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Tectonic evolution of the southern Kaoko belt, Namibia

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Abstract

The tectonic evolution at the junction of the Panafrican Kaoko and Damara belts is well recorded in the siliciclastic and carbonate successions of the Neoproterozoic Zerrissene turbidite system, metamorphosed to the biotite zone of the greenschist facies. The structures in the turbidites are attributed to two main deformational events. The older one generated two continuous folding phases, D1 and D2, and the younger one resulted in D3 deformation. D1, of dominant E–W shortening, caused upright kilometre-scale folds with well-developed axial planar cleavage, N–S trending axial planes and subhorizontal axes. This phase graded into D2 that refolded the first folds coaxialy and developed a crenulation cleavage at a high angle to the first cleavage. D2 is interpreted as a phase associated with shear movement to the north and minor continuing E–W shortening. The third phase, D3, of apparently sinistral transpression caused localised fold trains on a metre to kilometre scale with NE to NNE trending subvertical axial planes. The first two phases predate intrusion of a 530 Ma syenite and probably correlate with the main deformation in both the Kaoko and Damara belts. D2 is likely coeval with motion of the sinistral strike-slip Purros lineament in the central part of the Kaoko belt. D3 postdates the syenite intrusion and is restricted in occurrence to the Damara and southern Kaoko belts. Deformation in the Damara belt therefore outlasted that in the central and northern parts of the Kaoko belt.

Keywords: Tectonic evolution; Kaoko belts; Damara belts

1. Introduction

Two related Panafrican mobile belts dominate the geology of northwestern Namibia. The Kaoko belt, trending NW–SE, flanks the coast. The Damara belt trends NE–SW and is also referred to as the inland branch of the Damara Orogen (Miller, 1983; Fig. 1). The Damara belt, with predominantly south to southeast vergent structures, has been subdivided into a northern, central and southern tectono-stratigraphic zone (Miller, 1983; Porada, 1989; Fig. 1) and was interpreted as a typical intracratonic belt or aulacogen (e.g. Porada, 1989; Kröner, 1981). It is now generally considered that the Damara Belt is a collisional belt between the Congo and Kalahari cratons/paleocontinents, associated with subduction towards the north under the Congo Craton, with an accretionary prism constituting essentially the southern zone (Miller, 1983; Kukla and Stanistreet, 1991; Prave, 1996). The Kaoko belt, with predominantly ENE-vergent structures, has been subdivided into eastern, central, western and southern zones (Miller and Grote, 1988; Porada, 1989; Fig. 1). These zones are separated by shear belts, such as the Purros lineament and the Sesfontein thrust (Figs. 1 and 2), and exhibit increasing metamorphic grade from greenschist facies in the east to upper amphibolite/granulite facies in the west. Apart from E–W shortening, an important left-lateral displacement component was recognised in the Kaoko belt (Dürr et al., 1995). The Kaoko belt was also originally interpreted to be intracratonic (Dürr and Dingeldey, 1996) based on the inferred presence of basement inliers of the Congo Craton throughout the belt, but this interpretation is now questioned on the basis of new field and age data (Seth et al., 1998). Seth et al. (1998) have shown that most of the so-called basement in the belt consists either of lenses of Proterozoic or Archaean rocks with ages significantly different from ages in the adjacent part of the Congo Craton, or of synkinematic granitoids of Neoproterozoic age. A collisional origin seems therefore more likely for the Kaoko Belt (Stanistreet and Charlesworth, 2001). The Brazilian counterpart of the Kaoko

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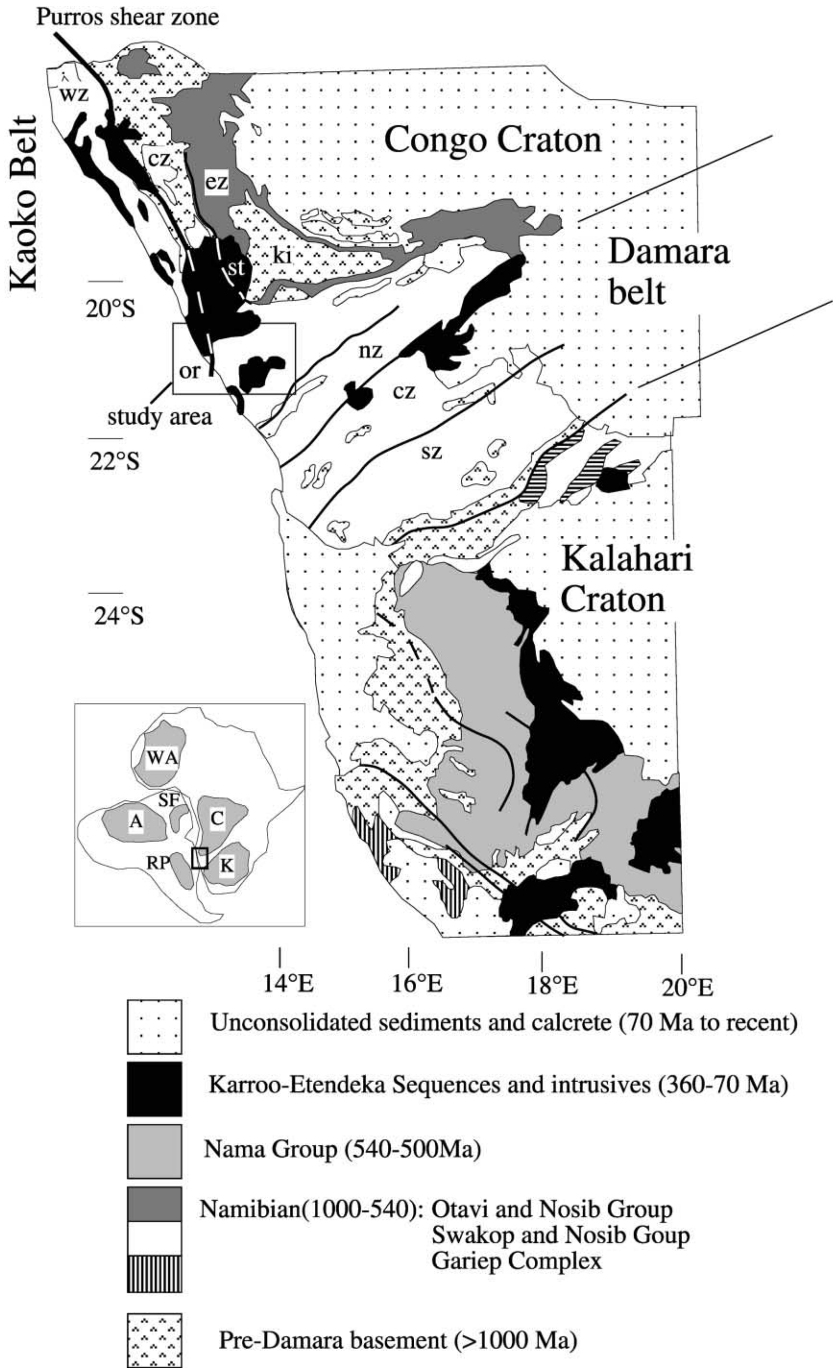

Fig. 1. Schematic location map of the Kaoko and Damara belts; the investigated area (Fig. 2) is marked by a box. ki—Kamanjab Inlier; nz, cz, sz, wz, ez; northern-, central-, southern-, western-, and eastern-zones of the Damara or Kaoko belts; or—Ogden Rocks shear zone; st—Sesfontein thrust. Cratons indicated in inset: A—Amazon; C—Congo; K—Kalahari; RP—Rio de la Plata; SF—São Francisco; WA—West African. The Dom Feliciano Belt lies east of the Rio de la Plata Craton.

belt, the Dom Feliciano belt, is also interpreted as collisional (Fernandes et al., 1992; Basei et al., 2000).

The junction of the Kaoko- and Damara belts lies in the catchment area of the Ugab River in northwestern Namibia (Figs. 1 and 2). Several studies were undertaken in this area (e.g. Miller et al., 1983; Miller and Grote, 1988; Hoffman et al., 1994; Coward, 1981, 1983; Swart, 1992) but these have not yet fully explained the nature and internal kinematics of the junction, nor the relative age of tectonic activity in either belt. This paper reports on structural analysis in the Lower Ugab Domain, and contributes new data to the tectonic interpretation of this area. The Lower Ugab Domain (Fig. 2; Hoffmann, 1987) is considered by most authors (Miller,

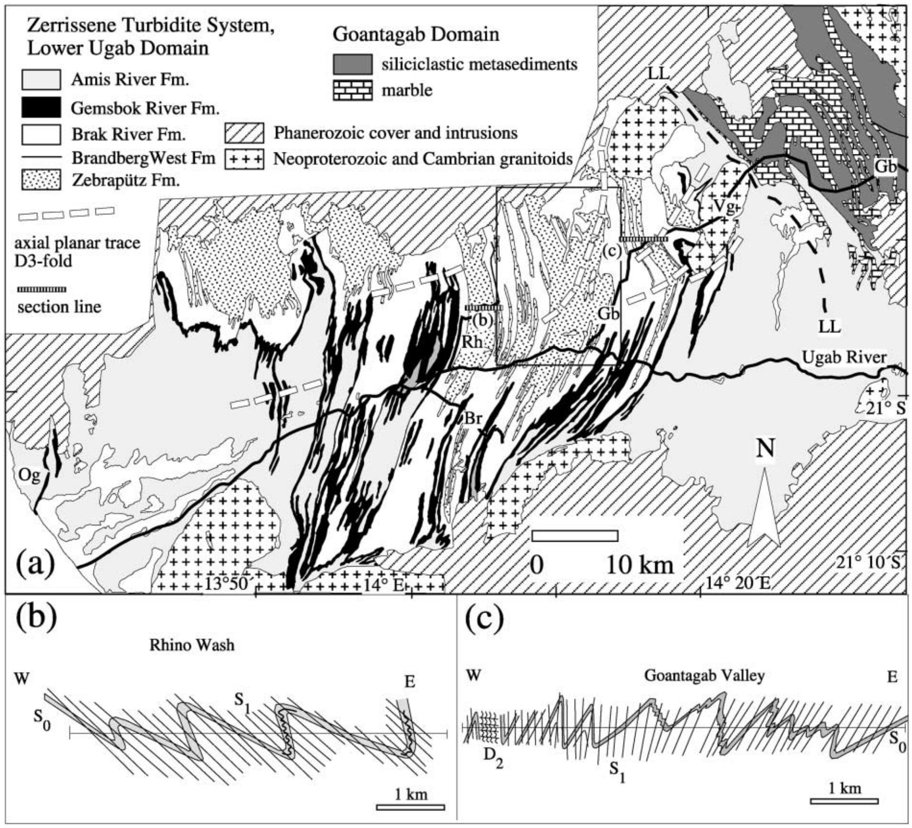

Fig. 2. (a) Schematic map of the Lower Ugab Domain (after the Geological Map of the Damara Orogen, 1:500.000, Miller and Grote, 1988). Ornamented bars indicate position of sections (b) and (c), showing the typical geometry of D1-folds. Dashed white lines show the axial planar trace of major D3-structures. Square in (a) shows the location of the area covered by Fig. 8. Og—Ogden Rocks; Rh—Rhino Wash; Br—Brak River; Gb—Goantagab River; Vg—Voetspoor intrusion; LL—western limit of stretching lineations in the Goantagab Domain.

1983; Miller and Grote, 1988; Hoffman et al., 1994) as the southern zone of the Kaoko belt, separated from the main part of this belt by an area with Phanerozoic cover, mainly basalts of the Cretaceous Etendeka Group (Fig. 1).

2. The Zerrissene turbidite system in the Lower Ugab Domain

A Neoproterozoic turbidite succession known as the Zerrissene turbidite system (Miller et al., 1983; Swart, 1992), crops out continuously over about 100 km in E–W direction in the Lower Ugab Domain between the Ogden Rocks in the west and the Goantagab Domain in the northeast (Fig. 2). Sedimentary and volcanic rocks of the Karroo sequence and volcanic rocks of the Etendeka Group cover the turbidites towards the north and south (Fig. 2; Miller and Grote, 1988).

The Zerrissene turbidite system consists of a cyclic succession of siliciclastic and carbonate turbidites (Miller et al., 1983; Swart, 1992; Fig. 2). Following Swart (1992) and Miller et al. (1983) the succession is about 1.7 km thick without base or top being exposed, and is subdivided into five formations, in stratigraphic order from bottom to top: (1) Zebrapüts Formation (>350 m greywackes and pelites); (2) Brandberg-West Formation (15–20 m marble and pelites); (3) Brak River Formation (500 m greywackes and pelites); (4) Gemsbok River Formation (200 m marble and pelites); (5) Amis River Formation (>550 m greywackes and pelites). In this study sandy, muddy and heterolithic facies were identified in all formations. Both carbonatic and siliciclastic sandstones and mudstones are present. The sandstones are fine, medium and coarse feldspathic arenites and calcarenites. Siliciclastic mudstones dominate over carbonatic ones and are represented by siltites and argillites. Locally conglomerates and intraformational breccias are also present. The recognised lithofacies constitute thin- and thick-bedded classical turbidites and massive sandstone beds. The main sedimentary structures are massive, normal graded, plane and micro-cross-lamination (ripples) and rare flute casts. Few isolated pebbles, possibly of glacial origin (Swart, 1992) were identified in massive sandstone beds of the Brak River Formation.

Miller et al. (1983) interpreted the Zerrissene turbidite system as deepwater deposits, equivalent to the shelf successions of the Swakop Group that crop out towards

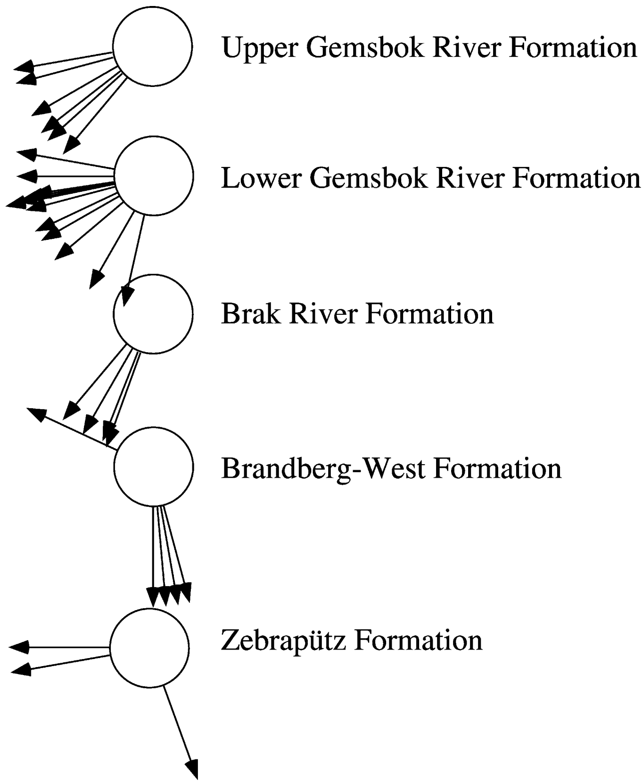

Fig. 3. Frequency diagram of paleocurrent directions of the Zerrissene turbidite system. Each arrow represents one station.

the northeast, with a depositional age between 750 and 540 Ma (Miller, 1983). Swart (1992) presented 23 paleocurrent measurements in the three siliciclastic formations, indicating preferred directions towards the southeast and northeast, and inferred a possible source area to the west that would presently be part of Brazil. We measured paleocurrent directions in thirty localities, mainly from asymmetric cross-lamination seen in three dimensions but also from flute casts, with directions different from those of Swart (1992). Our data indicate a progressive shift of the currents from southward in the Brandberg-West Formation, to southwestward in the Brak River Formation, and finally to westward in the Gemsbok River Formation (Fig. 3). These new data are consistent with a source area in the Congo Craton toward the northeast as originally proposed by Miller et al. (1983). The turbidites are deformed and metamorphosed to middle greenschist facies conditions, and intruded by granite and syenite plutons.

3. The Ogden Rocks and the Goantagab Domain

The Ogden Rocks (Og in Fig. 2; at the coast) consist of poorly exposed turbiditic metasedimentary rocks at least partly similar to the Zerrissene turbidites, intruded by veins and small bodies of granitoids (Hoffman et al., 1994). In contrast to the Zerrissene turbidites the rocks are mylonitised in upper greenschist to lower amphibolite facies along steep to moderately west-dipping shear zones with N–S strike.

The Goantagab Domain (Fig. 2a) has a different lithofacies association. The deformation is more intense than in the Lower Ugab Domain and hampers stratigraphic correlations. Nevertheless individual beds can be mapped from one domain into the other suggesting a gradual transition. The lateral equivalent of the Brak River Formation, apart from similar metasandstones and pelites, has diamictite and quartzite intercalations. A greenschist body of several kilometres in diametre appears intrusive in this formation. The correlate to the Gemsbok River Formation is thicker and contains mappable carbonate breccia strata. The Amis River Formation includes several 1–10 m thick marble intercalations that can be traced 5–7 km into the Lower Ugab Domain. The Goantagab Domain succession seems to represent slightly more proximal deepwater turbidites than the Zerrissene turbidite system, with associated debris flows represented by diamictite and marble breccia. However, the difference between the domains appears mainly the result of a different style of deformation as described below.

4. Deformation

Only short descriptions of deformation in the Lower Ugab Domain have been published to date, e.g. by Coward (1981, 1983), Miller et al. (1983), Miller and Grote (1988) and Freyer (in a geological excursion guide: Hoffman et al. (1994)). Our observations largely confirm their findings but we expand on them as outlined below. Three deformation phases were recognised, labelled D1, D2 and D3.

4.1. First deformation phase, D1

Spectacular meso- to macroscopic open to tight N–S to NW–SE trending folds (Figs. 2, 4a and 6a) with subhorizontal axes (Fig. 6e) were generated during the main deformation phase D1, which corresponds to F1 of Coward (1981, 1983) and to D1a of Freyer (in Hoffman et al., 1994). The folds are accompanied by a main penetrative slaty cleavage S1 in a roughly axial planar position. In the central and western part of the Lower Ugab Domain, a westward vergence of the folds is defined by a 20–70° eastward dip of axial planes and S1 cleavage (Figs. 2 and 6b). Gently dipping limbs, in normal position and at a small angle to S1 alternate with steep, commonly overturned limbs at a higher angle to S1 (Figs. 2b and 4a). The steep limbs usually exhibit parasitic minor folds that tend to be symmetric or weakly asymmetric (Figs. 2b and 4a) while the flat-lying limbs show few parasitic folds with strong asymmetry.

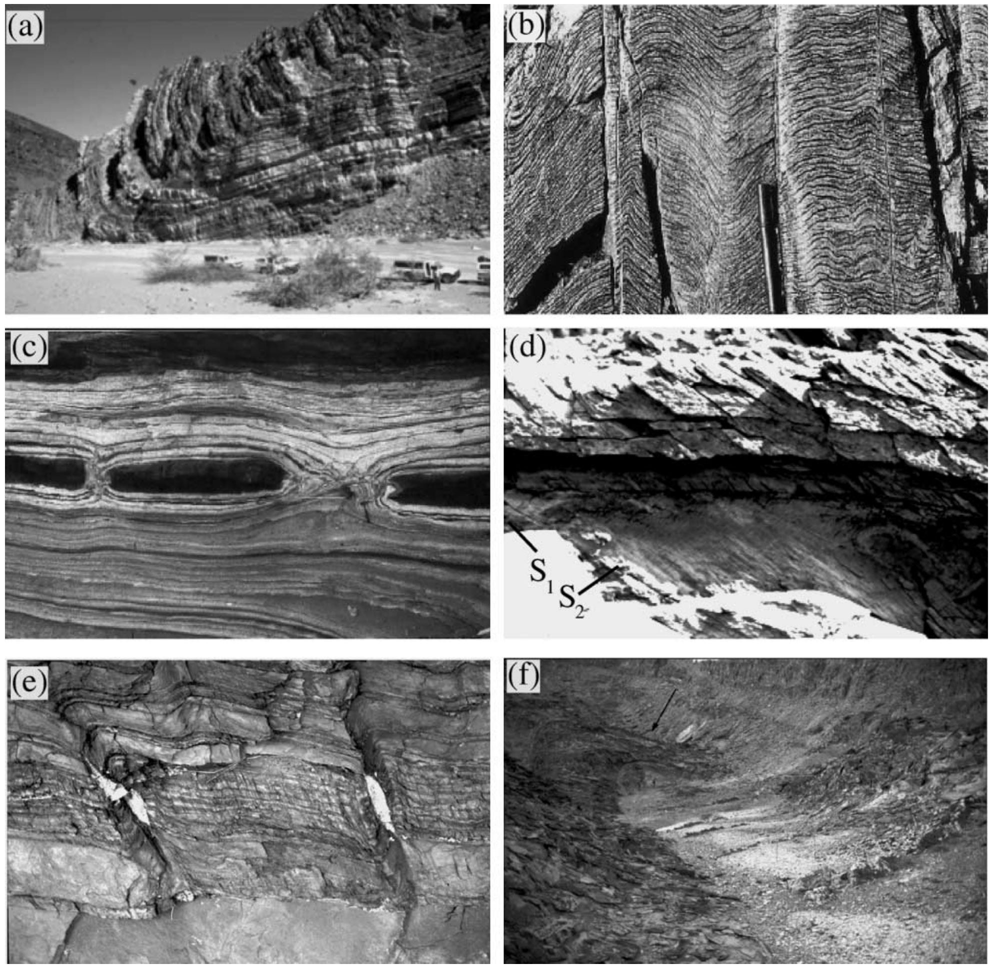

Fig. 4. (a) Typical D1-fold with E-dipping axial plane in carbonatic metaturbidites of the Gemsbok River Formation. View looking south. 14.0466°E, 20.9543°S. (b) Spaced S1 foliation folded in D2-folds; note the anastomosing nature of S1. 1 km E of the Voetspoor intrusion. (c) Boudinage of black pelitic layer in marble. 14.0372°E, 20.9689°S. (d) S1 cleavage in sandstone (top) and metapelite (bottom), overprinted by oblique S2 in the metapelite Width of view 1.5 m, viewing N. 14.05308°E, 20.9176°S. (e) D1-foliation boudins in a pelitic bed between psammite layers, shortened during D3. Typical is the trend of S3 (top left to bottom right) connecting small D3-folds that form the shoulder of the boudin neck on one side, and the tip of the core of the boudin neck on the other side. Compare with Fig. 9h. Width of view 1 m. 14.0849°E, 20.8702°S. (f) Quartz veins up to 100 m long normal to the axial plane of a D1-fold (arrow). The veins developed as a result of foliation boudinage similar to veins in (e). Compare with Fig. 9g. View looking south. 14.05308°E, 20.9176°S.

In the eastern part of the area the folds become upright and are locally eastward verging (cf. Hoffman et al., 1994), for example in the Goantagab valley west of the Voetspoor intrusion (Fig. 2c). The S1 cleavage in pelitic lithotypes has the morphology of a slaty cleavage defined essentially by subparallel white mica, chlorite and biotite grains. In most metasandstones no visible cleavage is developed. In calcsilicate rocks biotite and muscovite-rich domains separating carbonate-rich microlithons define a spaced, disjunctive cleavage of anastomosing character with millimetre to centimetre spacing. This latter cleavage is especially common in the eastern part of the area in the Amis River Formation (Fig. 4b). In many outcrops this is the only conspicuous foliation and could be mistaken for bedding by a casual observer.

Boudinage related to D1 deformation can be observed in the lower Rhino Wash and adjacent Ugab River valleys (Fig. 2); especially a black pelitic layer within the blue marble of the Gemsbok River Formation shows spectacular boudins with "fish-mouth" structure (Fig. 4c). Throughout the Lower Ugab Domain, strong foliation boudinage of S1 developed in pelitic layers on a decimetre to 100 m scale, with quartz veins filling most of the boudin necks (Fig. 4e and f). The boudins are probably late-syn-D1 after the foliation formed, since the quartz veins cut S1 but are not folded by D1. The boudin necks are roughly E–W oriented, indicating N–S extension approximately parallel to the D1-fold axes. Estimates for this extension based on the spacing of the boudins vary from 10% to 40%. No penetrative stretching lineations are developed, but deformed objects (intraclasts and concretions), boudinaged quartz veins and quartz strain fringes around pyrite crystals confirm that the principal extension direction of the bulk finite strain ellipsoid related to D1 was approximately N–S and subhorizontal, parallel to D1-fold axes. At one station in the Brak River valley (14.0242°E, 20.9983°S) deformed concretions in marble define a prolate strain ellipsoid with $X/Y = 3$ and $Y/Z = 1.6$ (average of 12 measurements). The bulk strain related to D1 can be estimated from restoration of the folding pattern (Fig. 2b and c) to have an approximately E–W shortening component in the order of 40–60%. This bulk strain must have deviated from plane strain where extension along the fold axes is evident.

Throughout the area isolated internally laminated quartz veins, mostly 1–3 cm but a few up to 10 cm thick, occur parallel to bedding. These veins, of a type known as striped bedding veins (Koehn and Passchier, 2000), consist of partly recrystallised fibrous quartz crystals subparallel to vein edges and bedding. Such veins are interpreted to have formed by bedding parallel slip (Koehn and Passchier, 2000). The fibres in the veins are close to orthogonal to D1-fold axes, which could be explained by flexural slip movements during D1. However, some veins continue uninterrupted and without a change in thickness or internal structure across the hinge zones of D1-folds. We envisage three possible scenarios to explain this situation. (i) The veins are related to pre-D1 bedding-parallel shear movements—in this case the orthogonal relationship between the fibres and D1-fold axes could be a coincidence; (ii) the veins result from inhomogeneous early syn-D1 shortening, predating D1-folding; (iii) the veins were produced by flexural slip during D1-folding, combined with fold hinge migration. The fact that no other indications for pre-D1 deformation such as ramps or repetitions of stratigraphic sequences were found makes the first scenario less probable.

4.2. Second deformation phase, D2

D2 is much less conspicuous than D1, producing only local folding of S1 and bedding (S0), mainly in open to gentle folds visible on micro- and meso-scale (Figs. 2c and 4b). This deformation phase corresponds to F2 of Coward (1981, 1983) and Miller et al. (1983) and to D1b of Freyer (in Hoffman et al., 1994). D2-fold axes have approximately the same orientation as D1 axes (Fig. 6e and g). D2-structures are unusual in that the axial planes and S2 do not have a preferred orientation throughout the area, but an orientation that is linked with that of S1. D2-folds are generally coaxial with respect to D1-folds (Fig. 6g) but invariably have axial planes at a high angle to S1 (Figs. 2c, 4d and 6c); if S1 is vertical, D2 axial planes are horizontal and vice versa. Along the axial planes of D2-folds a crenulation cleavage labelled S2 is locally developed, especially in pelitic layers. The morphology of S2 as seen in thin sections shows a complete transition from a well-developed differentiated crenulation cleavage to a disjunctive cleavage (Fig. 4d and 7a–c). Biotite-rich cleavage domains separating biotite-poor microlithons define S2. Apparently, growth of biotite porphyroblasts along S2 cleavage domains tends to mask progressively the crenulation cleavage microstructure. The fact that D1 and D2 are coaxial and acted under similar metamorphic conditions appropriate for the growth of biotite suggests that D1 and D2 were closely associated and probably continuous, resulting from a single tectonic event. Although we have no direct evidence it is conceivable that D2 formed in some localities while D1 continued to develop in others, resulting in a diachronous sequence.

4.3. Third deformation phase, D3

D1 and D2-structures are locally overprinted by microscopic to macroscopic scale D3-folds with ENE to NNE trending axes (Figs. 5e and 6h) and steep to subvertical axial planes, striking in the same direction (Figs. 2a and 6d; Coward, 1981, 1983; Miller et al., 1983). The orientation and intensity of D3 deformation is highly variable over the Lower Ugab Domain. Most D3-folds are asymmetric and occur as discontinuous fold trains, varying in amplitude along strike. S3 developed preferentially in tightly folded pelitic beds, usually as a spaced crenulation cleavage (Fig. 7b). In some samples S3 is selectively developed in differentiated S2 cleavage domains (Fig. 7b) similar to that described by Passchier and Trouw (1995, Section 4.11.2).

On a regional scale, the trace of the axial plane of open D3-folds with weak S3 development tends to be more EW-oriented than in tight folds with strong S3, which trend NE–SW (Fig. 2a). One of the best-exposed fold trains, with much higher D3 deformation intensity than in the surrounding rocks, is here named the Bushman fold train (Fig. 8). D1-folds and S1 are refolded in this kilometre-scale S-shaped D3-structure, the amplitude of which is highest in the centre of the structure and decreases to the SSW (Fig. 8). To the NE, the Bushman fold train wraps around the west-side of the Doros intrusion. This intrusion of hornblende syenite and younger biotite granite is of post-D1 to pre-D3 age (Figs. 2 and 8). The mean orientation of S3 within the centre of the Bushman fold train is 310/80 (dip direction), 15° different from the regional average 325/85 (Fig. 6d).

Another interesting feature is the post-D1 to syn-D3 Voetspoor intrusion, equally composed of hornblende syenite and younger biotite granite (Seth et al., in press; Vg in Fig. 2a). This intrusion cuts a major D1 syncline on its SW-side (Fig. 2a) that is deflected from the regional N–S trend to a NW–SE trend, associated with

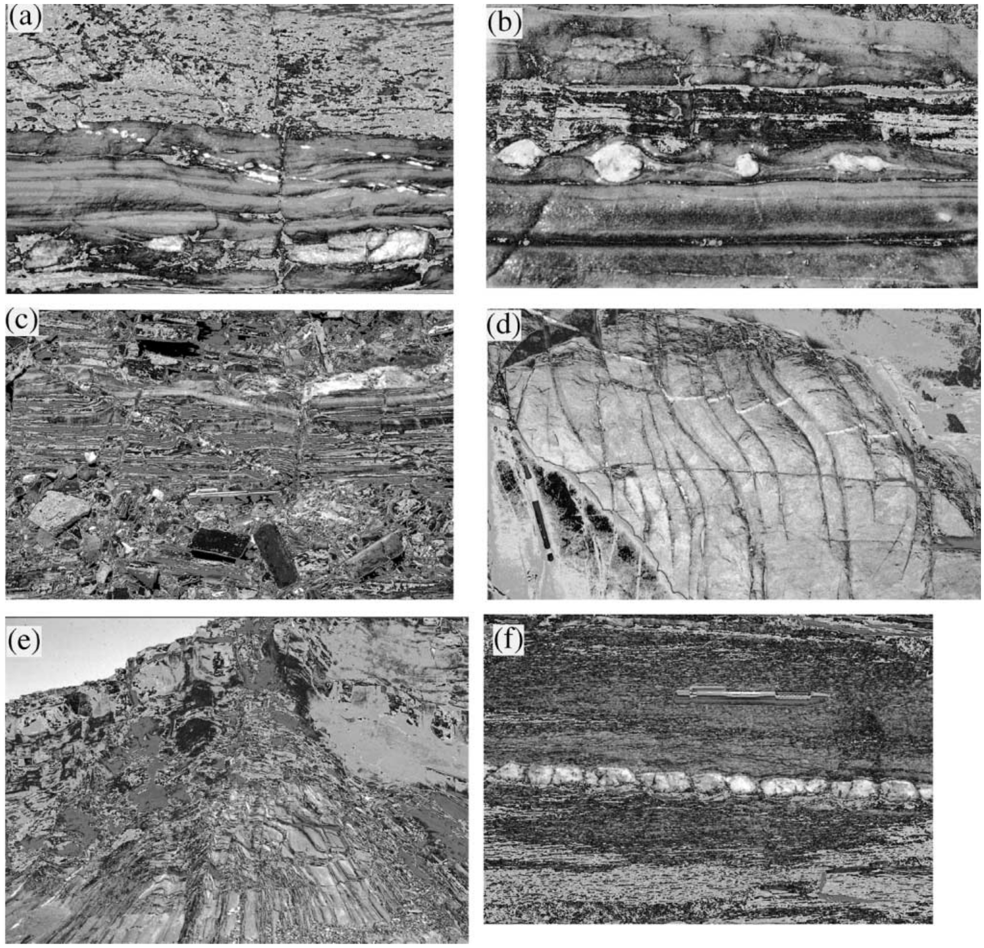

Fig. 5. (a) Quartz vein in marble of the Gemsbok River Formation is deflected and boudinaged in a shearzone parallel to the lithologic layering. Compare with Fig. 9a. Width of view 1.2 m. Brak River. (b) δ-type mantled porphyroclasts in marble of the Gemsbok River Formation, indicating sinistral shear sense. 14.1843°E, 21.0510°S. (c) Marble layer of the Gemsbok River Formation cut by quartz veinlets. The veinlets are boudinaged, and bedding is deflected into flanking folds alongside the veins (vein at left). Bedding and veinlets are subvertical. Compare with Fig. 9c. Width of view 80 cm. 14.1249°E, 20.8403°S. (d) Asymmetric tension gashes in the hinge zone of a D3-fold. 14.1716°E, 20.8923°S. (e) D3-fold in S1(bottom) and bedding (top). 14.1716°E, 20.8923°S. (f) Domino-type asymmetric boudins in a quartz vein in the D3-folds of (e). The boudinaged vein is slightly inclined towards a D3 crenulation on S1. The geometry of the boudins indicates sinistral shear sense.

D3-folds wrapping around the southern part of the intrusion (Fig. 2a). This arrangement suggests sinistral rotation of the intrusion during D3. The adjacent D1 syncline suffered drag with a sinistral sense of movement with respect to the regional trend of D1-folds, similar to that in small-scale n-type flanking folds (Passchier, 2001). The Doros pluton shows similar deflection of D1-folds along its SW contact and may have rotated in a similar way as the Voetspoor intrusion, but poor outcrop obscures the exact nature of the structure here.

The geometry of D3-structures like the Bushman fold train and folds around the Voetspoor intrusion, and the correlation of S3 foliation orientation and strain intensity suggest that D3 is a phase of sinistral non-coaxial flow with N–NW trending bulk shortening direction and steep NNE–SSW trending flow plane. Several small-scale D3-structures in the rocks of the Lower Ugab Domain are consistent with such sinistral non-coaxial flow:

(1) Ductile shear zones developed in subvertical limbs of D1-folds in marble layers of the Gemsbok River Formation (Fig. 2b and c). These shear zones have subhorizontal stretching lineation and subvertical planar shape fabric at a small angle to S0. Shear sense indicators, including σ- and δ-objects (Fig. 9b), shear bands and displaced boudinaged veinlets (Figs. 5 and 9a) all indicate sinistral movement. The shear zones are not overprinted by any younger structures although they are well foliated, but less deformed wall rocks have well-developed S2 and S3. We therefore interpret these sinistral shear zones as D3-features.

(2) Quartz-veinlets with sinistral flanking folds (Passchier, 2001) of S0 were developed in marble layers (Figs. 5 and 9c). The quartz veinlets have slightly

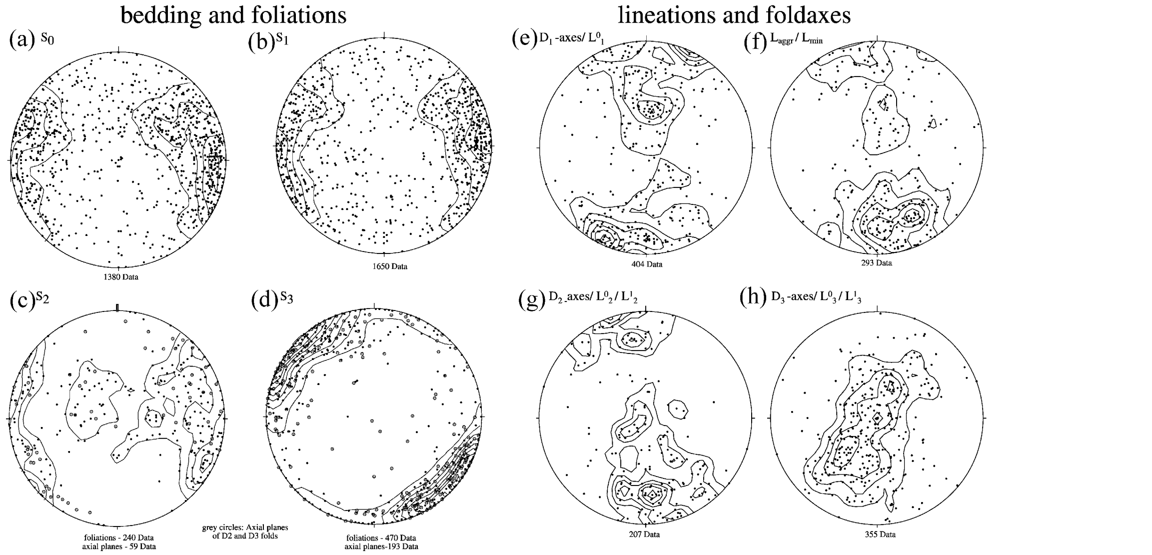

Fig. 6. Stereograms of the main structural elements in the Lower Ugab Domain; S0—bedding; S1, S2 and S3—cleavages and axial planes related respectively to deformation phases D1, D2 and D3. Lineations were measured in the Goantagab Domain. Data contoured at 1 2 3 4 times uniform.

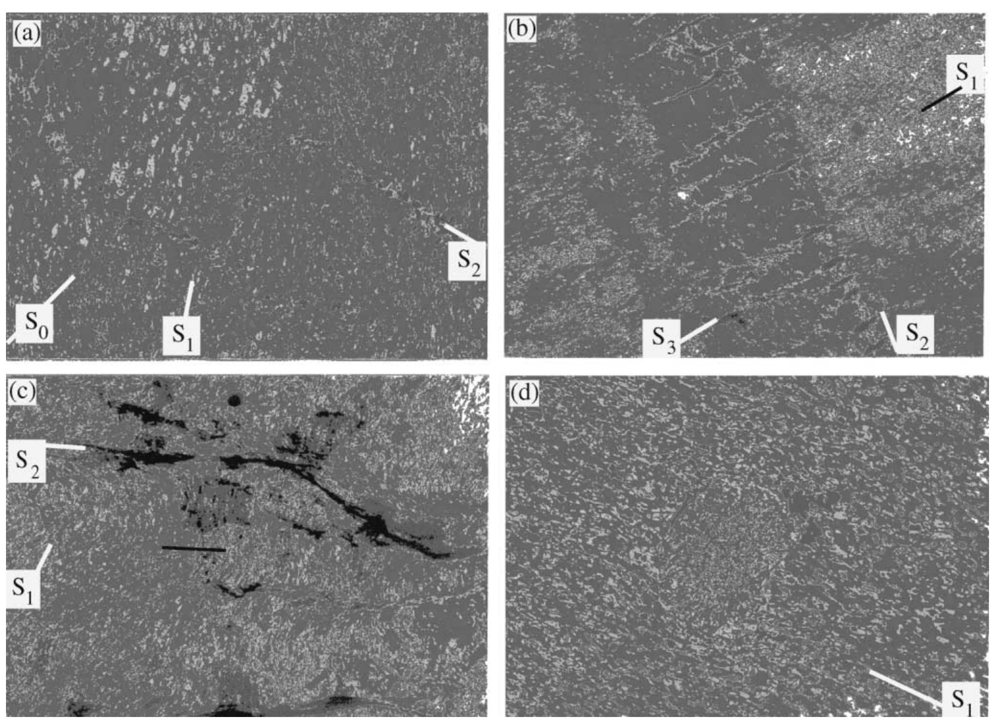

Fig. 7. (a) Incipient S2 cleavage development as seen in thin section; S0—top right to lower left; S1—vertical; S2—top left to lower right; S2 is a crenulation cleavage with concentration of biotite and opaque minerals along cleavage domains. Width of view 2 mm. Rhino Wash. (b) Photomicrograph showing selective S3 development in S2 differentiated cleavage band, caused by the different orientation of micas in these bands; S1—upper right to lower left; S2—dark band in centre of figure, from upper left to lower right; S3 parallel to S1. Width of view 2 mm. Upper Rhino Wash. (c) Photomicrograph showing biotite porphyroblasts that overgrew S2 crenulation cleavage; the concentration of biotite along S2 planes indicates that biotite grew late syn-D2. Width of view 1 mm. Brak River. (d) Photomicrograph of syn-D1 garnet with straight inclusion pattern at 90° with the S1 cleavage in the surrounding matrix, trending from upper left to lower right. The garnet must have grown relatively rapidly during D1, after incipient S1 had formed and before its rotation relative to S1 over about 90°. Width of view 2 mm. Brak River.

variable orientation, which is linked to either sinistral or dextral offset of bedding. This geometry is interpreted to result from vein development in conjugate extensional shear fractures due to N–S extension late during D1, and overprinting by sinistral D3 non-coaxial flow with a sinistral displacement component.

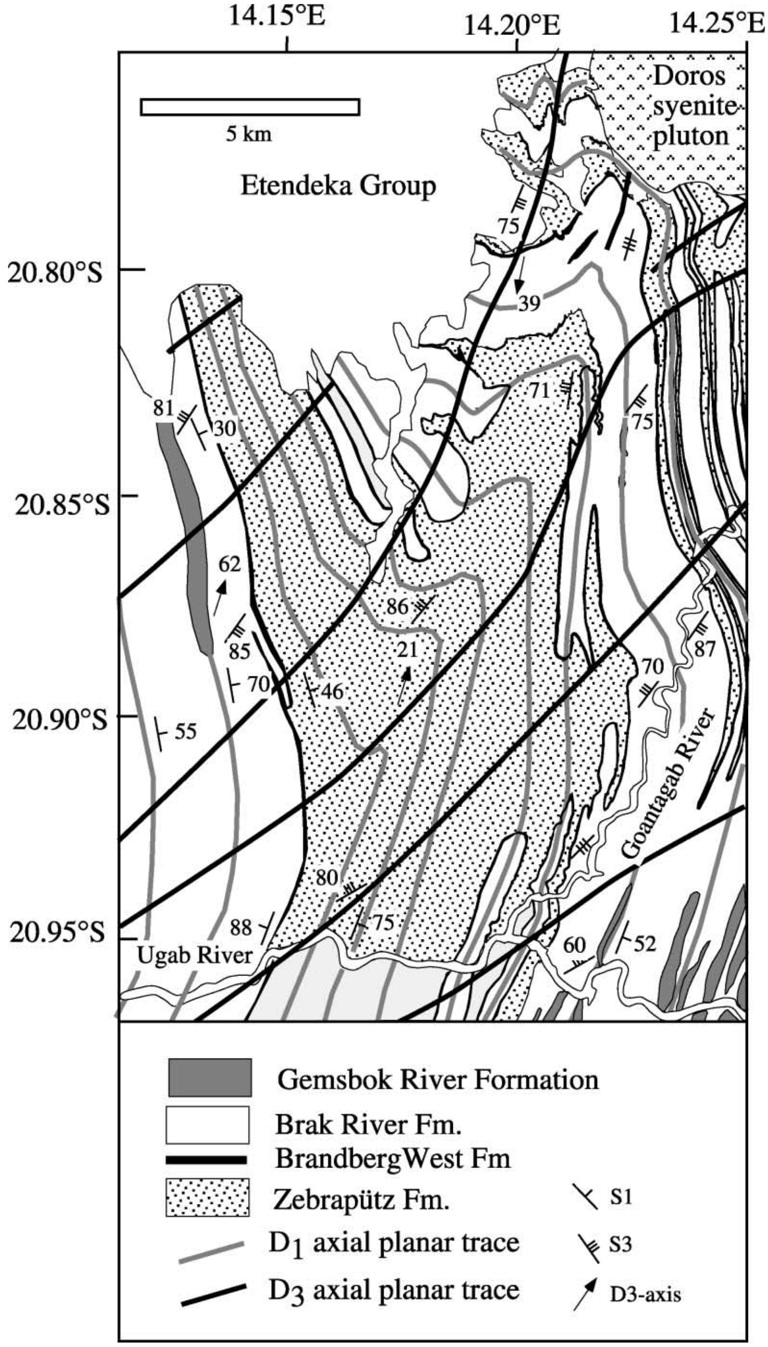

Fig. 8. Schematic map of the D3 Bushman fold train traced by mapping and from air photos. This kilometre-scale D3-fold appears from under the Etendeka flood basalts west of the Doros syenite pluton, wraps around the pluton and dies out towards the SW.

(3) Throughout the Lower Ugab Domain foliation boudins of S1 in steep limbs of D1-folds are shortened by D3 in a characteristic way, irrespective of the orientation of S1 and S0 (Figs. 4e and 9h). The intersection lineation between S1 and S3 is subparallel to the boudin axis. S3 is always slightly oblique to quartz veins in the boudin necks and lies axial planar to folds that involve the neck on one side of the boudin and the shortened shoulder of the boudin on the other side (Figs. 4e and 9h). These D3-folds apparently develop from the pre-existing curvature of the boudin neck zone. The geometry of the folds can be explained if originally a large angle existed between the layering and incipient S3 that progressively decreased during relative rotation of both elements during D3 (Fig. 9h). It seems likely that this consistent geometry formed in a sinistral ductile shear regime, although it is also possible in coaxial flow if the original variation in orientation of the boudins was very small.

(4) 1–10 cm thick quartz veins that developed in the necks of 100 m scale D1 boudins (Fig. 4f) are themselves

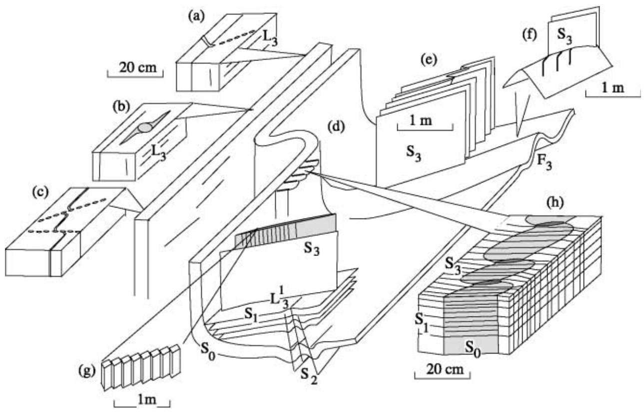

Fig. 9. Schematic diagram summarising the geometry and mutual relationship of minor asymmetric D3-structures overprinting a D1-fold. (a) Quartz vein in a marble layers is deflected and boudinaged in a shearzone parallel to the lithologic layering. Compare to Fig. 5a. (b) Mantled porphyroclasts indicate sinistral shear sense in marble layers. Compare to Fig. 5b. (c) Flanking folds are developed along quartz veinlets rotated by D3. Compare to Fig. 5c. (d) D3-folds are commonly asymmetric wit similar vergence. (e) Locally S3 is folded and a new cleavage S3b developed where S3 rotated into the shortening field. (f) Some D3-folds are obliquely transected by S3 and show asymmetric tension gashes in the hinge zone. Compare to Fig. 5d. (g) Some quartz veins truncate D1-folds at right angles, and develop asymmetric boudins during D3. Compare to Fig. 5f. (h) Boudins with quartz veins in the necks develop asymmetric folds with consistently similar vergence during D3. Compare to Fig. 4e. Further explanation in text.

boudinaged with vertical axes during D3. These boudins are asymmetric, with short, angular "domino-type" geometry (Figs. 5f and 9g) indicative of a sinistral shear sense (Swanson, 1992; Goscombe and Passchier, in press).

(5) Clockwise cleavage-transected folds (Johnson, 1991) on a metre scale occur in the D3 Bushman fold train (Fig. 9f). If cleavage developed later than initiation of the folds and if the folds rotated with the finite strain ellipsoid, this geometry is in agreement with a sinistral shear component (Borradaile, 1978; Soper, 1986; Pratt and Fitches, 1993).

(6) Brittle en-echelon sets of subvertical tension gradually occur in the hinge zone of several D3-folds. These sets are parallel to the hinge, and their geometry is indicative of a sinistral shear sense (Figs. 5d and 9f).

(7) SW of the Doros intrusion (Fig. 8, at 14.20298°E, 20.77883°S) S3 is affected by "D3b-folds" with vertical axes and an axial plane slightly inclined to the S3 enveloping surface as shown in Fig. 9e. These folds occur only here and have no regional significance. They can be explained if S3 was locally rotated into the shortening field of non-coaxial D3 flow such as could be expected in the D3 strain shadow of the Doros pluton; the geometry is similar to that of quarter folds adjacent to a porphyroclast (Passchier and Trouw, 1995).

4.4. Deformation in the Ogden Rocks and in the Goantagab Domain

The Ogden Rocks (Og in Fig. 2) are folded in kilometre-scale folds with NS trending axes and subvertical axial planes. A subvertical to moderately west dipping planar mylonitic shape fabric is usually parallel to the axial planes of these folds and contains a strong gently N-plunging stretching lineation. Shear sense indicators such as σ-shaped mantled porphyroclasts and asymmetric mica fish in the granitoid, and flanking folds (Passchier, 2001) in metasedimentary rocks indicate sinistral shear sense. Open upright folds trending SW–NE, which can be interpreted as related to D3, overprint these structures. Because of this relationship and the similarity in geometry of the folds, the pre-D3 mylonitic and fold structures in the Ogden Rocks can be classified as D1–D2 features. The intensity of deformation gradually decreases towards the east and 5 km from the Ogden Rocks no more stretching lineations are observed.

The mylonite zone of the Ogden Rocks is covered towards the north by the Etendeka Group. A similar mylonitic zone of sinistral transcurrent movement in the central Kaoko belt, known as the Purros Shear Zone (Hoffmann, 1987; Fig. 1), crops out north of the Etendeka Plateau, in line with the Ogden Rocks mylonites. The Purros Shear Zone is up to several kilometres wide and continuous over at least 400 km up to the Angolan border. On the aeromagnetic map of Namibia the Ogden Rocks and the Purros Shear Zone are connected by a linear anomaly. Brittle faults in the Etendeka Plateau also connect both zones and support the idea that both shear zone segments are part of a single structure which was remobilised by minor brittle faulting after deposition of the Etendeka Group.

Although our work was concentrated in the Lower Ugab Domain, preliminary study of structures in the Goantagab Domain led to the following conclusions. The rock units in the Goantagab Domain constitute a major D2 anticlinorium, subdivided into several synclines and anticlines on 2–3 km scale (Fig. 2a). 1–3 km west of the western limit of the Goantagab Domain, a N–S trending gently plunging stretching lineation (Fig. 6f) appears that increases in intensity towards the east, and is present throughout the Goantagab Domain. The lineation is parallel to the axes of metre to kilometre-scale D1-folds and mostly parallel to D2 axes, although it is locally folded by D2. S2 is present throughout the Goantagab Domain, and much more strongly developed than in the Lower Ugab Domain. D2-folds occur on a metre to kilometre-scale and refold D1-folds. D3 is relatively weak and only present locally. Shear sense indicators such as σ-shaped lenses, shear bands and flanking folds can be observed on horizontal surfaces parallel to the lineation in several steeply dipping marble horizons in the western Goantagab Domain; dextral and sinistral shear sense respectively are observed on alternating limbs of kilometre-scale D2-folds here. Shear sense in the transition zone between the Lower Ugab Domain and the Goantagab Domain, more precisely between the Amis River Formation and the equivalent to the Gemsbok River Formation in the Goantagab Domain, is inferred to be dextral. The arrangement of shear sense indicators and lineations in the marble horizons indicates flow parallel to bedding with the top units, i.e. the entire Lower Ugab Domain moving relatively to the north with respect to the lower units of the Goantagab Domain (cf. Goscombe and Trouw, 1999). A similar movement pattern is indicated by structures at Blauwpoort (20°34′S, 14°27′E) in the northern part of the Goantagab Domain. Marbles in this area contain recumbent D2-folds with subhorizontal N–S axes, parallel to a well-developed stretching lineation; clear S–C shear sense indicators show N-directed thrusting here.

4.5. Metamorphism

No detailed analysis has yet been made of metamorphism in the Zerrissene turbidite system but the following is known from field observations and thin sections. Metamorphic biotite is widespread in siliciclastic pelitic phyllites and actinolite–tremolite is common in calcsilicate layers. Locally garnet is present in phyllites indicating that, at these sites, the upper greenschist facies may have been attained whereas middle greenschist facies conditions prevailed elsewhere. Local occurrence of cordierite and andalusite porphyroblasts or their pseudomorphs and some biotite are related to contact aureoles of intrusive granites. No clear gradient of the regional metamorphism is apparent, but in the northern Rhino Wash valley (Rh on Fig. 2) biotite is absent in phyllites with muscovite and chlorite, suggesting an approximately E–W trending biotite isograd between this area and the southern Rhino Wash, Ugab and Brak River valleys (Br on Fig. 2). The metamorphic grade shows also a slight increase westwards towards the Ogden Rocks where granitoid rocks intrusive in metaturbiditic schists show mylonitisation under upper greenschist to lower amphibolite facies conditions as demonstrated by the presence of oligoclase in the metasedimentary rocks. In the Goantagab Domain (Fig. 1), the regional metamorphism is also of middle greenschist facies, as indicated by the generalised presence of biotite in metasandstones and phyllites and actinolitic amphibole in mafic rocks and in metadiamictites. At least at one site, in the Goantagab valley, garnet is also present in the metadiamictites.

Biotite occurs in the Zerrissene turbidites as small grains parallel to S1 and also as larger, intergrown crystals concentrated along differentiated S2. Many biotite porphyroblasts are late syn-D2 since they include open D2-folds, but also cause deviation of S2 (Fig. 7c). Garnet has straight inclusions interpreted to represent S1, at an angle of about 90° with the surrounding S1 cleavage (Fig. 7d). The garnet must therefore have grown relatively rapidly, early during D1, after an initial S1 had formed, but before the D1 deformation that caused its relative rotation. Even if the inclusions do not represent S1 but S0, rotation of at least 60° must have taken place since the angle between S1 and S0 in the same thin section is about 30°. It can be concluded that peak metamorphic conditions, represented by biotite and garnet growth, were attained during D1 and maintained during D2.

At most sites where an S3 crenulation cleavage was developed, no growth of new biotite occurred associated with S3, indicating that the regional metamorphic conditions during D3 had dropped below those of the biotite zone. However, at a few sites biotite grew along S3, possibly related to contact metamorphism caused by syn-D3 granite intrusions (e.g. close to the Voetspoor intrusion).

5. Discussion

5.1. Tectonic significance of D1–D2

Our observations concerning deformation and metamorphism in the rocks of the Lower Ugab Domain largely confirm earlier work by Coward (1981, 1983) and Fryer (in Hoffman et al., 1994). The three deformation phases recognised were apparently generated in two tectonic events: D1–D2 and D3. D1 and D2 seem to have been continuous and are attributed to a single tectonic event. At the presently exposed erosional level rocks in the stability field of biotite crop out throughout the entire Lower Ugab Domain. The general folding pattern of D1 shows it to be a phase of essentially E–W shortening but there are some additional features that need to be explained:

(1) D1 axial planes show a fanning structure (Fig. 2; see also Coward (1983, Fig. 8)).

(2) There is, at least in the central part of the area, a significant extension component parallel to D1-fold axes.

(3) Where D3-deformation is weak, D1-folds are unusually cylindrical and straight, even over an exposed length along the fold axis of 40 km (Fig. 2a).

(4) D2 axial planes and S2 are everywhere highly oblique to S1. In the fanning structure of S1, S2 is horizontal where S1 is steep, and steeply inclined where S1 is gently dipping.

(5) No evidence of thrusting was found within the Lower Ugab Domain. This means that a sedimentary succession in a domain of about 100×40 km with a measured exposed stratigraphic thickness of only 1.7 km

was shortened, folded and metamorphosed to the same grade, maintaining its lateral coherency.

Freyer (in Hoffman et al., 1994) interpreted D1–D2 as E–W shortening followed by a kind of gravitational collapse due to a hypothetical large overriding nappe thrust towards the east over a curved shear zone that corresponds to the Ogden Rocks mylonite zone (Fig. 1). In the present study no evidence was found to support this idea; D2 axial planes are not generally gently dipping (Fig. 6c) and the shear zones in the Ogden Rocks are steep and N–S trending with gently plunging stretching lineations, indicative of predominant N–S directed strike-slip movement.

According to Coward (1981, 1983) D2 was responsible for refolding of D1-structures around a common axis, and for N-directed thrusting of Neoproterozoic metasedimentary rocks over the Kamanjab inlier of the Congo Craton (ki in Fig. 1); in this scheme, D2 belongs to the same tectonic regime as D1 and is spatially restricted to the area SW of the Kamanjab inlier. Our observations are consistent with this model, with some modification. The pre-D3-structures in the Ogden Rocks and in the Goantagab Domain fit with relative N-directed movement of the Lower Ugab Domain units. The detachment zone related to this movement could be laterally continuous with the mylonitised rocks of the Goantagab Domain. The whole Goantagab Domain and the overlying lower part of the Amis River Formation with well-defined stretching lineations are actually interpreted as a massive antiformal shear zone with N-directed movement of the hanging wall. This explains the steep gradient in structural and stratigraphic changes between the Lower Ugab and the Goantagab Domains. Accommodation of this shear movement on the western side of the Lower Ugab Domain could well be expressed by the sinistral strike-slip Ogden Rocks shear zone (Fig. 10a).

E–W oriented fibres in folded striped bedding veins predating S1 indicate that during early stages of D1, E–W shortening was dominant. The foliation boudins cutting S1 show that the N–S extension component is of late D1 or D2 age. However, the alternation of shear sense on limbs of D2-folds and the occasional folding of stretching lineations by D2-folds in the Goantagab Domain suggest that E–W shortening was still active during D2 there. D1 and D2-folds and foliations are therefore produced by E–W shortening predating and contemporaneous with N–S extension (Fig. 10a). The cleavage fanning of S1 could be due to a vertical gradient in shortening during early D1, or may result from continuous relatively enhanced E–W shortening accompanying N–S extension in the Goantagab Domain below the present erosion level of the Lower Ugab Domain (Fig. 10a).

The local development of D2-structures oblique to S1 results from the complicated kinematic evolution during the D1–D2 event. Any late shortening at a high angle to S1 would have strengthened S1 and would go unnoticed, but shortening at a low angle to S1 would have caused development of younger folds (D2) and a new cleavage, S2. D2-folds and S2 cleavage, at a high angle to S1, apparently formed during N-directed movement of the Lower Ugab Domain accompanied by E–W shortening, during or after development of the D1 cleavage fan (Fig. 10a). Gently dipping S1 surfaces would have developed steep D2-structures because of lateral E–W shortening, while domains with steeply dipping S1 could have developed flat S2 planes due to activity of the gently dipping shear zone, or to collapse during reduced E–W shortening.

The development history of structures in the Lower Ugab Domain is therefore one of initial (early D1) E–W shortening that graded into predominant N-directed shear movement combined with minor E–W shortening (Fig. 10a). Such a sequence of events is unusual but not unique. A very similar interpretation was given to structures in the Dalradian in Scotland where two phases of deformation have a common axis, but show near orthogonal foliations (Treagus, 1999). The presence of a stretching component parallel to the fold-axes, as is the case for D1-folds, has also been proposed as a characteristic feature of lateral constriction in an active

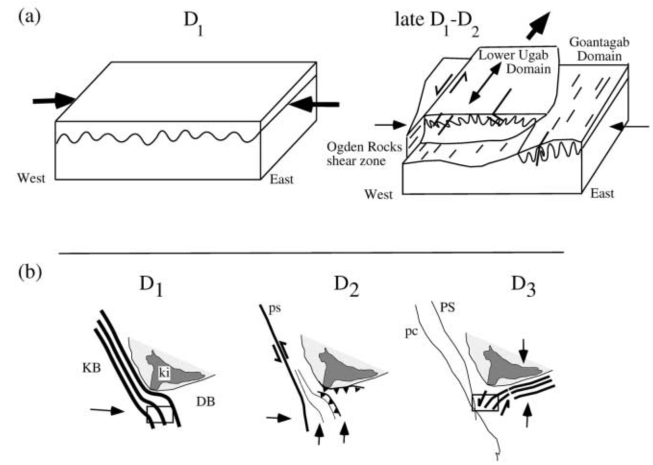

Fig. 10. (a) Schematic presentation of a possible sequence of events during D1–D2. Initial EW shortening of the turbidites is followed by a more complex history of NS extension and EW shortening. This is associated with reorientation of D1-folds and local development of D2-folds and S2 (bold lines) at a high angle to S1. (b) Regional tectonic sequence of events, partly after Coward (1981, 1983) and adapted for data from the Lower Ugab Domain (square). Initial contraction during D1 in the Kaoko- (KB) and Damara Belts (DB) against the Congo Craton is followed by more restricted shear movement to the north associated with D2; strike-slip movement in the Kaoko Belt is apparently of this age. During D3, contraction is restricted to the Damara Belt and sinistral D3 transpression in the Lower Ugab Domain may be due a decrease in contraction towards the west. ki—Kamanjab inlier of the Congo Craton. ps—Purros shear zone. pc—present coastline.

shearzone strongly deviating from plane strain, as in transpressional and transtensional regimes (Soper et al., 1987; Passchier, 1998; Tikoff and Peterson, 1998).

5.2. Interpretation of D3

Throughout the area, D3-folds differ considerably from D1–D2-structures in orientation and kinematic association. The metamorphic grade dropped in most places from biotite zone conditions during D1–D2 to sub-biotite zone conditions during D3. D3-folds and associated structures can be interpreted as the result of sinistral non-coaxial progressive deformation with an approximately vertical rotation axis and a steep NNE–SSW trending flow plane. D3 deformation is highly partitioned into domains of high strain, such as the Bushman fold train, and domains of low strain, suggesting that basement may have been involved in the movement, possibly along deep-seated ductile shear zones.

5.3. Regional tectonics

The scheme of deformation events presented above is consistent with reconstructions of the tectonic evolution of north-western Namibia given by Miller (1983), Prave (1996), Dürr and Dingeldey (1996), Seth et al. (1998, in press) and Stanistreet and Charlesworth (2001). According to Coward (1981), kilometre-scale D1-folds in the Kaoko belt extended into the Zerrissene turbidites and wrapped around the edge of the Kamanjab basement inlier towards the east as shown in Fig. 10b (Coward, 1981). The curvature may be due to an original embayment in the shelf. D2-structures only developed at the SW corner of the Kamanjab inlier, probably in response to a gradual change in the tectonic setting (Coward, 1981; Fig. 10b). The effect during D2 was N-directed shear movement of the Lower Ugab Domain over the Goantagab Domain while being shortened in E–W direction, and thrusting of the Goantagab Domain to the north onto the Congo Craton (Fig. 10b). A possible cause for the sequence and combination of E–W shortening and N-directed shear movement include initial shortening across the Kaoko belt due to the collision of the Congo and Rio de la Plata cratons, followed by the partially contemporaneous collision of the Congo and Kalahari cratons (Fig. 10b; cf. Prave, 1996). The development of the Purros-Ogden Rocks strike-slip shear zone may also have played a role (Fig. 10b).

D3 deformation is restricted to the Damara belt and the southern part of the Kaoko belt (Fig. 10b). According to Coward (1981) it decreases in intensity to the west, and in the Lower Ugab Domain it is indeed mostly weak and strongly partitioned. At least in the Lower Ugab Domain, D3 is not a phase of simple coaxial shortening but associated with sinistral non-coaxial flow, probably transpression. No such transpressional features are reported further east (Coward, 1981). This may be due to lack of data, but it is also possible that the transpressional nature is unique for the western zone. D3 sinistral transpression could well fit N–S shortening in the Damara belt, which at its western edge transfers into sinistral transpression to accommodate decreasing N–S shortening in that direction (Fig. 10b).

The absolute age of deformation in the Zerrissene turbidite system is not yet clear. The turbidites have been correlated with the Swakop Group (Miller et al., 1983) with a depositional age between 750 and 540 Ma (Miller, 1983). D1-folds are transected by syenite of the Voetspoor intrusion that has been dated by Seth et al. (in press) as 53 ± 03 Ma by Pb–Pb single zircon evaporation technique. It is interesting that all dated granitoid intrusions in the Kaoko belt north of the Etendeka cover are older than 550 Ma, the last of which are still synkinematic (Seth et al., 1998). An age range of 560–550 Ma is also typical of syn- to late kinematic granitic magmatism in the Damara Belt to the east of the Zerrissene turbidites (Miller, 1983; Seth et al., 1998). This implies that D1–D2 deformation in the Lower Ugab Domain turbidites may be contemporaneous to deformation in the northern parts of the Kaoko belt (Seth et al., 1998), but that deformation may have continued longer in the south.

An attempt was made to date different generations of biotite from the Lower Ugab Domain using the Ar–Ar method. First results show that all biotite in the Lower Ugab Domain shows cooling ages in the range 497–530 Ma, irrespective of the relative age or location of the biotite (Gray, personal communication). This implies that the entire area probably remained above the blocking temperature of biotite until D3, possibly related to the intrusion of the syenite plutons and that therefore only the later part of the deformation history can be dated with the Ar–Ar method.

5.4. Comparison with the Dom Feliciano belt

The counterpart of the Kaoko belt in South America, the Dom Feliciano belt in southeast Brazil and Uruguay has a general vergence to the W or NW (Basei et al., 2000). From the inland Rio de la Plata Craton (Fig. 1) to the coast it consists of deformed foreland basins; a schist belt with polyphase deformation and greenschist to lower amphibolite facies metamorphism; and a granite belt. The latter is composed of several aligned calc-alkaline batholiths, dated between 620 and 590 Ma, and interpreted as the root of a continental magmatic arc, active in the period 620–600 Ma that collided with the schist belt in the interval 600–590 Ma (Basei et al., 2000). If the Dom Feliciano belt is joined with the Kaoko belt a bivergent belt results with the magmatic arc in the centre. The main difference between the two belts is their age. Deformation and metamorphism in the Dom Feliciano schist belt is considered to be pre-620 Ma, whereas deformation and metamorphism in the Kaoko belt are estimated to be approximately 560–530 Ma.

According to Porada (1989) the main subduction in the Dom Feliciano belt was towards the west, while sedimentation on the African side would be on a passive or intraplate margin. The schist belt would be at least partly a back-arc basin in this model and the different ages could be explained by an early collision of the arc against its back-arc region (600 Ma), followed by the final closing of the oceanic basin around 550 Ma, resulting in the tectonics of the Kaoko and Gariep belts (Fig. 1). An alternative model, proposed by Basei et al. (2000) postulates eastward subduction between the schist belt and the arc. In this model, largely based on Sm–Nd signatures, the main suture would lie west of the arc that would be installed on the margin of the Congo and/or Kalahari cratons. According to Basei et al. (2000) the Kaoko and Gariep belts would represent essentially back-arc basins. Early collision (600 Ma) of the Rio de la Plata craton with the arc would be followed by later closure of the back-arc (550 Ma). Our results seem to fit the model of Porada (1989) better, mainly because the turbidites and paleocurrents point to passive margin sedimentation from a northeastern source and no evidence for back-arc sedimentation associated with volcanic material was encountered. Further work in the northern Kaoko belt could help to decide this issue.

6. Conclusions

In the Lower Ugab Domain turbidites, D1 and the less conspicuous D2 deformation phase are part of a complex system of deformation that includes a change from E–W compression at the onset of D1 to N-directed hanging-wall transport with N–S extension, combined with E–W shortening during D2. The D1–D2 tectonic event is probably associated with shortening due to collision across both the Kaoko and the Damara belts. D3 is significantly different from D1 and D2 because of its deviant orientation and lower metamorphic grade. It is considered to be related to final adjustments in the Damara belt associated with the collision between the Congo and Kalahari Cratons, slightly postdating events in the Kaoko belt. D3-structures illustrate that transpression in slate belts may be recognised by other deformation structures than cleavage-transected folds (Figs. 9 and 4e–h), including asymmetric fold-train domains on the map-scale, and small-scale structures such as those shown in Fig. 9. It may therefore be useful to search for this kind of structures in slate belts with lesser outcrop, since their presence can indicate the activity of more complex kinematic regimes than at first sight assumed.

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References

Basei, M.A.S., Siga Jr., O., Masquelin, H., Harara, O.M., Reis Neto, J.M., Preciozzi, P., 2000. The Dom Feliciano Belt of Brazil and Uruguay and its foreland domain, the Rio de la Plata Craton, framework, tectonic evolution and correlation with similar provinces of southwest Africa. In: Cordani, U.G., Milani, E.J., Thomaz Filho, A., Campos, D.A. (Eds.), Tectonic Evolution of South America, pp. 311–334, 31, IGC Rio de Janeiro.

Borradaile, G.J., 1978. Transected folds a study illustrated with examples from Canada and Scotland. Bulletin of the geological Society of America. 89, 481–493.

Coward, M.P., 1981. The junction between the Pan-African mobile belts in Namibia its structural history. Tectonophysics 76, 59–73.

Coward, M.P., 1983. The tectonic history of the Damaran belt. In: Miller, R.McG. (Ed.), Evolution of the Damara Orogen. Special Publication of the Geological Society of South Africa, vol. 11, pp. 409–421.

Dürr, S.B., Dingeldey, D.P., Okrusch, M., Franz, L., 1995. Transpression in the Panafrican Kaoko belt of northern Namibia structural and petrological implications. Zentralblatt Geologie Paläontologie 1996, 495–505.

Dürr, S.B., Dingeldey, D.P., 1996. The Kaoko belt (Namibia): Part of a late Neoproterozoic continental-scale strike-slip system. Geology 24, 503–506.

Fernandes, L., Tomassi, A., Porcher, C., 1992. Deformation patterns in the southern Brazilian branch of the Dom Feliciano Belt: a reappraisal. Journal of South American Earth Science 5, 77–96.

Goscombe, B., Trouw, R.A.J., 1999. The geometry of folded tectonic shear sense indicators. Journal of Structural Geology 21, 123–127.

Goscombe, B., Passchier, C.W. Asymmetric boudins as shear sense indicators—an assessment from field data. Journal of Structural Geology, in press.

Hoffmann, K.H., 1987. Application of tectonostratigraphic terrane analysis in the Damara province of central and northern Namibia. In: Hartnady, C. (Ed.), Proceedings of the Alex du Toit Golden Jubilee conference, Cape Town, pp. 25–27.

Hoffman, P.F., Swart, R., Freyer, E.E., Guowei, H., 1994. Damara Orogen of Northwest Namibia. In: Niall, M., McManus, C. (Eds.), Geological Excursion Guide of the international conference Proterozoic crustal and metallogenic evolution. Geological Society and the Geological Survey of Namibia. 55 pp.

Johnson, T.E., 1991. Nomenclature and geometric classification of cleavage-transected folds. Journal of Structural Geology 13, 261–274.

Koehn, D., Passchier, C.W., 2000. Shear sense indicators in striped bedding veins. Journal of Structural Geology 22, 1141–1151.

Kröner, A., 1981. Precambrian plate tectonics. In: Kröner, A. (Ed.), Precambrian Plate Tectonics. Amsterdam, Elsevier.

Kukla, P., Stanistreet, I., 1991. Record of the Damaran Khomas Hochland accretionary prism in central Namibia: refutation of an ensialic origin of a late Proterozoic orogenic belt. Geology 19, 473–476.

Miller, R.McG., 1983. The Pan-African Damara Orogen of Southwest Africa/Namibia. In: Miller, R.McG. (Ed.), Evolution of the Damara Orogen. Special Publication of the Geological Society of South Africa, vol. 11, pp. 431–515.

Miller, R.McG., Freyer, E.E., Hälbich, I.W., 1983. A turbidite succession equivalent to the entire Swakop Group. In: Miller, R.McG. (Ed.), Evolution of the Damara Orogen. Special Publication of the Geological Society of South Africa, vol. 11, pp. 65–71.

Miller, R.McG., Grote, W., 1988. Geological map of the Damara Orogen of South West Africa/Namibia 1:5.000.000. Geol. Surv. Namibia, Windhoek.

Passchier, C.W., 1998. Monoclinic model shear zones. Journal of Structural Geology. 20, 1121–1137.

Passchier, C.W., 2001. Flanking structures. Journal of Structural Geology 23, 951–962.

Passchier, C.W., Trouw, R.A.J., 1995. In: Microtectonics. Springer, Verlag, 325 pp.

Porada, H., 1989. Pan-African rifting and orogenesis in southern to equatorial Africa and eastern Brazil. Precambrian Research 44, 103–136.

Pratt, W.T., Fitches, W.R., 1993. Transected folds from the western part of the Bala Lineament Wales. Journal of Structural Geology 15, 55–68.

Prave, A.R., 1996. Tale of three cratons: Tectonostratigraphic anatomy of the Damara orogen in northwestern Namibia and the assembly of Gondwana. Geology 24, 1115–1118.

Seth, B., Kröner, A., Mezger, K., Nemchin, A.A., Pidgeon, R.T., Okrusch, M., 1998. Archaean to Neoproterozoic magmatic events in the Kaoko belt of NW Namibia and their geodynamic significance. Precambrian Research 92, 341–363.

Seth, B., Okrush, M., Wilde, M., Hoffmann, K.H. The Voetspoor intrusion, Southern Kaoko Zone, Namibia: mineralogical, geochemical and isotopic constraints for the origin of a syenitic magma. Henno Martin Memorial Volume of the Geological Survey of Namibia, in press.

Soper, N.J., 1986. Geometry of transecting anastomosing solution cleavage in transpression zones. Journal of Structural Geology 8, 937–940.

Soper, N.J., Webb, B.C., Woodcock, N.H., 1987. Late Caledonian (Arcadian) transpression in north-west England: timing, geometry and geotectonic significance. In: Proceedings of the Yorkshire Geological Society, vol. 46, pp. 175–192.

Stanistreet, I.G., Charlesworth, G., 2001. Damaran basement-cored fold nappes incorporating pre-collisional basins Kaoko Belt Namibia and controls on Mesozoic supercontinental break-up. South African Journal of Geology 104, 1–12.

Swanson, M.T., 1992. Late Acadian-Alleghenian transpressional deformation: evidence from asymmetric boudinage in the Casco Bay area coastal Maine. Journal of Structural Geology 14, 323–341.

Swart, R., 1992. The sedimentology of the Zerrissene turbidite system, Damara Orogen, Namibia. In: Memoir of the Geological Survey of Namibia, vol. 13, 54 pp.

Treagus, J.E., 1999. A structural reinterpretation of the Tummel Belt and a transpressional model for the evolution of the Tay Nappe in the central Highlands of Scotland. Geological Magazine 136, 643–660.

Tikoff, B., Peterson, K., 1998. Physical experiments of transpressional folding. Journal of Structural Geology 20, 661–672.