtful because penetrative shear has overprinted the sedimentary structures and does not allow the classification of the rock as a matrix-supported conglomerate: careful observation shows that soft pebbles have been metamorphosed into quartz-sericite schist. The clasts are often extremely sheared and flattened and therefore appear to be part of the matrix.

A tectonic or sedimentary mixed series of rocks, about 500 m thick overlies the conglomerate. It is constituted by various rock types that form lenses and layers of generally limited lateral extent and often only a few metres in thickness. The main rock types are clast and matrix supported conglomerate or diamictite, which differ from the underlying conglomerate by the presence of carbonate clasts. Of minor proportions are intercalated phyllite, quartz phyllite, and various types of carbonate rocks, the latter forming lenses and layers up to 2 m thick and 6 km long. The mixed series has been interpreted as an olistostrome that resulted from submarine gravity sliding and slumping of unconsolidated sediments with a glaciogenic origin (Walter, 1999). This interpretation is challenged because carbonate rocks rarely form part of a glaciogenic succession. Instead the interpretation of the unit as a tectonic melange of platform and slope facies is favoured. Structural repetition is indicated by a second series of conglomerate and mixed series that form the top of the Berghof Formation.

A lateral transition has been described from the Berghof Nappe into overlying graphite schist of the Blaukrantz Formation which at the base contains isolated clasts and layers of conglomerate interpreted as drop stones and diamictite of glaciogenic origin (Walter, 1999). However, the geological map by the same author clearly demonstrates an unconformity between the two units. Tectonic slivers of talc schist at the subhorizontal contact, up to 25 m thick are probably derived from dolomite whereas rocks in the vicinity of the contact are often silicified. This reveals the activity of former hydrothermal brines that have been released during tectonic transport that is also documented by scc' fabrics in graphite schist near the contact (see below).

Chaibis Nappe

The Chaibis and Chausib nappes have been formerly classified as members of the Hakos (Hoffmann, 1983, 1989) or Hakosberg Formation (Miller, 2008). The sediments are interpreted as middle and outer fan facies of a deep-water, passive margin turbidite system (Hoffmann, 1983).

The Chaibis Nappe forms the core of the Hakos Fold Nappe (Figs 2-4). It is exposed over an area of about 200 km^2 with the tectonic thickness attaining up to 2500 m. Figure 5 illustrates the contacts with adjacent lithologic units. In its south-western part on farm Schlesien 483, the Chaibis Nappe has been transported along the Horosib Thrust over pre-Damaran Basement of the RNC (Pfurr *et al*. 1987). A tectonic mélange a few metres thick is locally developed along the contact and consists of mylonitic yellow quartzite studded by centimetric graphite flakes. Boudins of magnesite derived from dolomite also form part of this zone. The rocks of the RNC adjacent to the contact are contorted around unoriented fold axes and stretching lineations vary widely in plunge from a north-westerly to south-easterly direction. Already several hundred metres north of that contact, the Chaibis Nappe is overthrust by the Chausib Nappe. It is in turn thrust off on the northern, normal limb by the higher Blaukrantz and Naos nappes, that here are juxtaposed with the Chaibis Nappe.

The Chaibis Nappe consists almost entirely of massively bedded fine-grained well-sorted K-feldspar-quartz meta-arenite that alternates at m to 10 m-scale with thin graphite schist layers. Except for apparent planar bedding the quartzite lacks any other sedimentary structures. The sedimentary pile has been interpreted as a proximal turbidite sequence (Porada & Wittig, 1976, 1983). An interesting profile at the north-western flank of the Natas Dome (Tantus 30) shows the transition from basal fluviatile arkose and conglomerate of the Nosib group into an overlying sequence of thickly-bedded meta-arenites alternating with thinly-bedded graphite schist which resembles Chaibis Nappe facies. Provided that the contact is sedimentary this would imply the presence of marine sediments with a Nosib age and constrain the upper stratigraphic position of the Chaibis Nappe (Vietor, 1999). However, since this is the only place where such sedimentary transition is suspected, more detailed work is necessary to confirm the observation.

Few amphibolite horizons, up to 20 m thick are intercalated with the metasediments and transposed into the s_{0X} fabric of the adjacent rocks; the most prominent of these former dykes or sills can be traced for more than 25 km strike from Chaibis 29 to Klein Chausib 408 and attains more than 50 m thickness. They record the continuation of crustal extension after deposition of the sediments.

Chausib Nappe

The Chausib Nappe is exposed over about 250 km^2; it envelops the core of the Hakos Fold and further south-east is folded into the closed late D2, SE-facing Klein Chausib and Rooisand synclines characterised by several km wave-lengths (Figs 2-4). The thickness of the nappe varies considerably along strike from a few metres in the south-western extremity of the Hakos Fold where the rocks are transformed into ultramylonite, to about 1900 m in the hinge zone of the structure. The variation results from layer-parallel gliding combined with isoclinal folding at various scales and subsequent tectonic stacking.

The contacts to the neighbouring nappes are illustrated in Figure 5. In general, the Chausib Nappe structurally overlies the Chaibis Nappe around the Hakos Structure and the contact can be traced from the SE overturned limb around the hinge to the NW normal limb where the unit is thrust off by the higher Naos Nappe. Between the Hakos Fold and the Natas Dome, the Chausib Nappe underlies rocks grouped into the Otsus Nappe; all units were subsequently folded into the Klein Chausib Syncline. The south-eastern contact of the Chausib Nappe in this area is delimited by the late Berghof Thrust which juxtaposes the unit in the southwest with underlying rocks of the RNC (Hoffmann, 1983) and along the north-eastern continuation of the thrust with rocks of the Blaukrantz Nappe. Rootless folds and sporadic boudins of exotic orthoquartzite along the contact document the importance of the shear displacement arguing for its tectonic nature. On Schlesien 483, a thrust sliver of Chausib rocks, about 20 m thick and a few hundred metres long, is tectonically stacked between basement rocks of the RNC. Another sliver, about 50 m thick and 500 m long occurs at the northern flank of the Natas Dome where it is stacked between underlying conglomerate of the Nosib Group and overlying graphite schist of the Blaukrantz Nappe. These rocks change along north-eastern strike into a tectonic mélange consisting of ultramylonitic quartzite with graphite schist flakes.

Metasedimentary rocks, comprising fine-grained micaceous sandstone alternating at cm to m-scale with graphite schist, are interpreted as distal turbidites (Porada & Wittig, 1976). Most common are quartz-feldspar sandstone beds up to 1.5 m thick that alternate with up to 30 cm thick layers of graphite schist. Graded bedding from coarse-grained quartz-feldspar sandstone into quartz-mica (+/- kyanite) schist has been observed in a few places. Few lenses and layers of small-pebble conglomerate up to 1 m thick are intercalated within the succession. A general increase in the thickness of individual layers from the stratigraphic bottom to the top of the sequence has been described by Porada & Wittig (1976) but was not confirmed in the present study.

A thrust separates the lower sequence from the overlying succession of rocks interpreted as proximal turbidites: Basal, up to 100 m thick graphite schist with occasional lenses of small pebble conglomerate resembles facies of the Blaukrantz Nappe into which they could be grouped. Higher up, the graphite schist is interlayered with meta-arenite at 3 to 30 cm scale. Local, thinly bedded dark meta-psammite contains varying amounts of graphite and is marked by gradual transition into thin beds of graphite schist. One, up to 10 m thick horizon of mafic volcanic rocks occurs close to the contact with the overlying Chaibis Nappe. It forms a prominent marker that can be traced from Klein Chausib for more than 20 km along the north-eastern strike.

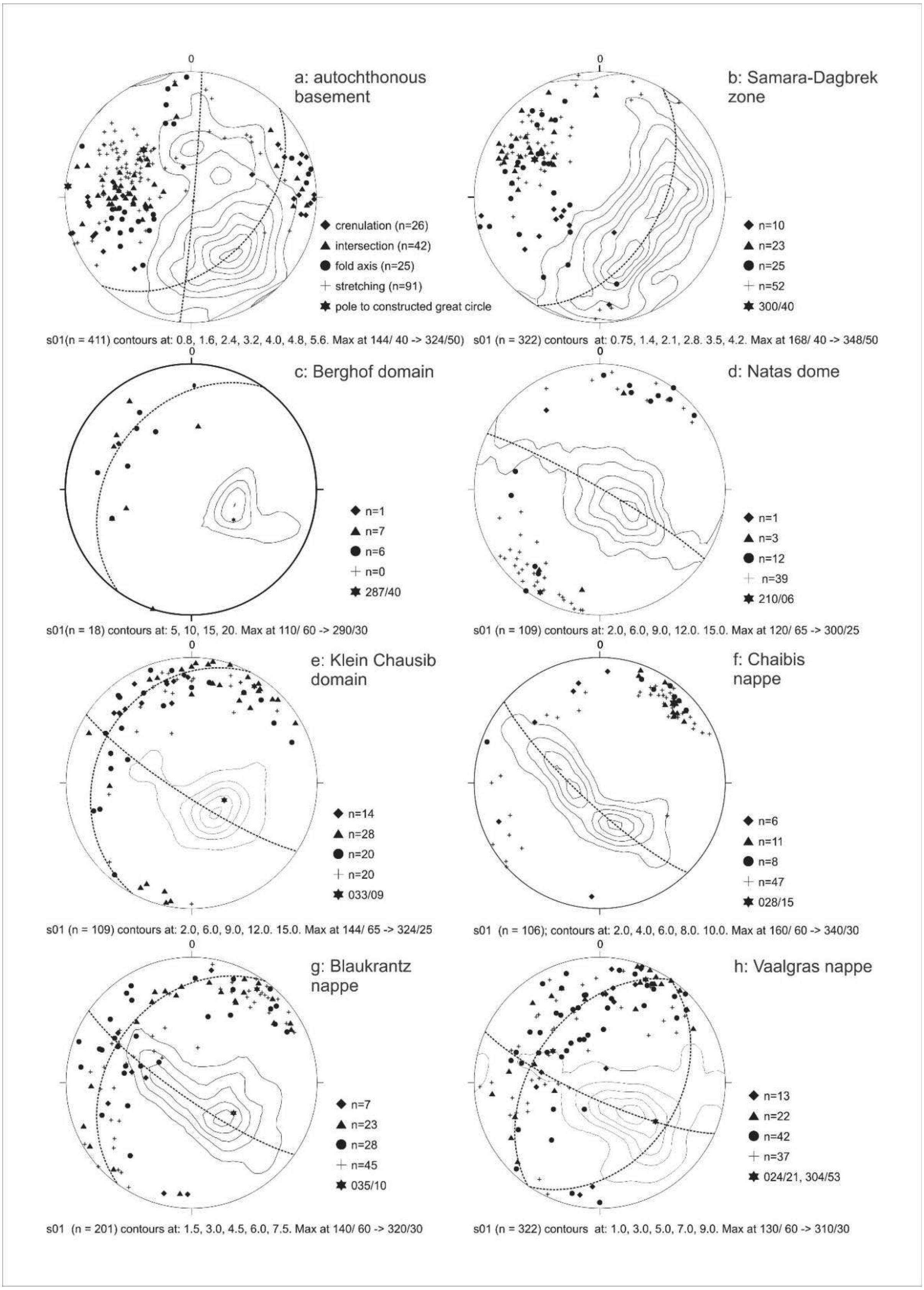

Figure 6. Presentation of structural data of the analysed domains in stereograms.

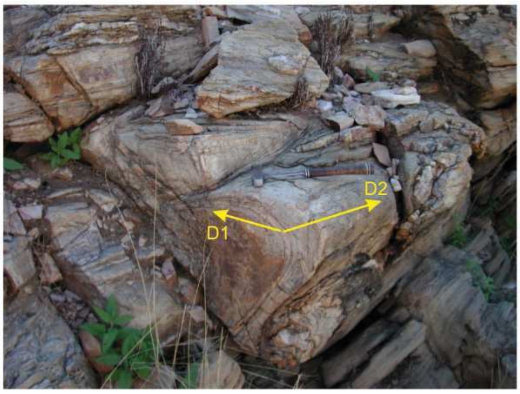

Fig 7a: Interference pattern of D1 (330/15) and D2 (060/10) folds in quartzite of the Nosib G (Berghof 222)

Fig 7b: Complex D1-D2 fold interference pattern; Berghof Fm (Berghof 222)

Fig 7c: Commonly planar orthogneiss turns in vicinity of major D1 shear zones into L-tectonite (Tantus 30)

Fig 7d: Early D1 s-lineations are rotated around D2 axes resutling in a pseudo-sc-fabric (Kos 28)

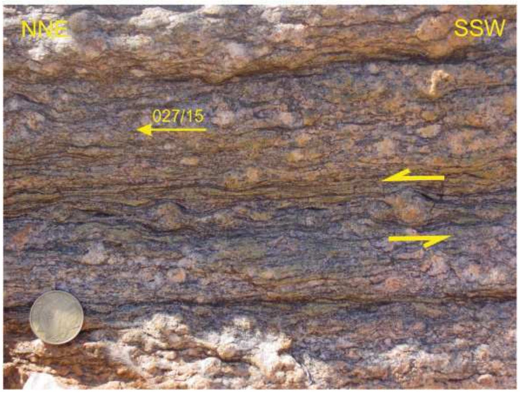

Figure 7e: Top to the NNE shear sense in orthogneiss of the Rostock Nappe Complex

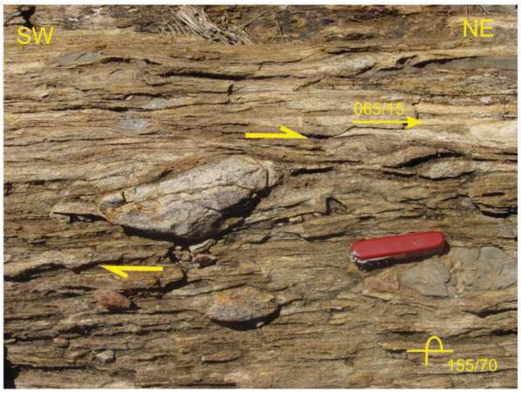

Figure 7f: D1 sigma-clasts are rotated during D2 around NE-plunging fold axes and, hence, indicate at the steeply overturned limb of the Hakos fold a dextral shear sense with the top (i.e. the lower part of the photo) to the SW (Natas 220)

Figure 7. Tectono-metamorphic structures related to the NE-directed D1-deformation.

Otsus Nappe

The Otsus Nappe is here defined and mapped for the first time (Figs 3-5). It consists of quartzite, calcareous conglomerate, polymict conglomerate, blue dolomite, white marble, graphite schist and biotite schist, resembling facies of the Berghof Nappe of which it could be a stratigraphic correlate. However, it is separated from the latter unit because of its stacking at a higher structural level between the Chausib and Blaukrantz nappes. The unit extends from the hinge zone of the Hakos Fold for about 50 km to the south-west where it forms the core of the Klein Chausib Syncline. The maximum tectonic thickness is about 1200 m.

Mapping along strike shows that the Otsus Nappe is internally sheared and hence, profiles across the unit display considerable variation in rock composition and sedimentary successions. The most variable, overturned sequence is preserved in the hinge zone of the Hakos Structure. Here, quartz-biotite schists are thrust over Chausib Quartzite and are in turn tectonically overlain by talc-bearing coarse-grained white marble and blue fine-grained dolomite marble up to 30 m thick. These rocks also occur further north in direct contact with the Chausib Nappe.

The carbonate rocks are overlain by an oligomict breccia or conglomerate of about 40 m thickness that is chiefly made of strongly stretched angular grey dolomite clasts up to 30 cm size that are supported by a sandy to gritty quartz-carbonate matrix. Higher up the breccia changes into a clast-supported polymict conglomerate which is constituted in decreasing order by granite, quartzite and carbonate clasts. Yellow quartzite alternating with thin layers of graphite schist and a few lenses and layers of conglomerate forms the top of the sequence.

These contrasting lithotypes are missing at the overturned limb of the Hakos Fold and instead a monotonous succession of quartz-biotite schist, up to 1000 m thick extends to the south-west for more than 35 km along strike. On Klein Chausib 408 it underlies polymict conglomerate made of granite and dolomite clasts supported by quartz-mica-carbonate matrix that possibly can be correlated with the conglomerate in the hinge zone of the Hakos Fold; thin slivers of carbonate rocks are associated with these rocks.

Blaukrantz Nappe

The Blaukrantz Nappe has been transported over all rocks described before (Fig. 2-4). However, in front of the Hakos Fold the unit was in turn overthrust locally by the Berghof and the Chausib nappes during subsequent deformation. Prolonged deformation history is also recorded in the Weissenfels Syncline where Blaukrantz facies rocks, marked by m-scale ramp structures, are stacked between underlying mylonitic granite (Gamsberg Suite) and overlying magnetite-rich quartz-mica schist (Waldburg Formation of the parautochthonous SDZ). In the eastern corner of the Natas Dome, complicated tectonic stacking is suggested by the succession of rocks that, from the base to the top, are classified as Chaibis, Blaukrantz, Berghof, pre-Damara basement and again Blaukrantz facies. Most of the nappe forms a continuous outcrop, about 70 km^2 in surface area, where the unit varies in thickness from more than 800 m in the core of the Djab Syncline to a few metres at the northern limb of the Hakos Fold (Fig. 3).

On Berghof 222, the Blaukrantz Nappe overlies the Kos Nappe along a basal unit, up to 20 m thick, interpreted as a tectonic melange. It consists of complexly folded granite boudins and quartz-muscovite schist slivers enveloped by graphite schist. Pervasive post-tectonic magnetite-garnet-grunerite mineralisation close to the contact attests to important fluid activity. This horizon can be traced for about 500 m to the east before it is thrust off by overlying graphite schist. Its regional extension is documented by a number of other outcrops along strike where exotic fragments from underlying nappes (white marble, conglomerate and amphibolite) are stacked within graphite schist. Higher up follows quartz-mica-graphite schist, about 180 m thick that in turn is overlain by brown micaceous carbonate, up to 600 m thick. In the west this rock becomes the dominant facies in the upper part of the nappe. Elsewhere, beds of very fine-grained dark quartzite, varying in thickness from cm to dm are rarely intercalated with the graphite schist. Magnetite quartzite or iron rich chert, up to 3 m thick locally forms the top of the Blaukrantz Nappe (Vietor, 1999). The sediments of the Blaukrantz Nappe have been interpreted as deep water (Miller, 1983) facies that mark marine transgression in the wane of Neoproterozoic global (Marinoan?) glaciation (Vietor, 1999).

Metasedimentary rocks of the Blaukrantz Nappe display considerable variation along strike. Strongly foliated, parallel-bedded quartz-mica-graphite schist constitutes the main rock type in the central and eastern outcrops. Where less deformed, fining-upward cycles at cm to dm-scales are recognisable which are marked by increasing graphite and mica and decreasing quartz towards the top of the beds. Lenses and layers of brown micaceous carbonate and calcarenite, up to 100 m long are intercalated with the schist; however, their thickness generally does not exceed a few decimetres. Rock types in the western outcrops are more variable in their composition. On Berghof 222, several horizons and lenses of small-pebble conglomerate, up to 1 m thick are intercalated at the base with graphite schist (Walter, 1999). Laterally, they change into a sequence of yellow meta-arenite alternating with graphite schist that resembles facies of the Chausib Nappe.

Vaalgras Nappe

Rocks of the Vaalgras Nappe overlie the Blaukrantz, Chaibis and Chausib nappes and in the past the contact has been interpreted as a sedimentary unconformity (Hoffmann, 1983, 1989). However, the often completely different geometry of folds across the contact together with the rotation of stretching lineations in the direction of transport, the sporadic occurrence of magnesite boudins and kyanite crystallising in the S_{01}-foliation are evidence for tectonic decoupling during regional thrusting. Hence, this unit is classified here as the highest nappe of the HNC.

It is noteworthy that the Vaalgras Nappe does not occur into the overturned limb of the Hakos Structure, suggesting emplacement during a second phase of SE-directed thrusting concomitant with or postdating the isoclinal folding. Although mapping and structural analysis reveal internal thrusting and polyphase folding, the original stratigraphy appears to be largely preserved (Hoffmann, 1983, 1989).

1) The basal Naos Formation is exposed in one continuous outcrop of about 420 km^2 area in the core of the Djab and Weissenfels synclines and from there continues into the “root zone” at the northern limb of the Hakos Fold (Figs 2-4). In many places, a polymict clast-supported conglomerate of about 30 m thickness marks the onset of sedimentation. It is constituted by well-sorted quartzite, schist and carbonate clasts, 3-6 cm in size supported by the quartz-mica-carbonate matrix. Intercalated discontinuous lenses and layers of quartzite are up to 1 m thick and several tens of metres long. Contorted bands of white marble are part of the basal succession and it is uncertain whether they are highly deformed clasts, sediments or tectonic slivers. Laterally and vertically the conglomerate often grades into thickly bedded, locally pebbly quartzite. Pebbly schist of possibly glaciogenic origin (Hoffmann, 1989) unconformably overlies this basal succession and prevails in the upper part of the Naos Formation. They are intercalated with varying proportions of thickly bedded quartzite, garnet-chlorite phyllite, magnetite quartzite, conglomerate, carbonate and mafic volcanic rocks. The thickest of these metabasaltic flows has been distinguished as the Choaberib Member (Hoffmann, 1989). It forms a marker horizon, up to 150 m thick that can be followed over more than 30 km along strike. The irregular outline of the member documents complex interference folding.

2) The Melrose Formation, up to 900 m thick is constituted by garnet-chlorite phyllite interlayered with minor quartzite interpreted as marine slope facies that stratigraphically correspond to carbonate platform facies of the SDZ (Hoffmann, 1989).

3) Overlying schist and quartzite on the northern limb of the Hakos Fold are distinguished as the Mahonda, Haris and Gomab formations (Hoffmann, 1989). However, two profiles show discontinuity of individual facies along strike; instead intense thrusting and folding results in tectonic stacking of the facies.

4) The Hartelust Member of the Gomab Formation forms the top of the HNC. It is composed of felsic and mafic volcanic rocks, the latter being altered into reddish-stained silicified carbonate-amphibole-chlorite schist. The unit records polyphase tectono-metamorphic deformation and late-stage hydrothermal overprint and brecciation. En-echelon arrangement of vein quartz suggests normal faulting during orogenic uplift.

The contact to the overlying Kuiseb Formation of the Khomas Group defines the boundary between the Southern Margin Zone and the Southern Zone of the Damara Orogen (Miller, 1983), which were later classified as Hakos and Khomas terranes (Hoffmann, 1989) Up to now, this important contact has not been studied in detail, even though it marks drastic changes in sedimentation and tectonic style (Coward, 1983).

Tectono-metamorphic Fabrics : Deformation History

Structural analysis shows that the SMZ has been affected by two major (D1-2) and one minor (D3) phase of deformation. Marked differences exist in the structural record of the lower and upper part of the nappe pile documenting two stages of north-east (D1) and south-east (D2) directed movements, respectively. D1-fabrics are chiefly preserved in the autochthonous basement, and the parautochthonous to allochthonous RNC. In contrast, the D2-event obscured all older metamorphic fabrics in the upper Blaukrantz and Vaalgras nappes, which therefore only record SE-directed transports.

The cross section (Fig. 4), parallel to the D2-transport direction, shows the main D2 fold structures that also characterise the geological map (Figs 2, 3). Supposed D1 and D2 thrusts are marked in red and blue, respectively. The anastomosing network of thrusts shown in the section represents only an idealisation marked by major uncertainties since the observed thrusts have been partly projected over several tens of km from their outcropping position into the plane of the cross-section.

D1: NE-directed nappe emplacement related to regional dextral transpression

The first deformation event D1 is characterised by subhorizontal SW-NE stretching lineation marked by 1) elongated quartz, carbonate, feldspar crystals, 2) trails of mica, 3) elongated quartz-feldspar aggregates, 4) elongated clasts. Such D1 markers are well-recorded around the Natas Dome at the thrust contact between gneissic granite and conglomerate. Commonly planar gneissic fabrics turn in the thrust into L-tectonite (Fig. 7c) while clasts that elsewhere are oblate change into prolate forms. Top to the NE shear sense is indicated by SC-fabrics in orthogneiss of the RNC (Fig. 7e) and, less stringent, asymmetric pressure shadows around clasts of meta-conglomerates. Subsequent re-activation of the thrusts is documented in several places where D_1-lineations are folded around oblique D_2 fold axes (Fig. 7d). Lineations attributed to D1 have been identified in the RNC, the Kos, the Berghof and the Otsus nappes and argue for initial stacking of all nappes up to the Otsus Unit during a first phase of NE directed transport (Fig. 2). The present-day configuration of the nappes with respect to the autochthonous basement implies an important dextral horizontal component since the contact between the two domains strikes subparallel to the inferred transport direction.

Regional shallow-dipping, bedding-parallel mylonitic S_{01}-foliation is developed in both the RNC and the HNC but may be diachronous and polyphase in evolution; it mostly parallels the outline of the geological units (Figs 2, 3) and probably parallels the axial plane of isoclinal folds of both D1 and D2 deformation (Pfurr *et al.* 1987; Pfurr, 1990). Small-scale rootless D_1 isoclinal folds are recorded in the basal part of the tectonic pile within rocks of the RNC (Pfurr *et al*. 1987; Leiss, 1990). Their fold axes plunge shallow to the NW or SE and the folds occasionally form interference patterns with D_2-structures (Figs 7a, b). Large scale NE-verging isoclinal D1 folds of deca-km-scale have been mapped in the western extension of the Samara-Dagbrek zone on Kromhoek 416 (Leiss, 1990).

The orientation of D1 structures and kinematic indicators is strongly affected by subsequent D2 deformation. 1) Such overprint is well-documented in the D2-Aris Syncline where an early subhorizontal NE-stretching lineation is preserved on the southern normal limb but rotated into a moderate north-westerly down-dip on the northern, overturned limb that is limited by a D2 thrust (Walter, 1999; Fig. 2). 2) In the hinge zone of the Hakos Fold, sigma-clasts of a conglomerate grouped into the Otsus Nappe are rotated around the D2-fold axis of this structure, hence suggesting a southwest directed shear sense on the overturned limb (Fig. 7f). This demonstrates the D1 transport of the Otsus Nappe and subsequent deformation by isoclinal folding of the Hakos Nappe.

M1 Metamorphism

The paragenesis garnet-biotite-hornblende-epidote-calcite-albite-Mg-chlorite within mafic rocks and muscovite-biotite, quartz-albite-Mg-chlorite-epidote ± garnet in metapelite constrains upper greenschist to lower amphibolite facies M1 conditions (Pfurr, 1990). Calc-silicate rocks are constituted by ribbon quartz with thin interlayers of early tremolite parallel to the S1-foliation. It is partly transformed into syn-kinematic, zoned D1 garnet with a core characterised by early rotational snowball structures and abundant quartz inclusions whereas the rim shows helicitic growth with less abundant inclusions of tremolite. A second tremolite generation crystallised together with black tourmaline and rutile (Warkus, 1997).

D2-deformation: SE-directed thrusting and folding

The D2 tectonic phase corresponds to the main collision stage of the Damara Orogeny and, in the SMZ is marked by SE-directed nappe tectonics. D2 structures, described below, include several generations of linear and planar fabrics, folds of small- to regional-scale and flat to moderately NW-dipping thrusts. The latter constitute the main boundaries in the HNC but also affected the SDZ and, by the Areb Shear Zone, define the limit with the autochthonous domain.

1) SE-directed D2 transport is best recorded in the thrusts that limit the Blaukrantz Nappe. The tectonic melange that, on Berghof 222, marks the contact with the underlying RNC has been described in a previous paragraph. The crustal-scale importance of the structure is further demonstrated by the presence of m-scale S/C/C' fabrics (Fig. 10a), progressively rotated stretching lineations and axes of non-cylindrical, sheath, and often rootless folds (Figs 8e, 10b). Underlying rocks of the RNC turn in the vicinity of the contact into mylonite (Fig. 10c). High-resolution airborne geophysical data and field observation document magnetite enrichment at the base of the Blaukrantz Nappe (Fig. 11) that is associated with post-kinematic grunerite and garnet. Further, irregular enrichment of magnetite in underlying rocks of the SDZ (Fig. 11) argues for mobilisation and precipitation of iron-oxide from hydrothermal or metamorphic fluids. The overlying graphite schist of the Blaukrantz Nappe acted in this model as an impermeable barrier. A sedimentary origin of this iron-rich layer (Vietor, 1999) is therefore considered unlikely.

Structural decoupling of the Blaukrantz and Vaalgras nappes has been described before. Occasional talc and magnesite boudins in the thrust, separating the two units, are probably derived from alteration of sheared tectonic slivers of dolomite rock and may have acted as lubricant during tectonic transport.

2) SE-verging D2- folds, up to several 10 km size are the most prominent features of the geological map. Field observation shows their progressive evolution in tightness from early isoclinal through tight to late open. The spectacular Hakos Fold Nappe of about 10 km wave length and with limbs more than 50 km long is probably the largest fold structure of the Damara Orogen; it is characterised by the isoclinal tightness, the recumbent to shallow SE-plunging attitude, and lineations (fold axes, stretching and S_{01-2} intersection) that consistently dip shallow to the NE (Figs 2, 4, 6). The present plunge of the front of the fold nappe towards the SE is the result of subsequent coaxial deformation that also refolded the axial plane of the structure into a series of open folds.

Rocks of the Chaibis Nappe form the core of the Hakos Fold; here the normal and overturned limbs are in tectonic contact along a refolded blind thrust tracing the fold axial plane. Outwards follow the Chausib and Blaukrantz nappes, which on the overturned limb were refolded into isoclinal to open folds of 100-m scale (Fig. 8c). A giant talc-quartz-dolomite body of about 800 m length and 200 m width crystallised in tension gashes in the centre of the Hakos Fold Hinge (De Waal, 1966; Behr *et al*. 1983; Marais *et al*. 1995).

A second, less conspicuous major fold structure below the Hakos Fold is outlined in the NE sector of the Natas Dome by basement slivers of the RNC that are mixed with graphite schist and carbonate rocks of the Blaukrantz Nappe (Fig. 3). Their arrangement defines a major tight to isoclinal, recumbent to SE-plunging D2-anticline probably combining the Aris and Rooisand synclines that in this locality are sheared off (Figs 3, 4). Along strike, the north-western normal limb of this anticline is sheared off along the Berghof Thrust, resulting in the tectonic juxtaposition of the Chausib Nappe with the RNC. The fold geometry and the orientation of lineations are marked by systematic change from the SMZ to the autochthonous domain. Folds in the SMZ, such as the Hakos Structure, are cylindrical in shape and their axes generally plunge shallowly towards the NE. Drastic change of this fold geometry occurs through the SE limb of the Djab Syncline, within the SDZ and, at a smaller scale, elsewhere in zones of high shear (Fig. 2). Here, SE directed D2-thrusting resulted in the development of non-cylindrical folds and progressive rotation of stretching lineations and fold axes into the direction of transport (Fig. 6b, h; 8c-e). The shape of these folds passes from open through isoclinal to sheath fold geometries while the curvilinear axial plane commonly parallels the basement relief. The Weissenfels Syncline and Picadilly Structures are km-scale examples of these fold types. On Corona 223, the geological map shows 1) that the latter consists of closed synclines with opposing axial plane orientation that are linked by a gentle anticline and 2) the constant plunge of all lineations to the NW (Figs 2, 3, 9) and 3) that a dolomite marker horizon defines a closed structure (Figs 2, 3). These features suggest a km-scale sheath fold combined with tectonic stacking. The interpretation as fold interference structure is excluded because of the parallel orientation of all lineations. Similar fold geometries have been mapped in graphite schist of the Weissenfels Syncline (Fig. 9).

The progressive evolution from cylindrical folds with parallel axes to non-cylindrical and sheath folds with curved fold hinges is also illustrated in stereo plots by the continuous spread of D2-lineations from the NW to the NE quadrant (Fig. 6c, e, g, h). The pole of the plane defined by the lineations coincides with the maximum of the S_{01}-poles, indicating their rotation into the normal plane of maximum D2 shortening, parallel to the S2-schistosity. In contrast, stretching lineations in the Natas Dome and the Chaibis Nappe plunge with little variation shallow to the SW and NE (Fig. 2, Fig 6d, f). The shear sense in these domains is consistently to the NE, hence their origin is mainly attributed to D1. Their orientation subparallel to D2 lineations elsewhere (outside shear zones) suggests that the D2 event may have enhanced the lineations without developing a distinct generation. It is interesting to note that the maximum of lineations within the Natas Dome and Hakos Fold varies at 10 degrees from the maxima determined in the other domains supporting their different evolution. The apparent lack of NW-plunging lineations in the Natas Dome and Hakos Fold further implies that these units were less affected by D2 thrusting than the Samara Dagbrek Zone. All observations demonstrate the significant increase of strain in the SMZ towards the contact with the autochthonous domain (Vietor, 1999).

Rocks of the study area record up to three generations of schistosity related to D2. A regional S_{21}-schistosity dips commonly moderately to the NW (Fig. 6a-h) and, due to the isoclinal geometry of the folds is generally subparallel to the S_{01}-fabrics. However, in the vicinity of the basement and in domains of high strain the S_2-foliation parallels again the trace of the axial curvature of non-cylindrical D_2 folds. Subsequent co-axial deformation produced closed to open folds and here the $S_{0,1/2,1}$-$S_{2,2}$ relationship allow the establishment of the structural normal or overturned position of rocks with respect to these $D_{2,2/3}$ structures (Fig. 8f). This is best illustrated in the front of the Hakos Fold where $S_{2,1}$ to $S_{2,3}$ foliations are locally developed in the axial planes of closed to open folds (Fig. 8a). They all dip in the same direction but vary in their inclination, hence demonstrating the continuum of deformation in an overall constant stress field.

Local ramp structures in the hinge of the Weissenfels Syncline demonstrate continuation of SE-directed transport (Fig. 10d) resulting in late-tectonic stacking of rocks from the RNC, the Blaukrantz and the Vaalgras nappes. Simple shear also affected underlying foliated basement granites. In the vicinity of the contact they turn into L-tectonite marked by a prominent stretching lineation with shallow to moderate plunge from N to NNE).

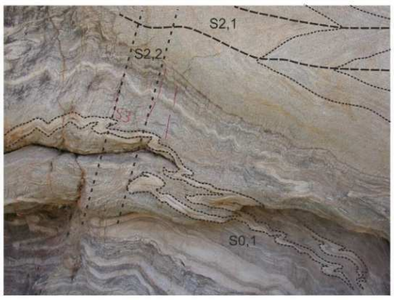

Fig 8a: Complex interference pattern in homoaxially folded schist of the Vaalgras Nappe; front of the Hakos Fold (Hakos 25).

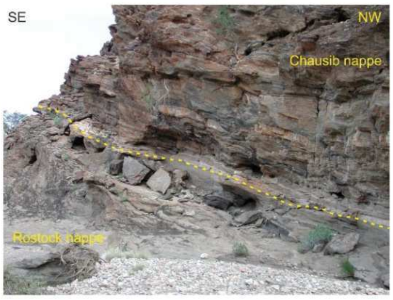

Fig 8b: The Berghof Thrust separates pre-Damaran rocks of the Rostock Nappe (base) from distal turbidites of the Chausib Nappe (top). (Berghof 222)

Fig 8c: Internal closed to isoclinal folding at the overturned limb of the Hakos Structure (Tantus 30).

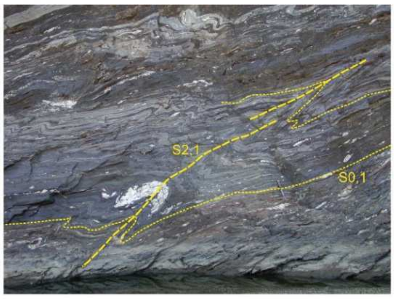

Fig 8d: Cylindrical, dysharmonic folds at the northern limb of the Hakos Structure (Chaibis 29)

Fig 8e: Refolded sheath folds occur in zones of high shear (Djab 26)

Figure 8f: The D2 Rooisand Syncline (Berghof 222)

Fig 8g: An overturned position of the S2,1 schistosity is shown by the steeper dip of S0,1 (Klein Chausib 408).

Figure 8. Tectono-metamorphic structures related to the SE-directed D2-deformation.

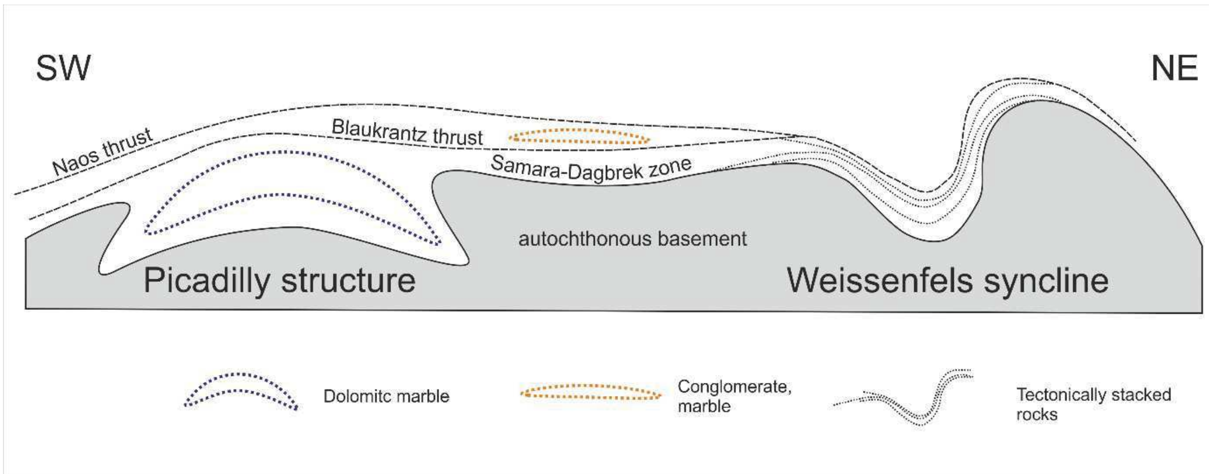

Figure 9. Schematic cross-section across the Samara-Dagbrek Zone, which illustrates the development of major sheath folds and/or tectonic stacks. Fold axes and stretching lineations are perpendicular to the cross section.

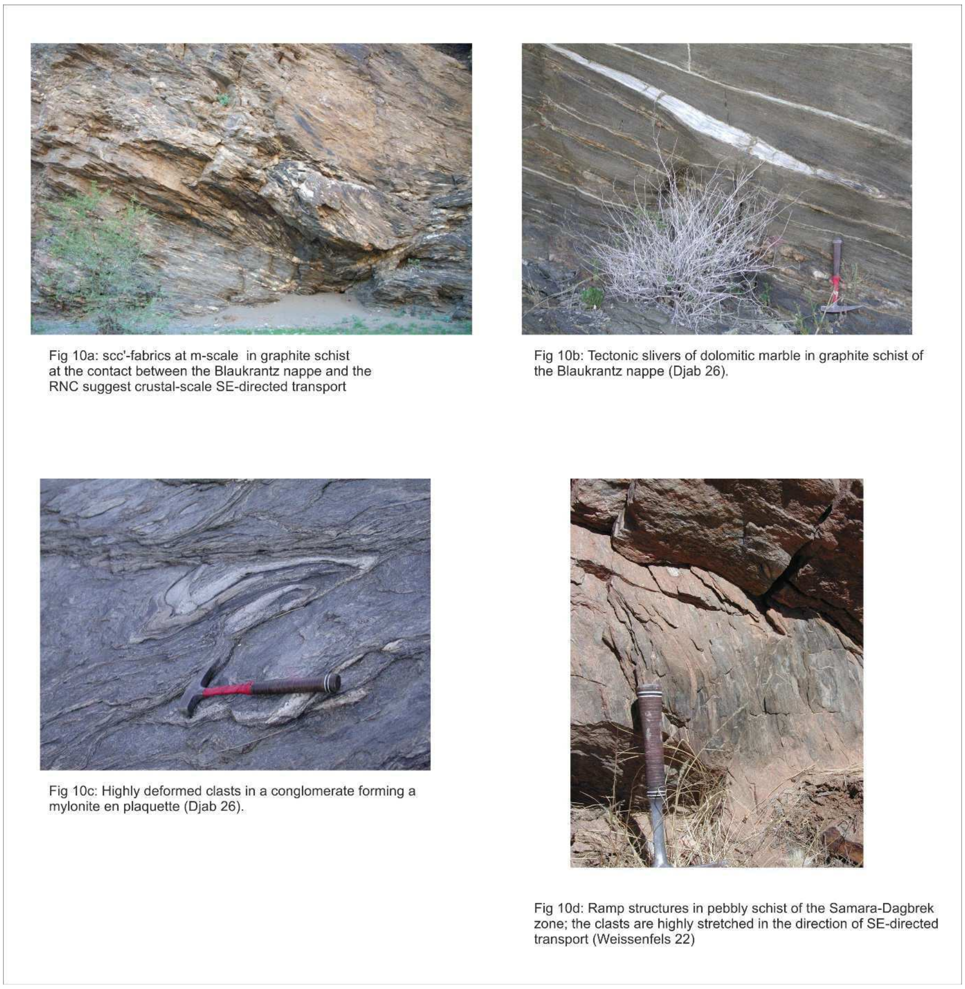

Fig 10a: scc'-fabrics at m-scale in graphite schist at the contact between the Blaukrantz nappe and the RNC suggest crustal-scale SE-directed transport

Fig 10b: Tectonic slivers of dolomitic marble in graphite schist of the Blaukrantz nappe (Djab 26).

Fig 10c: Highly deformed clasts in a conglomerate forming a mylonite en plaquette (Djab 26).

Fig 10d: Ramp structures in pebbly schist of the Samara-Dagbrek zone; the clasts are highly stretched in the direction of SE-directed transport (Weissenfels 22)

Figure 10. Structures at the basal contact of the Blaukrantz Nappe.

M2 Metamorphism

M2 conditions are again characterised by the mineral paragenesis almandine-biotite-muscovite-feldspar and quartz which crystallised in metapelitic rocks both syn- and postkinematic and show that metamorphism outlasted deformation. Biotite and garnet reacted during retrograde metamorphism frequently to chlorite. The syn- to postkinematic paragenesis kyanite-quartz-(mica) occurs mainly in alumina-rich graphite schist of the

Blaukrantz Nappe and in the upper portion of graded turbidites of the Chausib Nappe (Porada & Wittig, 1976). However, kyanite also crystallised in the foliation of quartz-biotite schist, which locally occurs at the base of the Vaalgras Nappe. Unoriented quartz-kyanite aggregates up to 30 cm size occur close to and within major shear zones around the Natas Dome. This suggests continued activity of metamorphic fluids along impermeable barriers under minimum pressures of 6-8 kb corresponding to a depth of 18-24 km. D2 late-kinematic staurolite has been proven in quartz-muscovite-chlorite schist of the Melrose Formation in the northern part of the study area (Hoffmann, 1983).

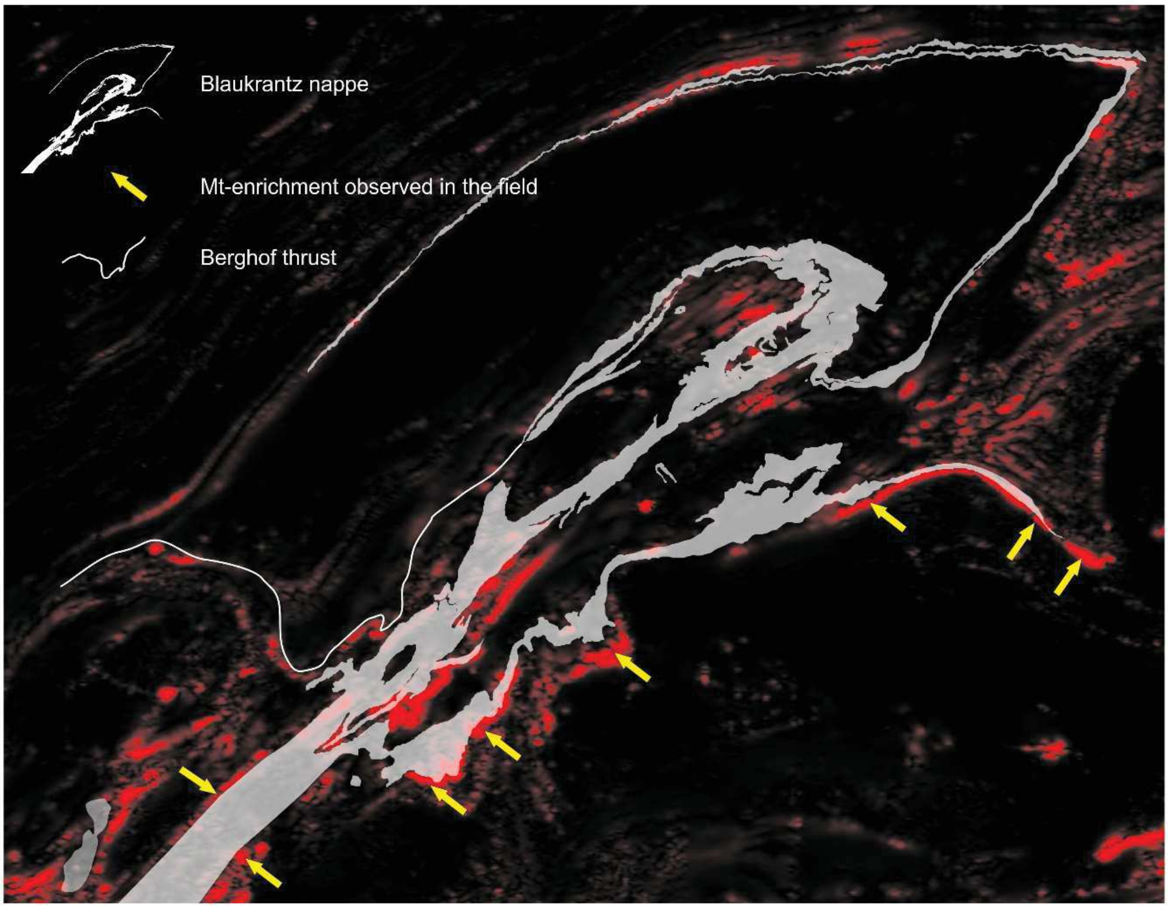

Figure 11. High-resolution airborne geophysics (magnetics; 1st vd) demonstrates? near the basal contact of the Blaukrantz Nappe? irregular positive anomalies in various underlying formations caused by the enrichment of magnetite. The iron was probably transported by metamorphic fluids and crystallised in the thrust; the overlying graphite schist may have acted as an impermeable barrier.

D3-deformation: East to south-east directed compression and up-doming

The D3 deformation folded the Hakos area around N-S axes into gentle kilometre-scale folds but did not develop a regional D3 axial plane foliation or fracture cleavage. The interference of the D_3-folds with D_1/ D_2 structures further caused up-doming of the RNC in the type locality, created the Natas Dome and formed the crescent shape of the western tail of the Hakos Fold. In the Natas Dome S_{01} dips at shallow to moderate angles off the centre of the structure. Small-scale structures comprise local steeply dipping strike slip cleavage and a shallow plunging crenulation lineation that results from the intersection of this cleavage with S_{01} and S_2 foliations, respectively. The crenulation lineation changes from an N-S direction in the west to a NE-SW direction in the east. It coincides with a change in the vergence of the strike slip cleavage from E to SE. At microscopic scale, D_3 rotated mica into the D_3 axial plane of microfolds but did not lead to new growth of minerals.

Discussion and Conclusion

The lack of fossils renders difficult the reconstruction of the stratigraphy in any Precambrian fold and thrust belt and it is often even challenging to define the main nappes and their thrust boundaries in pervasively sheared rocks. Such boundaries are sometimes indicated by different structural fabrics across the contact, by a jump in the metamorphic grade, by juxtaposition of rocks of different age and/or deposition environment, or by concentration of shear along the contact. Reconstruction of the Neoproterozoic stratigraphy is even more problematic in the study area because of 1) the absence of igneous Neoproterozoic rocks whose age can be determined and 2) the simultaneity of Wilson cycle processes and Neoproterozoic glaciations. They both trigger important sea level changes independently from each other and result in possibly unusual sedimentary sequences that in the following periods may have been disrupted by tectonic stacking during the Pan-African Event.

Stratigraphic Reconstruction

Previous models are based on either 1) the autochthonous to parautochthonous position of Neoproterozoic sediments with largely preserved stratigraphy (DeWaal, 1966; Porada and Wittig, 1976, 1983; Weber, 1994), or 2) nappe tectonics but without any further implications to the stratigraphy (Hartnady, 1978), or 3) nappe tectonics, accompanied by largely preserved sedimentary sequences hence, allowing stratigraphic reconstruction (Hoffmann, 1983, 1989; Vietor, 1999).

The present study shows that large-scale thrusting and nappe tectonics during two main deformation events affected all units of the SMZ largely obscuring the former stratigraphy. This conclusion is based on a number of key observations demonstrating the large-scale thrust nature of all contacts in the SDZ, the RNC and the HNC.

1) The predominantly thrust nature of contacts between rocks close to and within the SDZ is best demonstrated in two places: (A) In the Weissenfels Syncline, sheared slivers of the Berghof Nappe (diamictite, carbonate, quartzite) overlie a melange of rocks from the SDZ (carbonate phyllite) and the Blaukrantz Nappe (graphite schist). Graphite schist is even locally in contact with autochthonous orthogneiss that, due to the intensity of shearing turned locally into L-tectonite. Further east, the SDZ is thrust off by the HNC that here is in contact with pre-Damara basement rocks. Abundant flats and ramps within graphite schist and diamictite document further complication of the tectonic stack. (B) On Kromhoek 416, isoclinally-folded rocks of the SDZ (dolomite and phyllite) are in thrust contact with the Pre-Damara basement (Blecher, 1991).

2) The RNC is comprised of pre-Damara basement rocks that are stacked with Nosib Group sediments. It is structurally characterised by the penetrative subhorizontal s1-fabric and low-angle shear zones (Pfurr *et al*. 1987; Pfurr, 1990) implying large-scale transports which in turn constrains the allochthonous position of the overlying HNC.

3) In the Natas Dome, interlayering of pre-Damara rocks, the Nosib Group and the Hakos Group demonstrates tectonic stacking of the HNC and RNC.

4) Low-angle thrusts of the RNC are obscured below overthrust graphite schist of the Blaukrantz Nappe arguing for polyphase nappe emplacement.

5) Kyanite within the Berghof low-angle thrust constrains minimum pressures of 6-8 kb corresponding to a depth of 18-24 km and implying deca-kilometric transports.

6) The boundary between the Chaibis and Chausib nappes is defined by the abrupt change in the composition of sedimentary rocks that are interpreted as proximal and distal turbidites. Such change is unusual in tectonically undisturbed fan systems marked by tran-

sition of these facies and again argues for the tectonic nature of the contact. Important thrusting between these nappes is further demonstrated by change in thickness of the Chausib Nappe from more than 1500 m in the hinge zone of the Hakos Fold to a few metres at the north-western flank, where the rock turns into ultramylonite.

7) Distal slope facies of the Chausib Nappe overlie the Pre-Damara Basement along a thrust contact implying that former coeval platform and proximal slope facies are sheared off. The tectonic melange of graphite schist and mylonitic quartzite in the thrust demonstrates the intensity of deformation.

8) The Otsus Nappe is constituted by slivers of various rock types that resemble those of the Berghof Nappe. This illustrates the problem of stratigraphic reconstruction of units in different structural positions.

9) The crustal-scale of the thrust below the Blaukrantz Nappe is demonstrated by the low-angle inclination, the presence of m-scale S/C/C'-fabrics, rootless and sheath folds, tectonic slivers of underlying rocks and the regional tectonic discordance outlined by the unit.

10) The tectonic nature of the contact between the Blaukrantz and Vaalgras nappes is shown by the change in structural style across the contact and, within the shear zone the presence of contorted folds, boudins of talc schist and magnesite as well as syn-to post-kinematic kyanite. The basal beds of the Vaalgras Nappe are sheared along strike into many tectonic slivers of quartzite, conglomerate and pebbly schist documenting further internal thrusting.

Vietor (1996, 1999) proposed a revised stratigraphy for the southern margin of the DO assuming that, despite important thrusting, stratigraphic relationships were largely preserved. In his model, the basal Nosib Group consists of fluviatile conglomerate and quartzite that grade laterally into marine proximal and distal turbidites of the Chaibis and Chausib formations deposited within a half-graben. This is at great variance to other workers who classify the Chausib and Chaibis formations as part of the Swakop Group (SACS, 1980; Hoffmann, 1989).

Marine carbonate rocks that in our study are mapped as Samara Formation of the SDZ, the Berghof Nappe and the Otsus Nappe have been combined by Vietor (1999) into the Corona Formation. The sediments are interpreted as recording regional down-warping that possibly marks the transition from the rift to the spreading stage of the Kuiseb Basin to the north. The overlying black shales of the Blaukrantz Formation are believed to reflect marine transgression over an extensive shallow marine platform of a starved passive margin. Glaciogenic sediments of the Naos Formation (Vaalgras Subgroup) are thought to have been deposited within a half-graben thus explaining the important changes in thickness between the Djab Syncline in front of the Hakos Fold and the north-western limb of the structure (Vietor, 1999). In contrast, Hoffmann (1989) subdivided the carbonates of the SDZ into the Waldburg Formation at the base of the Kudis Subgroup and into the Samara Formation that postdates the Naos Formation (Table 1).

However, crustal-scale shear zones demonstrated by our study, imply that, based on field observation only, it is impossible to establish the Neoproterozoic stratigraphy. This is illustrated by two examples.

1) Rocks from the Chaibis, Chausib and Blaukrantz nappes all consist of graphite schist and sandstone. Based on the variation in the proportion of the facies, they may be interpreted as proximal (Chaibis), distal (Chausib) and very distal (Blaukrantz) slope sediments of a continental passive margin. Alternatively, the graphite schist has been classified as transgression facies unrelated to the Chausib and Chaibis sediments (Vietor, 1999). On Berghof 222, graphite schist is intercalated with rare horizons of sandstone, hence documenting the transition of basal Blaukrantz into overlying Chausib facies and arguing for their deposition in one prograding turbiditic fan system (SACS, 1980; Hoffmann, 1989). However, it is impossible to decide whether this system was active at Nosib times, or post-dates deposition of platform carbonates of the Waldburg Formation (Hoffmann, 1989). The transition of Nosib sediments into Chaibis facies rocks at the northern flank of the Natas Dome is probably the result of tectonic stacking that is characteristic for this domain.

2) Similarly, the stratigraphic relationship between the Otsus, Berghof and Samara units remain unsolved. Although similar in rock composition, they are stacked in various positions of the nappe pile or overlie the autochthonous basement along sheared contacts. The same holds true for the Vaalgras Nappe and all overlying rocks derived from the adja-

cent 'root zone' to the north of the HNC. Although sheared, the stratigraphy may have been preserved within individual nappes. However, there are no constraints at all, which would allow establishment of the former stratigraphic position of the nappes.

In conclusion, our structural observations show the uncertainty of all previous stratigraphic schemes. The definition of the main tectono-metamorphic domains and the structural frame work is considered as a first step in reconstructing the regional geology. In the absence of absolute time markers a detailed metamorphic study may characterise the individual nappes and quantify their specific PT-paths. Geochemical analyses of mafic rocks of possible volcanic origin occurring in the Chaibis, Chausib and Vaalgras nappes may help to elucidate stratigraphic relationships should they record systematic transition from intraplate basalts to MOR-basalts or show similar composition. Age determination of detrital zircons may trace systematic change in the source area and help correlating of nappes. C-O isotope signatures of (cap) carbonate rocks are powerful for their correlation with Sturtian (710 Ma) or Marinoan (635 Ma) global glaciations and may allow correlation of structurally distinct units (e.g. Otsus and Berghof nappes, Waldburg, Samara formations).

Deformation History

Structural analysis of the study area reveals three stages of deformation D1-D3. NE-directed transport during D1 is recorded by S/C/C'-fabrics, asymmetric pressure shadows and vergence of D1 folds. It resulted in the initial stacking of the lower nappes grouped into the RNC and the HNC. Possible D1 fabrics are obscured in the upper Blaukrantz and Vaalgras nappes due to intense D2 overprint. A regional flat lying bedding-parallel S_{01}-foliation is developed in all rocks of the SMZ but may be diachronous.

The D2 event is characterised by SE-directed transports, recorded again by S/C/C'-fabrics, asymmetric pressure shadows and the fold vergence. D2-folds up to deca-km scale dominate the outcrop geometry of the study area. A NE plunging stretching lineation parallel to the fold axis is commonly developed. Continuous deformation in the same overall stress field is demonstrated by homoaxial early isoclinal through tight to late open folds that are associated with schistosities $S_{2,1}$ to $S_{2,3}$. SE-directed shear is focused at the contacts between the nappes and in the SDZ. Here, stretching lineations are progressively rotated into the SE transport direction and folds become non-cylindrical.

Late-stage east-west compression D3 resulted in the local development of a crenulation cleavage and gentle folding around N-S trending axes leading to dome and basin interference structures.

Regional Implications

Miller (2008) summarised the history of the DO distinguishing the coastal Kaoko and Gariep branches from the inland branch, the DO s.s. but presenting one protracted tectono-metamorphic history for all terranes. A wide Adamaster Ocean existed between the South American and Congo cratons from about 800 Ma onwards. It was closed between 595-575 Ma resulting in the formation of the N-S striking Kaoko Belt. A first phase of transpressive sinistral continental collision changed subsequently into a SE and, about 550 Ma ago, into the final W over E direction of transports concluding the orogeny in the Kaoko Belt.

The inland branch of the DO is interpreted to record prolonged rifting from 800 Ma onwards. According to Miller (2008) it was followed by limited ocean spreading between 609-600 Ma leading to a narrow marine basin, the Khomas Sea. A possible first phase of sinistral transpressive NE over SW D1 deformation is assumed between 580-575 Ma resulting within the SMZ in the stacking of (unmetamorphic?) nappes. This phase of deformation is, however, so far not constrained by any geochronological data. The onset of syntectonic Salem-type granites and the predominant regional M1-D2 tectono-metamorphic event commenced about 550 Ma ago being largely responsible for the present geometry of the inland branch. Final closure of the Khomas Basin is proposed at 542 Ma and was followed around 535 Ma by regional post-tectonic M2 metamorphism leading in the western part of the central zone to anatexis. Voluminous intrusions of post-tectonic granites (Donkerhook Suite) occurred between 535-510 Ma. The isostatic rebound of the orogen resulted at 495 Ma in the gravity-induced tectonic emplacement of the Naukluft Nappe Complex.

The history of the Gariep coastal branch is summarised by Frimmel (2008). The first phase of rifting is constrained in this terrane by alkaline magmatism at about 850 Ma followed by continental break-up 750 Ma ago. A major hiatus exists from then onwards that includes the Marinoan glaciation at 635 Ma. Oceanic crust formation within a back arc basin took place approximately 600 m.y. ago. The formation of this basin is explained by reactivation of the previous rift in response to a large, 640-590 Ma active, volcanic arc system in the Dom Feliciano Belt in south-eastern Brazil and Uruguay. Therefore the main suture lies west of that arc in South America. Transpressive continental collision occurred between 550-545 Ma. It was directed towards the SE and is characterised by a strong sinistral wrench component. The Gariep Belt is described as a fold and thrust belt characterised by a steep orogenic front and metamorphic conditions up to medium-grade in the lower part of the tectonic pile.

The summaries given by Miller (2008) and Frimmel (2008) show that the tectonometamorphism recorded in the Kaoko Belt precedes the orogenies in both the Gariep and Damara belts, the latter of which appear to be coeval. Therefore, the collision of the Kalahari Craton with the South American and the Congo cratons (i.e. the Damara Orogeny) is regarded by us as an event distinct from the Kaoko Orogeny implying, in contrast to Martin & Porada (1977a, b), Porada (1979, 1983, 1985, 1989; Miller 2008) a wide ocean between the Kalahari and Congo plates (Hartnady, 1978; Barnes & Sawyer, 1980; Kasch, 1983; Coward, 1983; Goscombe *et al*. 2005; Jacobs *et al*. 2008; Gray *et al*. 2008).

In the study area, D1 fabrics are best preserved in the lower nappes of the RNC and document west-directed oblique collision of the Kalahari with the Congo Craton, with an important dextral horizontal component (Fig. 12a). This corresponds in the Gariep Belt to early eastward transport of the oceanic Marmora Terrane over, and tectonic stacking with, the Porth Nolloth Terrane constituted by continental to passive margin sediments (Frimmel, 2008).

In the study area D2 records the change in tectonic style and direction into compressive, top-to-the-SE directed transports resulting in the stacking of D2 fold and thrust nappes (Fig. 12b). The variation in the orientation of D2 lineations as well as the local variation in shape from cylindrical to sheath-fold geometry is a function of the strain rate which is highest in the vicinity of the autochthonous basement and within individual shear zones. In the Gariep Belt, D2 corresponds to transpression marked by overall top-to-the-SE transport with an important sinistral horizontal component. East- and south-east verging folds and thrusts developed along north-eastern lateral and southern frontal ramps, respectively (Gresse, 1994).

D3 east-west compression is in our model related to the final adjustment between the Kalahari and South American cratons.

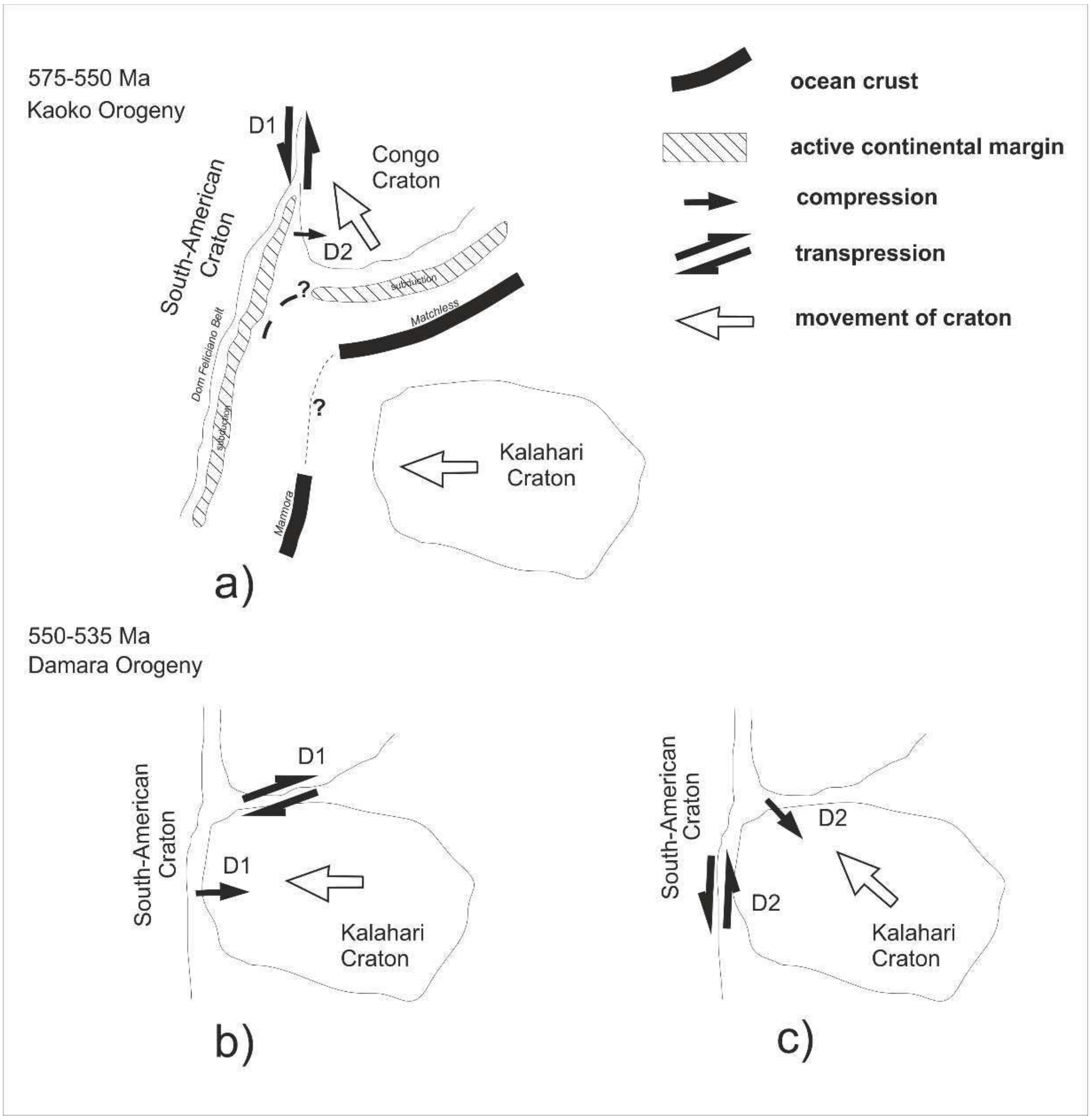

Figure 12. Interpretation of the observed deformation stages D1 and D2 in relationship to the relative plate movements of the Kalahari, Congo and South American cratons during the Kaoko and Damara orogenies.

Problems

One major problem of the proposed model is the apparent absence of a Neoproterozoic magmatic arc that should form the active southern margin of the Congo Craton parallel to the strike of the "inland" branch of the DO. Such an arc has nowhere been observed but could have been eroded during uplift of the orogen as it was situated on the overriding plate. However, flysch sediments of the Khomas Group interpreted as a former accretionary prism (Kukla & Stanistreet, 1991) and molassic sediments of the Mulden and Nama groups should record this arc magmatism. The study of detrital zircons across the Damara Orogen (Foster *et al*. 2015) shows the presence of zircons as young as 602 +/- 26 Ma in Kuiseb Schist of the Southern Zone whereas zircons in the the Nama Group yielded Neoproterozoic peaks at 574, 596, 652, 683 and 796 Ma (Blanco *et al*. 2011). Facies and palaeocurrent analyses, silicified volcanic ash beds and chromian spinel-bearing sandstones of the Nama Basin points to a volcanic arc in the Damara Belt (Blanco *et al*. 2009) located in the adjacent Damara Belt (Blanco *et al*. 2011).

Alternatively the Khomas Sea may represent the inland continuation of the back arc basin that developed in the Gariep Belt (Fig. 12, stippled line) with the Matchless Member forming the equivalent of the Chameis Group described by Frimmel (2008, 2011). However, even though the Matchless Member and the Chameis Group may have developed at the same time there are major differences with respect to the proportion of contemporaneous sedimentation: the Matchless shows affinity with ocean ridge basalts (Finnemore, 1978, Schmidt & Wedepohl, 1983) but is drowned in sediments of the Kuiseb Group whereas the Chameis Group is characterised by the predominance of basalts with an oceanic intra-plate affinity, minor gabbros with mid-ocean ridge affinity and minor reef facies interpreted as remnants of former ocean islands (Frimmel, 2008, 2011).

Acknowledgments

The present study has been performed within the regional mapping program of the Geological Survey of Namibia. The authors would like to thank the Director Gabi Schneider for her support throughout the project. Martin Pickford is thanked for the careful review of the manuscript.

Digital Supplementary Data

The making of the Hakos area file and the georeferenced structural data are available on request from the corresponding author.

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