Superposed folding at the junction of the inland and coastal belts, Damara Orogen, NW Namibia

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Two adjoining dome structures in Neoproterozoic Otavi Group sediments are located at the intersection of the Otjojo (inland branch) and Kaoko (coastal branch) fold and thrust belts of the Damara Orogen on the farm Vrede, northwestern Namibia. Systematic mapping of outcrop-scale cleavage and folding relationships has unraveled three temporally distinct folding events in the two dome structures. In present-day coordinates, the first shortening event (D₂) produced E-W trending folds. The second contractional phase (D₃) developed N-S trending folds. The third shortening episode (D₄) featured a renewed production of E-W trending folds. If this chronology of deformation is applied to similarly oriented structures along the inland and coastal belts, a history of ocean closure and the amalgamation of southwestern Gondwanaland can be inferred. Closure of the Adamastor ocean and collision along the Kaoko margin began before closure of the Khomas sea and collision along the inland belt.

Introduction

The Vrede domes are a pair of doubly plunging anticlines located on Vrede farm, at the junction of the Pan-African age (650-450 Ma) Otjojo fold and thrust belt (Northern Zone of the inland branch in Miller, 1983) and Kaoko fold and thrust belt (Central Kaoko Zone of the coastal branch in Miller, 1983; Fig. 1). Domes have been recognized as a prevalent feature of the Otjojo thrust belt (Frets, 1969; Miller, 1983; Weber et al., 1983) and Swakop zone (Smith, 1965, Barnes and Downing, 1979; Miller, 1983; Kröner, 1984; Oliver, 1995; Fig. 1). Kröner (1984) proposed that late-tectonic, anatexically produced granites intruded and ballooned into pre-existing anticlines in the cover sediments of the Swakop zone, producing domes. Further north in the area of the Vrede domes, the Otjojo thrust belt is riddled with post-tectonic granites (520-500 Ma) and the diapir mechanism for dome formation appears plausible. Superposed tectonic folding has also been evoked as a mechanism for doming in the Swakop zone (Smith, 1965) and Otjojo thrust belt (Frets, 1969; Miller, 1983). In theory, two orthogonally-oriented fold trains should interact like waves to form basin and dome fold interference patterns (Ramsay, 1958, 1962; Weiss, 1959; To-bisch, 1966; Ghosh, 1970; Wathkinson, 1981; Theissen, 1986; Ramsay and Huber, 1987; Lisle et al., 1990). In the Swakop zone, the northward extension of outcrop-scale refolded structures has usually led to the adoption of the diapir model (Barnes and Downing, 1979; Kröner, 1984). A third mechanism for dome formation has been proposed for the Swakop zone involving oblique collision and lateral tectonic escape along mid-crustal ductile detachments (Oliver, 1995). This model is probably not applicable to the lower-grade rocks of the Otjojo thrust belt.

The Vrede domes are low-grade, thin-skinned structures containing a plethora of outcrop-scale refolded folds and multiply deformed axial-planar cleavages. The goal of this contribution is to deconvolve the refolded folds of the domes into chronologically distinct components of finite strain. Four models are presented to explain the structural evolution of the domes. The chronology of deformation in the domes is then used to speculate on the relative timing of ocean closure along the inland and coastal belts during the amalgamation of southwestern Gondwanaland.

Geologic Setting

The Vrede domes consist of two northeast verging domal anticlines that are separated by an east-west trending keel-shaped syncline (Fig. 2). The domes owe their ring-like appearance in plan view to a drainage network that follows the weakest Otavi stratigraphy (Ghaub Formation and Lower Ombombo Subgroup silstones), eroding circular, low-order channels that eventually feed higher order ephemeral streams which drain into the Huab River. The domes are each about 1.5 km wide (east-west), and together they are about 4 km long (north-south).

The domes lie in the Huab River channel, which drains the southwest heel of the Kamanjab Inlier (1.7-2.0 Ga basement; Fig. 1). Creaceous volcanics related to the opening of the Atlantic Ocean occur west and south of the Vrede domes and post-tectonic (~500 Ma) granites border them to the southeast (Fig. 1).

Regional Tectonostratigraphy

The Otavi Group is a carbonate-dominated passive-margin succession bordering the southern promontory of the Congo Craton (of which the Kamanjab Inlier is a part). The megasequence records the ~750 Ma breakup of the supercontinent Rodinia, a stable platform with two discrete glacial intervals capped by distinctive Neoproterozoic carbonate (Fairchild, 1993; Schmidt and Williams, 1995;
Kennedy, 1996; Hoffman et al., 1998a; Kennedy et al., 1998), and the $\sim$550 Ma amalgamation of Gondwanaland (Fig. 3b; Hoffman et al., 1998D). The Otavi Group is underlain by pre-760 Ma Nosib Group syn-rift clastics and a low grade, 2.0-1.7 Ga basement complex. It is overlain by pre-540 Ma Mulden Group foreland basin siliciclastics (Frets, 1969; Hoffman et al., 1998D).

The Otavi passive margin is segmented into basins by transverse basement ridges (Henry et al., 1990). Henry et al. (1990) and Stanisstreet et al. (1991) have suggested that much of the early rifting was controlled by large, low-angle detachments that exploited older, Mesoproterozoic structures. The Otavi rocks of the Vrede domes are located south of the Huab Ridge, the southernmost transverse basement ridge, and south of Rockeys Fault (Fig. 1). Henry et al. (1990) proposed that the Rockeys fault is the northernmost extensional detachment and the postulated shelf-slope transition bounding the Otavi platform to the north.

**The Stratigraphy of the Vrede domes**

In the Vrede domes, the Otavi Group is a 400+ m sequence of carbonates and siliciclastic metasediments, variably folded and metamorphosed to lower greenschist facies. Figure 2 is a geologic map of the Vrede structures, and Figure 3 provides stratigraphic columns for each dome.

The Ombombo Subgroup forms the bulk of the Vrede stratigraphy. The southern dome is cored by 50+ m of Ombombo-1 Formation (Fm.) black-maroon limestones (partially dolomitized) and sandstones. The limestones/dolostones of this unit form the resistant ridges that contain metre-scale refolded folds at location M9 (Fig. 2). A 15+ m thick lens of dinitite/tektite lies in the hinge of a 150 m wavelength, E-W trending fold at M9. The northern dome shows less structural relief than the southern dome and does not expose Ombombo-1 Fm. rock.

The Ombombo-2 Fm. varies significantly across the domes. In the northern dome, the Ombombo-2 Fm. is a 100+ m thick unit of nearly continuous coarse-grained, polymictic conglomerate (Maloof, 1998), broken only by thin sandstone beds and rare dolostone ribbon beds. In the southern dome, the Ombombo-2 Fm. contains thin, medium-grained conglomerate beds separating thicker beds of fine-medium sandstone and dolostone ribbons.

The Ombombo-3 and Ombombo-4 Fms. are lithologically consistent across the domes, displaying two or three parasequences containing pinkish red *Tungusia* stromatolite biostromes (Semikhov, 1962) with ooids and dolostone ribbons. The silicified ooid beds lie parallel to bedding and serve as useful marker horizons. A volcanic ash layer near the top of Ombombo-3 in a stratigraphically correlative section 150 km to the north has a U-Pb zircon age of $759 \pm 3.5$ Ma (Fig. 3; Hoffman et al., 1998D).

The Abenab Subgroup is extremely thin in the Vrede domes. The Chuos Fm. glaciogenic dinitite marks the base of the Abenab Subgroup and rests unconfornably on the Ombombo-4 Fm. Without the presence of the overlying Rasthof Fm., the Chuos Fm. cannot be distinguished conclusively from the overlying Ghaub dinitite of the Tsumeb Subgroup. However, where the Rasthof Fm. is present, the Chuos Fm. is a pink-grey, carbonate-clast, carbonate-matrix dinitite with very rare striated basement
clasts. For the Rasthof Fm., only the lowermost 4 m of black, finely laminated rythmites with rollover algal mats was preserved prior to the downcutting of the Ghaub glacia
tion (Fig. 3).

The Ghaub Fm. marks the base of the Tsunam Sub-
group and cuts into the Abenab and Ombombo Subgroups creating an unconformable contact with 6+ m of local re-
lief. The Ghaub Fm. is a glaciomarine diamicitic charac-
terized by large (up to 2 m diameter) carbonate and base-
ment granitoid/gneiss dropstones within a fine silstone or carbonate matrix. In the northern dome, the unit contains
only rare <5 cm diameter dropstones in the silstone diamicritic, while in the southern dome, the unit is packed with large and lithologically diverse dropstones.

Along the western edge of the Vrede domes, an up to
2.8 m thick green ash bed occurs between the Ghaub and the Maieberg Formations. The ash is silicified and fer-
ruginated, perhaps because of its proximity to the Mulden-
Tsunam exposure surface. Four zircon grains were sepa-
rated from the ash unit for U-Pb geochronology. Unfortu-
nately, the grains gave upper concordia intercept ages of
1.7-1.9 Ga and lower intercept ages of 0.51-0.52 Ga (S.A.
Bowring, pers. com., 1998). The zircons are probably de-
trital grains that were affected by Pan-African Pb-loss.

Where the ash unit is not present, the Ghaub Fm.
grades almost conformably into Maieberg Fm. dolostone
rythmites characterized by distinctive undulating waves of
0.5-2 cm wide isopachous cement, 10-200 cm in wave-
length and 5-100 cm in amplitude. The Maieberg cement
layer grades into variably dolomitized limestone rythmites,
followed by massive recrystallized dolomite with calcite
vugs up to 1 m in diameter.

The outermost shell of the Vrede domes consists of
Mulden Gp. silstones, phylites, quartzites, feldspathic
quartz arenites, quartz-pebbie conglomerates and dolomite-
clast breccias. A highly ferruginated and silicified subsala-
unconformity is present where the Mulden Group truncates the Maieberg Formation.

The Structure of the Vrede domes

At the map-scale, 1-2 km wavelength folding of the thick and competent Ombombo 2, 3, 4 Fm. conglomerates and dolostones created a 1.5 km wide and 4 km long pair of lobate domes separated by a keel-shaped synclinal basin (Fig. 2). Associated Otavi Group silstones and fine-
grain sandstones are deformed more passively around folded dolostone and conglomerate horizons, leading to map-scale layer thickness variations in the less competent units. The less competent silstone and limestone units contain the metre-scale, multiply deformed folds and cleavages that are the subject of the remainder of this paper.

Cleavage

The Otavi Group carbonates rarely preserve a cleav-
age, and when they do, it is spaced, discontinuous and of-
ten just a series of stylolitic discolorations (Alvarez et al.,
1976). The carbonates never contain two cleavages in hand sample. Fortunately, the Ombombo carbonates are closely associated with silstones and sandstones that con-
tain one and often two sets of penetrative cleavage. In
some localities, the intersection of the two cleavages leads
to typical pencil-shaped debris (e.g. M9). In other loca-
tions, the cleavages interact to create an older, tightly
spaced crenulation cleavage accommodating large amounts
of strain, and a younger, spaced, planar cleavage accounting
for relatively less volume loss (e.g. M20). Occasionally,
a thin sandstone or dolostone bed within a silstone matrix will preserve a fold train that displays the geometric and chronologic relationship between the folds and the cleavages.

Cleavage formation in the Otavi rocks of the Domes
was accomplished, at least partially, by pressure dissolu-

tion. Even in silstones, insoluble clay and iron residues
remain on the cleavage surface and no mica growth is visi-
ble. The older cleavage tended to remove the soluble ma-
terial to accommodate shortening, making it difficult for
the younger cleavage to develop because of the dearth of
water and soluble minerals after the first deformation.
However, when one of the two cleavages is not present, the cleavages are not physically distinctive in hand sample. Quartz veins are often present, but it is difficult to tell whether they represent precipitation of material mobil-
ized by pressure dissolution, or whether they are related to
other processes.

If granite diapirism was the cause of doming, a single,
early horizontal cleavage inclined away from the core of
the dome would be expected. The common occurrence of
two sets of mutually perpendicular penetrative cleavage,
each set axial-planar to a group of physically distinct folds,
supports the multiple-event, superposed folding hypothesis. Outcrop-scale relationships between multiple cleavages shed light on the sequence of deformational events within the Vrede domes.

Outcrop-scale evidence of three discrete folding events

Silstones along the central north-south axes of the
Vrede domes tend to preserve two cleavages. The two
cleavages intersect at 80-90° and display obvious crosscut-
ting relations. The S1 cleavage is tightly spaced, penetra-
tive and generally strikes E-W. The S2 cleavage is widely
spaced (up to 1-2 cm), nearly planar, variably penetrative
strikes consistently ~160° and dips ~60°. At M20 (Fig. 2)
and in the region along the axis of the keel-shaped syn-
cline, S1 is sheared into S-shaped micro-folds between the
S2 cleavage (Fig. 4). Towards the east and west flanks of
the domes, S1 becomes discontinuous and eventually disap-
ppears, while S2 becomes progressively more closely spaced and continuous.

It is possible to assign both S1 and S2 to a distinct set
of folds (F1 and F2 respectively), based on axial-planar
cleavage relationships. Measurable F1 folds are rare, but
where preserved, appear as stranded dolostone hinge zones
~5-20 m across with attenuated limbs (Fig. 5). F1 folds
trend approximately E-W to ENE-WSW, verge to the
north, and are gently (0-20°) plunging (Fig. 7b). Without
exception, the F1 folds preserve relatively undisturbed
axial-planar S1 in their hinges, while S1 is either crenulated
by S2 or not present at all outside of the hinge zones (Fig.
5). S2 transects both limbs of F1 folds (Fig. 5, 7c). F1 folds
are best developed along the N-S central axis of the domes
(e.g. M20, M9).

F2 folds are variable in style but are always smaller in
scale, more regularly developed throughout the domes and
more steeply plunging than F1. For example, at M9, F2
Figure 4: Thick siltstone unit located at M20. Bedding (Sδ) is at an angle to both S1 and S2, though Sδ is difficult to see in this outcrop. S1 is a closely-spaced cleavage sheared into sigmoidal micro-folds between the younger, widely-spaced S2 cleavage planes. S1 was sheared by flexural slip between S2 layers when S2 was broadly folded by D2. The crosshair marks an F3 fold axis going into the page.

Figure 5: Isolated E-W trending F1 fold hinge zone (marked by crosshair) and attenuated limb in Ghaub Fm. dolostone layer at M20. Within the siltstone inside the hinge zone (upper left corner of figure), the S1 cleavage is undisturbed and axial-planar to the F1 fold. In the siltstone outside the hinge zone (right side of figure), S1 is crenulated by a more widely spaced S2 cleavage that intersects S1 at a 90° angle and transects both limbs of the F1 fold.

Figure 6: Quartz vein precipitated between tectonically developed S1 cleavage planes at M20. D3 deforms S1 into open F2 folds and treats the original compositional layering (Sδ) as a passive marker (see text for discussion). The S2 cleavage is axial-planar to F2, and is more widely-spaced than S1. S2 is not obvious in this photograph, but can be recognized as attenuations between fine sandstone and siltstone at an angle to both S1 and S2.

In rare locations, clustered in the region of the keel-shaped syncline (M20, M54), entire outcrops of F1 and F2 folded massive siltstone are refolded into broad, 5 m-wavelength, 1 m-amplitude, E-W trending N-verging folds (F3). During F3 folding, flexural slip along the S2 planes sheared S1 planes into sigmoidal microfolds (Fig. 4).

Summary of deformational history

Three discrete deformational events are visible at outcrop-scale (D1, D2, and D3; Table 1). The chronology of deformation recorded at outcrop-scale is assumed to be consistent with, and representative of, the map-scale structural history (Ramsay and Huber, 1987; Davis and Reynolds, 1996; Passchier and Trouw, 1996). This approach is supported by the similarities in the relative orientation of cleavages with respect to outcrop-scale and map-scale folds.

It is interesting to note that none of the outcrop-scale structures show the Type-1 basin and dome interference pattern so obvious in the map-scale expression of the thick competent units such as the Ombombo 2, 3 and 4 Fm. that control the geometry of the Vredefort structures (Fig 2). Younger deformations in less competent siltstones exploit a tectonically developed set of nearly uniformly dipping layers (S4) rather than the primary bedding (Sδ) (Figs. 4, 5 and 6). In Figure 7, a model is presented for the sequence.
Table 1: Summary of the chronology and style of deformation in the Vrede domes.

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<th>Deformational Phase</th>
<th>Outcrop-scale style of deformation</th>
<th>Map-scale expression (interpretation)</th>
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<td>D₁</td>
<td>Large (5-100 m wavelength) E-W trending, shallowly plunging open folds (F₁) and a well-developed, steeply dipping, closely spaced axial-planar cleavage — F₂ and S₁ are typically strongly refolded/rotated by D₂.</td>
<td>Folded Otavi Gp. sediments into a broad E-W trending anticline-syncline-anticline train with cuspato-lobate geometry.</td>
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<tr>
<td>D₂</td>
<td>Small (0.5-2 m wavelength), tight, steeply plunging, N-S trending folds (F₂) in S₁ layering. F₂ folds are associated with a variably spaced axial-planar cleavage that dips to the west—F₁ and S₁ are consistently oriented throughout the domes and are rarely deformed by D₂.</td>
<td>Folded E-W trending anticline-syncline-anticline into a Type-I NE-verging basin and dome structure (Fig. 2). The Type-I two-dimensional interference pattern is the result of discrete D₁ and D₂ buckling phases interfering at nearly 90° angles (Ramsay, 1967; Theissen, 1980; Theissen, 1986). Pushed southern dome further into northern dome asymmetrically, squeezing intervening syncline into the shape of a keel and developing northern sense of vergence in both domes.</td>
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<td>D₃</td>
<td>Rare E-W trending open F₂ folds—style and degree of development vary over short distances due to the complex structural grain that had developed after D₁ and D₂. Renewed slip along S₁ planes shears S₁ planes into sigmoidal micro-folds.</td>
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of contractional events that may be representative of outcrop-scale deformation in less competent units throughout the domes.

Implications for Regional Tectonics

Tectonic Setting

The Damara-Kaoko-Gariep Orogen is a late Pan-African (560-500 Ma) trench-trench-trench triple junction recording the simultaneous convergence of the Congo, Kalahari, and Rio de la Plata cratons (Fig. 8; Hoffman, 1996; Prave, 1996; Hoffman, 1997). Trench sedimentation (Kukla and Stanistreet, 1991), formation of the Khorixas accretionary prism (Fig. 1), and south-verging fold belts (Miller, 1983) suggest that the Congo craton was the upper plate with respect to the Kalahari craton. Andean-type arc magmatism in the Camaqua retro-arc basin of southern Brazil (Gresse et al., 1996) and East-verging fold belts (Gresse and Gresse, 1991; Coward, 1983) suggest that the Congo craton was the lower plate with respect to the Rio de la Plata craton (Fig 8; Hoffman, 1996). Kinematic indicators imply that collisions along each arm of the triple junction were left-lateral oblique (Fig. 8a, Coward, 1983). One of the major problems that remains unsolved involves the relative timing of ocean closures and continental collisions between the three plates.

The timing of continental collision between the Congo, Kalahari, and Rio de la Plata cratons

In the Vrede domes, outcrop-scale crosscutting relationships between genetically linked folds and cleavages have distinguished a chronological sequence of deformational events for the domes (Table 1). In order to attach these phases of deformation to specific tectonic events, structures in the Vrede domes are linked with previous fold cleavage style and relative timing observations from structures along the Outjo and Kaoko thrust belts (Table 2). Then, the change in strength of each chronologically distinct set of structures with their relative distance from the inland and coastal deformation fronts will suggest which tectonic event is responsible for each set of structures.

Along the coastal belt, Coward (1981) identified predominantly N-S trending structures in the north which experience an abrupt 90° change in trend near Khorixas (Fig. 1), about 120 km east of the Vrede domes. The NE-SW (D₁) and N-S (D₂) trending structures weaken systematically to the east (Coward, 1983) and D₁ structures appeared before collision along the inland branch began (Miller, 1983). Therefore, it is suggested that D₁ and D₂ structures are related to deformation focused along the coastal belt associated with closure of the Adamastr ocean.

In the Vrede domes and further east in the Outjo thrust belt (Miller, 1983; Coward, 1983), D₁ and D₂ were weakly deformed by an E-W trending D₃. Southeast of the domes

Figure 7: Model for the kinematics of outcrop-scale D₁ and D₂ folding in less competent units: (a) undeformed slab of massive siltstone (S₀ is bedding); (b) D₁ deforms S₀ into open, E-W trending, N-vergent, gently plunging F₁ folds with S₁ axial-planar cleavage; (c) During D₂, the tightly-spaced S₁ cleavage is deformed into steeply plunging F₂ layer-parallel slip folds while S₁ is deformed as a passive marker. The plunge of the F₂ folds is determined by the dips of the S₁ cleavage planes, which are in turn related to the vergence of the F₁ fold and to the degree of fanning in the S₁ cleavage. Note: map-scale folding was controlled by thick competent units such as the Onombo 3-4 Fm. carbonates. Therefore, during D₂ S₁ remained the fabric along which layer-parallel slip was accomplished. The resulting map pattern reflects the basin and dome architecture typical of orthogonally interfering deformation fronts (see Ramsay, 1958, 1962; Weiss, 1959; Tobisch, 1966; Ghosh, 1970; Watkinson, 1981; Theissen, 1986; Ramsay and Huber, 1987; Lisle et al., 1990).
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<td><strong>D₉</strong> Local. Open to tight, upright to northward verging, E-W trending folds pre-dating Mulden Gp. deposition. (D₉)</td>
<td>NE-SW trending, SE-verging asymmetric folds. Deformation strengthens significantly in the Kaoko thrust belt to the north along the coastal belt. (K₁)</td>
<td>NE-SW trending structures post-dating Mulden Gp. deposition. (D₉)</td>
<td>Bedding-parallel cleavage formation pre-dating Mulden Gp. deposition. (D₉)</td>
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<tr>
<td><strong>D₂</strong> N-S folds and vertical cleavage of the Zerrisse fan. Effects of this deformation die out rapidly to the East. (D₂)</td>
<td>N-S trending, W-verging chevron folds and vertical cleavage. Post-dates the Sleaf granites (495 Ma). (D₄)</td>
<td>N-S trending, E verging folds and almost ubiquitous axial-planar cleavage.</td>
<td>Weak ENE–WSW trending folds and no cleavage development.</td>
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Galiep belt (Frimmel and Frank, 1998). The tectono-thermal evolution of the Galiep belt suggests earliest closure of the northern Adamastor Ocean, followed by destruction of the Khomas Sea, and succeeded by closure of the southern Adamastor ocean (Galiep belt) (Frimmel and Frank, 1998). Tectonostratigraphic studies within the Damara Orogen, however, suggest that either the Khomas sea closed before the Adamastor ocean (Stanistreet et al., 1991) or vice versa (Prave, 1996).

The evolution of the stress field at the junction of the coastal and inland belts

Why did ocean closure along the nearly N-S oriented Kaoko margin give rise to an initial phase of locally E-W trending structures (e.g. in the Vrede domes) followed by the development of N-S trending structures? Coward (1981) did not recognize a consistent interference pattern between the two structural trends, and considered the entire coastal belt to represent a single phase of differential shortening related to the closure of the Adamastor ocean and continental collision along the coastal belt. Coward (1981) went on to suggest that the change in structural trend was related to the tectonic evolution of the inland branch. During ongoing collision along the coastal belt, northward subduction of the Khomas sea commenced (Fig. 9a). Subduction of the Khomas sea ocean floor beneath the Congo craton generated large volumes of granite in the Swakop zone and the Outjo thrust belt (Fig. 1). Heat from the intruded granite weakened the lithosphere along the inland branch. This weakening of the southern margin of the Congo craton led to differential E-W ductile shortening and a counterclockwise rotation of structural trends (Fig. 9b). This interpretation then predicts that D₁ was E-W directed and that ongoing deformation led to a 90° counterclockwise rotation of originally N-S trending D₁ structures (Fig. 9b). Thus, D₂ would represent a renewed pulse of E-W directed shortening, refolding the E-W trending D₁ structures across N-S trending axes. This model is consistent with the observation that D₁ structures are only locally E-W trending (Coward, 1983) and that D₁ structural orientations are frequently controlled by the shape and location of basement inliers which may have acted as individual rigid indentors.
Figure 9: Models for the 90° rotation of structural trends between D1 and D2 during Kaoko orogenesis. The legend in (a) is relevant throughout this figure. (a) Simplified map of the triple junction between the Rio de la Plata, Congo, and Kalahari cratons. Zipper indicates diachronous E-W directed closure of the Adamastor ocean (closing progressively from north to south) around an Euler pole to the north of the diagram; (b–h) represent variable magnifications of the dome region at the junction of the inland and coastal belts (dashed boxes in (a)); (b) Orocline hypothesis (after Coward, 1981): The first phase of deformation (D1) is E-W directed. During progressive E-W shortening, the F1 fold axes are rotated 90° counterclockwise into shortening-parallel orientation. A second pulse of E-W directed shortening (D2) will fold E-W trending F1 axes across N-S trending F2 axes; (c, d) Block rotation model (after Nur et al., 1986): The first pulse of shortening (D1) is E-W directed and forms N-S trending folds (F1). Progressive D1 shortening activates a set of parallel strike slip faults oriented approximately 60° from D1. (c) Slip on the faults causes counterclockwise block rotation. The original faults will accommodate rotation until relative normal stress gets too high and slip is no longer energetically favorable. At this point, a second set of faults will form to accommodate block rotation, while the original faults lock (d). New fault sets would continue to develop until the crustal sliver that contains the domes rotated 90°. Then a second pulse of E-W shortening (D2) would fold the E-W trending F1 axes around N-S trending F2 axes; (e) A representation of the triple junction in a relative plate velocity field shows that it is necessarily unstable. Assuming that the left-lateral component of slip along each plate boundary is contemporaneous, migration of the triple junction will induce clockwise rotation of the Congo craton. C: Congo craton, K: Kalahari craton, P: Plata craton, CP: Congo-Plata plate boundary, KP: Kalahari-Plata plate boundary, KC: Kalahari-Congo plate boundary; (f, g, h) The existence of a basement promontory (e.g. the southwestward extension of the Kamanjab inlier) causes a rotation of the stress field as the Adamastor ocean closes diachronously from north to south (a). The first pulse of deformation (D1) forms E-W trending structures parallel to the north-facing edge of the Kamanjab inlier (f). The next pulse of deformation (D2) builds N-S trending structures parallel to the west-facing edge of the Kamanjab inlier as the Adamastor ocean begins to close around the basement promontory (g). The final phase of contraction (D3) is associated with the closure of the Khomas sea and forms E-W trending structures parallel to the southern margin of the Kamanjab inlier (h).
during differential shortening. However, as Carey (1955) asserts, Coward’s orocline explanation for the changing trend of the coastal belt would imply greater degrees of shortening and the unroofing of deeper structural levels along the southern Congo craton. The orocline hypothesis also predicts an increasing degree of horizontal extension towards the outer arc of the bend in the mountain belt. Mapping in the Swakop zone (Miller, 1983; Stanistreet et al., 1991) and in the narrow NW-SE window of Otavi Group rocks just north of the Zerrisene fan (Hoffinan, pers. com., 1998) does not show evidence for widely varying degrees of crustal shortening or horizontal extension.

Alternatively, a similar 90° rotation of D1 structures in the Vrede domes region could be accomplished if the southwestern Congo craton acted as a discrete rotating crustal sliver (Freund, 1978; Nur et al., 1986; Nelson and Jones, 1987; Sylvester, 1988; Fig. 9e). According to Nur et al. (1986), the 90° rotation of a crustal block requires that the block is bound by multiple sets of parallel strike-slip faults, where each set of faults has rotated less than 45° (Fig. 9d). The block rotation hypothesis is attractive because, unlike the orocline model, it does not predict an increasing degree of horizontal extension towards the outer arc of the bend in the mountain belt (Hoffinan, pers. com., 1999). Unfortunately, such variably rotated strike-slip faults have not been observed to bound the southwestern Congo craton. A second test of the rotating crustal block or orocline hypotheses would be to document palaeomagnetic rotations by dating palaeomagnetic poles from nearby syn-tectonic granitic intrusions (e.g. Nelson and Jones, 1987; Beck, 1998).

Two other models may satisfactorily explain the rotation of structural trends between D1 and D2 without requiring the 90° rotation of crustal elements. Most simply, McKenzie and Morgan (1969) predict that trench-trench-trench triple junctions are inevitably unstable. Using relative plate velocity vectors with directions based on the sinistral-oblique nature of convergence along each trench and with arbitrary magnitudes, Figures 9a and c indicate that the Kaoko-Damara-Gariep triple junction is indeed unstable. As the plates move, the junction will be unable to retain its geometry and the relative motion of the plates will have to adjust accordingly (McKenzie and Morgan, 1969; Nitsuma, 1996). In order to reach a stable triple-point configuration, the Congo-Rio de la Plata plate boundary (CP) must rotate clockwise into parallelism with the Kalahari-Rio de la Plata plate boundary (KP), forming one long transform fault (Figs. 9a and c). Perhaps the differential clockwise rotation of the three plates varied with time as the plate boundaries adjusted towards this more stable orientation. The complex evolving stress-field associated with the variable clockwise rotations of three interacting plates may have led to the 90° change in structural trend between D1 and D2 in the Vrede area.

A fourth explanation for changes in orientation of D1 and D2 structures involves the existence of an irregular coastal margin. Thomas (1983, 1990) used the southern Appalachian example to describe how the trace of an orogenetic belt may be inherited from the geometry of the earlier rifted margin. Irregular margins composed of promontories and embayments result in diachronous collisions and rotations of the stress field (Figs. 9f, g and h). Indeed, Porada et al. (1983) suggested that the Kamanjab inlier extends to the coast (Figs. 1, 9a, f, g and h). The Kamanjab inlier (1.7-2.0 Ga basement) could act as such a cold, rigid lithospheric promontory around which the trace of Kaoko structures would rotate (Vaucluse et al., 1994). If the collision along the Kaoko margin closed the Adamastor Ocean like a zipper around an Euler pole located well north of the Kamanjab inlier (Fig. 9a), as stratigraphic (Porada, 1989; Stanistreet et al., 1991; Germs and Gresse, 1991) and 40Ar/39Ar thermochronology suggests (Frimmel and Frank, 1998), then the Kaoko collision would have begun in the north and propagated southwards (Figs. 9a and f). Although plate convergence is E-W, the initial collision in the north (D1) would have been strongly oblique, with a component of N-S directed stress across the WSW-ENE trending northwestern edge of the Kamanjab inlier (Fig. 9i). As the zipper closed southward beyond the NNW facing embayment and around the W-facing promontory, compressional structures (D2) would have assumed a N-S trend (Fig. 9g).

This hypothesis can be tested directly by checking to see if the Kamanjab inlier does extend to the coast with the proposed geometry. In fact, the Zerrisene fan (Fig. 1) deposits that currently cover the alleged extension of the Kaoko inlier are deep water turbidites that may have been deposited on oceanic crust (rather than debris flows deposited on a basement high) (Stanistreet et al., 1991). By examining Nd and Pb isotopes in syn/post-tectonic granites, we should be able to determine whether the granites traveled through 2.0-1.7 Ga granite basement or through an accreted sliver of oceanic crust (Hoffinan, pers. com., 1998).

Perhaps the boldest assumption made in this contribution and in previous work is that fold orientations and vergences describe relative plate motions and collision geometries directly. This is a risky assumption to make considering that pre-existing structure (Thomas, 1990; Vaucluse et al., 1994) and extensional collapse (Dewey, 1988) are only two of the many facors that may cause folds to mask true relative plate motions. Nevertheless, the striking correlation between outcrop-scale Vrede dome structures and the map-scale structures observed along the coastal and inland belts indicate that something in the tectonic evolution of NW Namibia is systematic and perhaps related directly to the relative movements of the Congo, Rio de la Plata and Kalahari plates.

**Conclusions**

The Vrede domes provide a rare opportunity to examine deformed Otavi Gp. sediments at the junction of the Kaoko and Damara belts. Unlike domal structures observed in the Outjo thrust belt and elsewhere in the Damara orogen, the Vrede domes are not (1) the result of post-tectonic granitic diapirism, nor are they (2) the expression of oblique convergence and lateral tectonic escape along a mid-crustal detachment. The Domes were formed by the interference of three discrete folding events, D1, D2 and D3. If one assumes that fold orientations and vergences are valid data for the prediction of relative plate motions, then the discrete deformational episodes in the Vrede domes suggest a chronology for the inland and coastal orogenies. Coupled with structural observations from the Kaoko (Guj, 1976; Porada et al., 1983, Porada, 1989; Coward, 1983; Dürr and Dingeldey, 1996) and Outjo (Frutos, 1969; Porada, 1979, 1989; Porada et al., 1983; Coward, 1981, 1983;
Miller, 1983; Weber et al., 1983) thrust belts, it is concluded that D₁ and D₂ reflect the closing of the Adamanstor Ocean and subsequent Kaoko orogeny. Closure of the Komas Sea and collision along the inland Damara branch (D₃) followed.

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