Journal of Structural Geology, Vol. 2, No. 1/2. pp. 19 to 27.
Pergamon Press Ltd. 1980. Printed in Great Britain.

0191-8141/80/0601-0019 $02.00/0

Shear zones in the Precambrian crust of Southern Africa

M. P. Coward

Department of Earth Sciences, The University, Leeds, England

(*Received* 25 *July* 1979; *accepted* 10 *October* 1979)

Abstract—This paper reviews the main features of the shear zones of southern Africa, their size, shape, orientation, state of strain and displacement.

There are two main types of shear zones, those that form the boundaries to orogenic belts or deformation domains within the belts and those which form discrete zones crossing earlier structures or otherwise undeformed rock. These two types of shear zones have different strain patterns.

The major shear zones may be steeply dipping or flat lying, with thrust sense or strike slip sense of displacement. Often the shear zones change their dip and strike along their length. The gently dipping shear zones are associated with crustal thickening, often with the uplift of lower crustal rock. None of the major shear zones in southern Africa forms a suture between two once separate plates but all must have propagated some distance from their source. It is suggested that the flat lying shears develop along low velocity zones within the crust.

The widespread Archaean shear zones suggest non rigid crust and as many of the intracratonic shears are flat lying there must have been considerable displacement and deformation within and just below the lower Archaean crust.

INTRODUCTION

The shear zones of southern Africa range widely in size and in age, from late Archaean zones in the Limpopo belt of Botswana, Rhodesia and the northern Transvaal to the late Proterozoic—early Phanerozoic shear zones in the Pan African mobile belts of Namibia and East Africa. The shear zones formed at different crustal levels, in rocks in granulite facies conditions in the Limpopo belt of southern Rhodesia, to rocks in low greenschist facies or below in the Pan-African belt of Namibia. The aim of this paper is to review some of the structural features of these major shear zones: their shape and orientation, amount of displacement and relation to Precambrian tectonics.

Holmes (1951) and later Kennedy (1964) and Clifford (1970) have divided the structure of Africa into cratons and mobile belts of different ages which cross each other to make a reticulate pattern. The belts are not necessarily Alpine type orogenic belts; Shackleton (1976) has shown that there is relatively little offset of structures across the mobile belts of Zambia and East Africa. Similarly, Briden (1976) and Piper (1976), using palaeomagnetic evidence, have detected no measurable displacement across the mobile belts of central and southern Africa. However, there is evidence that consumption of oceanic crust occurred in some of the more recent Precambrian mobile belts, as ophiolite complexes have been recorded from the Pan African (~ 500 Ma) belt of East Africa (Gass 1977) and possible ophiolites occur in Namibia (Kroner 1977, Shackleton 1977). Some of the tectonics in the Archaean mobile belts of Rhodesia and Botswana have been interpreted in terms of plate collision and ocean consumption (Coward *et al.* 1976a, Burke *et al.* 1976).

Shear zones have been recognised in most of these belts. Sutton *et al.* (1954) noted that parts of the (~ 2000 Ma) Ubendian mobile belt of East Africa could be likened to the shear belts of N.W. Scotland. Hepworth (1965, 1967) observed zones of more intense deformation in Uganda and Tanzania where the older structures were strained and reorientated. He introduced the term 'straightening zone' for these structures. Cox *et al.* (1965) and Mason (1969) observed similar straightening zones at the edge of the lower Proterozoic (~ 2600 Ma) Limpopo mobile belt (Fig. 1). Sanders (1965) and Shackleton (1951) have reported zones of intense deformation at the edges of the Pan African mobile belt of East Africa and similar shear zones and thrust zones have been reported from the edges of other Pan African and late Proterozoic mobile belts (Halbich 1977, Bickle & Coward 1976).

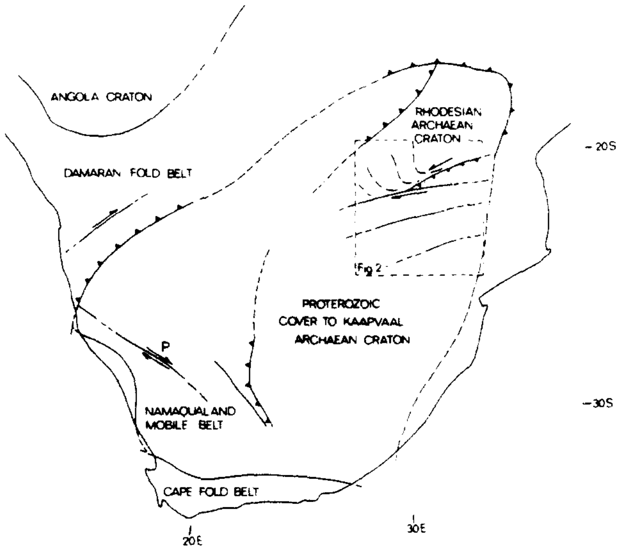

Fig. 1. Outline map of southern Africa showing the main shear zones (those with thrust sense shown by triangles), the main mobile belts and cratons and the areas of granulite facies rock (shown stippled). P = Pofadder shear zone.

Differential movement can be observed to have taken place across these straightening belts, the strain is more intense within the belts, earlier structures curve into the belts and there is no evidence of major volume change, therefore much of the strain must be that of simple shear. The belts are large scale examples of the s-type of Cobbold (1977). The transport direction of these shear zones may be determined from the trend of the linear structures, as with intense shear, early linear fabrics rotate towards, though never quite reach, parallelism with this transport direction (cf. Bridgewater *et al.* 1973).

The shear zones may be considered to be of two types:

(a) Shear zones which form at the margins of mobile belts or at the margins of sub-zones within the belts. These belts act as decoupling zones between rocks showing different deformation intensities.

(b) Discrete shear zones cutting rocks otherwise undeformed by this particular phase of deformation.

The distribution of some of the major steeply and gently dipping shear zones is shown in Fig. 1. The shear zones which are obvious on satellite and air photographs and on geological maps are those with steep to moderate dips. However, in the high grade mobile belts of southern Africa the shear zones are dominantly flat lying. The relationship between these flat lying shear zones and the structure of the mobile belts will be discussed separately.

DISCRETE SHEAR ZONES

Probably the largest discrete shear zone in southern Africa extends from north of Luderitz in Namibia to south of Pofadder in the North Cape, a distance of over 500 km (Fig. 1). This Pofadder shear zone is steeply dipping and has a minimum dextral displacement of 85 km in South Africa with a younger vertical displacement of unknown magnitude (Toogood 1976). Displacement varies along the shear belt; in Namibia it cannot be more than 15 km and the shear zone terminates near the Namibian coast (McDaid 1978). The shear zone crosses a recumbently folded charnockite–granulite gneiss province, which has been dated at between 2000 Ma and 1100 Ma (Reid 1977). In South Africa the shear zone is associated with a break in regional metamorphism, in that the rocks are higher grade on the NE side (Toogood 1976, Jackson 1979). The age of the shear zone is uncertain. It post dates almost all the regional deformation in this part of Namaqualand and Namibia. In Namaqualand pegmatites emplaced along the shear zone yield ages of 1000–900 Ma and give a minimum age of some of the shearing deformation (Joubert 1974).

Within the Pan African belts of Namibia, Zambia and East Africa there are steeply dipping discrete shear zones which formed late in the deformation history, after all the local deformation. Swardt *et al.* (1965) record several major sinistral shear zones in Zambia. In Namibia, Blaine (1978) records a dextral shear zone in the Okahandja area. In Kenya, there is a wide zone containing several discrete mylonite bearing shear zones (Bear 1952, Baker 1962) with gently plunging linear fabrics and hence mostly a strike slip sense of displacement. In Namibia, the shear zones post-date the main compression across the Damaran belt but the ages of the others in Zambia and Kenya are uncertain.

There are several discrete steeply dipping shear zones in the granites and greenstones of central Rhodesia (Coward 1976). In some zones, as in the Tati greenstone belt (Figs. 2 and 3B), the shearing appears to be synchronous with the regional deformation. Other zones

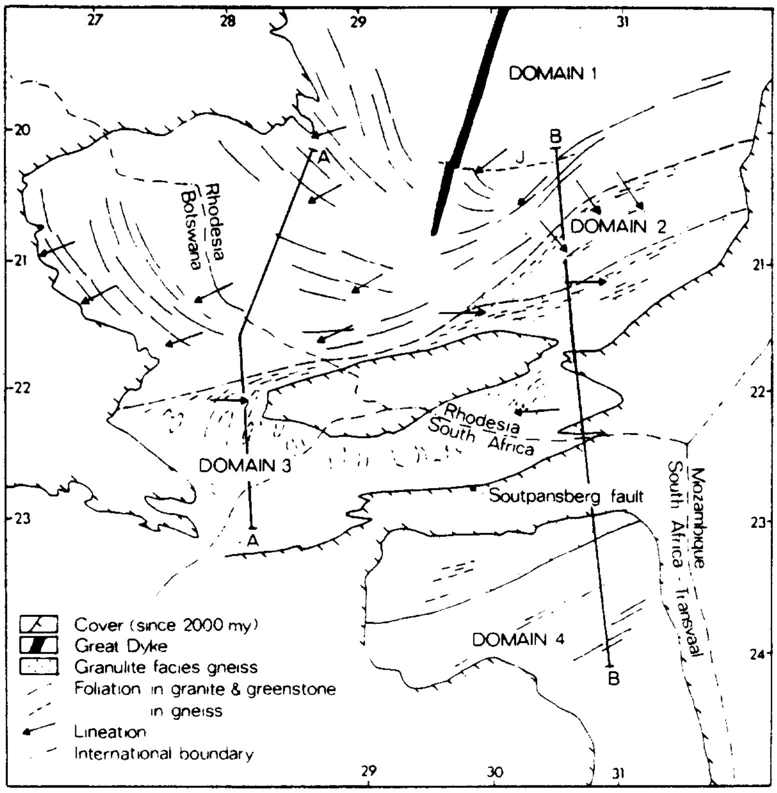

Fig. 2. Map of the Limpopo belt and adjacent Rhodesian and Kaapvaal cratons showing the different tectonic domains. Section line A refers to Fig. 7a, line B refers to Fig. 9. J = Jenya fault.

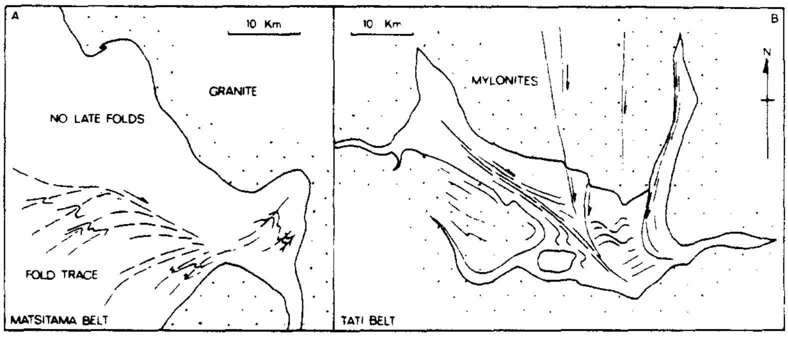

Fig. 3. Simplified maps of the Matsitama (3A) and Tati (3B) greenstone belts. Locations shown in Fig. 4. Fig. 3(B) after Coward & James (1974).

are later and have a complex displacement history. The Jenya–Mushandike shear-zone in Rhodesia (location in Fig. 2) can be traced for more than 100 km from Fort Victoria to south of Selukwe. It is a ductile shear zone with a core of mylonite, locally of quartz breccia. From the curvature of the foliation into the shear zone and the displacement of granites across it, the displacement at the western end of this zone appears to have been sinistral, of 8–10 km magnitude. At the eastern end of the shear zone, however, from the curvature of the fabric into the zone, the displacement appears to have been dextral. In both areas the displacement was dominantly strike slip but the relationship between the two displacements is unknown.

SHEAR ZONES AT THE BOUNDARIES OF MOBILE BELTS OR AT THE BOUNDARIES OF SUB-ZONES IN THE BELTS, EXAMPLES FROM THE LIMPOPO BELT

In the Limpopo belt, shear zones mark the boundaries of the different structural domains. These domains have been delineated from the changes in orientation of schistosity and linear structures (Fig. 2), the orientations of which reflect the different movement patterns for each domain (Coward 1976). Within Domain 1 the rocks consist of Archaean granites and greenstone belts and this domain is characterised by an arcuate pattern of schistosity but a constant trend to the linear structures. Coward *et al.* (1976a) considered this arcuate foliation pattern to be due to movement of the Rhodesian craton to the SW relative to the South African craton, producing shortening across the foliation in SW Rhodesia but simple shear with horizontal movement on a steep shear plane in the SE of this domain. This deformation must post-date the main phase of tonalite-granite intrusion dated at about 2600 Ma (Hawkesworth *et al.* 1975) but pre-date the intrusion of the Great Dyke (Fig. 2) dated at 2560 Ma (Allsopp 1965).

Domain 2 is largely confined to an area of locally sheared and mylonitised granulite facies rock. The northern margin of this domain in eastern Rhodesia is marked by a major shear zone, striking ENE, with a down dip lineation. Within this domain there are several shear zones but the lineations dominantly plunge to the SSE. James (1975) considered the granulites of this domain to have been thrust from the SSE along these gently dipping shear zones and used strain intensity to estimate a horizontal displacement of a 27 km and a vertical displacement of ~ 25 km across these zones. An interpretation of the gravity data across southern Rhodesia, supports this model of thrusting from the south-east (Coward & Fairhead in press). These shear zones post-date the deformation of Domain 1 but they and the granulites are cut by satellites of the Great Dyke. Domain 4 (Fig. 2) has a similar extension direction and also contains granulite facies rocks.

The shear zone which marks the northern boundary of Domain 3 crosses the boundary between Domains 1 and 2 (Fig. 2). In the east it is gently dipping but to the west, in Botswana, the shear zone is steeply dipping. The linear fabric in the shear zone is nearly horizontal and trends E–W. The zone has a dextral sense of displacement, as deduced from the curvature of structures into the zone (Coward *et al.* 1973). The mylonitic granulite-facies gneisses in Rhodesia grade upwards and southwards into flat lying amphibolite facies gneisses which have been folded into upright to westerly verging N–S trending folds (Fig. 2). These folds curve into and become sub-parallel to the shear zone at the northern edge of Domain 3, but to the north in Rhodesia, no folds of this age have been found. The marginal shear zone has acted as a decoupling zone for these folds.

A similar, though smaller scale, decoupling shear zone occurs in the Matsitama greenstone belt of northern Botswana (Figs. 2 and 3A). Here late phase folds, possibly the same age as those described from Domain 3, curve to become sub-parallel to a shear zone on its southern side, but do not occur on its northern side.

An important feature of these marginal shear zones is that they are asymmetric with different strains on either side of the zone. It is not always possible to decide whether these shear zones are structurally necessary zones to maintain compatibility between rocks which have suffered different deformation intensity, or whether the asymmetry is due to the shear zone dying out, producing differential displacement and hence longitudinal strain parallel to the shear zone. It is evident that the strains within these zones will contain components of longitudinal strain as well as shear strains and hence measurement of the strain intensity across the zone will not give an accurate measure of the amount of shear strain and hence the amount of displacement.

THE RELATIONSHIP BETWEEN SHEAR ZONES AND OVERALL STRUCTURE OF THE MOBILE BELTS

The cratons are characterised by upright foliations and shear zones, while the mobile belts, particularly the older belts, are characterised by gently dipping to flat lying foliations, folds and shear zones. Some of these gently dipping fabrics and shear zones were formed where the gneisses were thrust over the adjacent cratons. At the northern margin of Domain 2 in the Limpopo belt (Fig. 2) gently dipping shear zones formed where the gneisses of the mobile zone were thrust over granites and greenstones of the Rhodesian craton. However, elsewhere in the mobile belts, gently dipping shear zones do not uplift lower crustal rocks. An example will be described from the Limpopo belt and adjacent craton in Botswana and Rhodesia where there is evidence for a flat lying decoupling shear zone beneath Domains 1 and 3.

Strain measurements have been made across the gneisses, granites and greenstones of Domain 1 and these are summarised in Fig. 4. The strain increases from zero shortening ($\sqrt{\lambda_3} = 1$) east of Shabani to over 60% shortening ($\sqrt{\lambda_3} = 0.4$) across the Tati and Matsitama greenstone belts. There are zones of locally more intense deformation though the strain across a section between Shabani and Tati is thought to be dominantly that of pure shear (Coward 1976). If there was a major component of simple shear then there would have been considerable uplift of one end of this section; simple shear which produces a strain ellipsoid with a steeply plunging maximum extension direction must involve vertical displacement. However, from Shabani to Tati, a distance of 200 km, the rocks are granites and greenstones with no evidence of uplift of lower crustal rocks. The deformation is regional; the fabric crosses the granites and greenstones and the deformation is not due to localised diapiric intrusion of the granites.

Within the greenstone belts and granites the strain fields may be divided into areas of locally intense strain, generally closely associated with shear zones, and areas which show the background strain. Figure 5 shows the background strains $\sqrt{\lambda_3'}$ and $\sqrt{\lambda_1}$ plotted on a section line between Shashi and Shabani (Fig. 4), where $\sqrt{\lambda}$ = (final length/original length) and $\sqrt{\lambda_3'} = 1/\sqrt{\lambda_3}$. This figure assumes no volume change across this area so that $\sqrt{\lambda_1} \cdot \sqrt{\lambda_2} \cdot \sqrt{\lambda_3} = 1$ and hence the values of $\sqrt{\lambda_1}$ and $\sqrt{\lambda_3}$ may be obtained directly from the finite strain ratios. Strain values off this section line have been projected on to the line along the strike of the foliation. Assuming pure shear deformation, an estimate of the amount of shortening may be obtained by integrating the values of $\sqrt{\lambda_3'}$ (Hossack 1978) along this line. This gives a mean strain of $\sqrt{\lambda_3} = 0.55$, that is over 240 km displacement across this zone.

Figure 5(b) shows the values of $\sqrt{\lambda_1}$ for the background strains across this section. This maximum extension direction has a steep plunge throughout. The extension varies from zero ($\sqrt{\lambda_1} = 1$) east of Shabani to $\sqrt{\lambda_1} = 2.35$ in NE Botswana. It is unlikely that the whole crust would have been thickened by this amount unless there had previously been considerable crustal thining; a crust originally 40 km thick would have been thickened to over 100 km in NE Botswana. There is no geophysical evidence for thick crust in this area (Coward & Fairhead, in press) nor for any isostatic uplift, bringing lower crustal rocks to the surface. Alternatively the pure shear may decrease in intensity downwards in the crust and the transition from the zone of high intensity strain to the zone of low intensity strain may be some form of flat decoupling zone.

There is evidence for this decoupling zone in Botswana. To the south of the Tati belt, the shear zone at the northern margin of Domain 3 uplifts amphibolite facies gneisses. The gneisses show intense deformation and often carry a mylonitic fabric which is folded round the later upright to inclined folds of Domain 3 and so must originally have been flat lying (Coward *et al.* 1976b, Wright 1977). Lithological units and structures can be

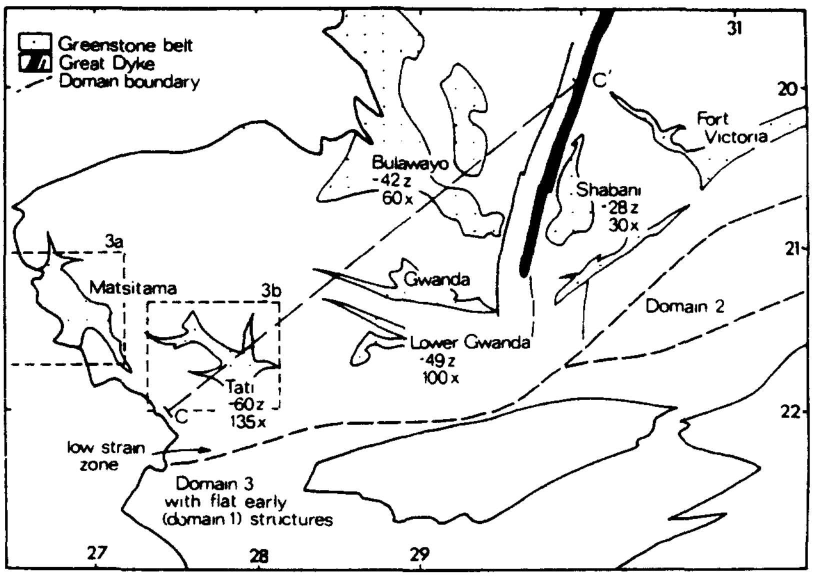

Fig. 4. Map showing variation in mean background strain in the greenstone belts of southern Rhodesia.

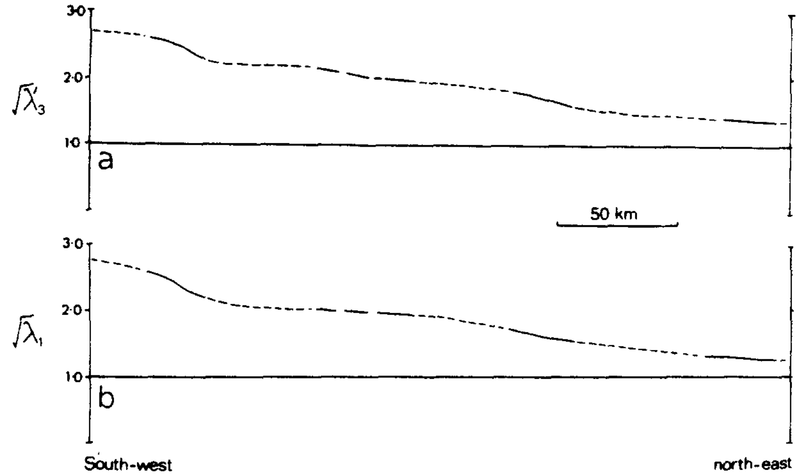

Fig. 5. Variation in $\sqrt{\lambda_3}$ and $\sqrt{\lambda_1}$ across the southern part of Rhodesia. Section line shown on Fig. 4. Dashed line denotes strain estimation rather than measurement.

traced across the shear zone at the northern margin of Domain 3 and the intense originally flat lying fabric in the south can be correlated with the more upright fabrics with down dip extension direction of Domain 1 in NE Botswana (Coward *et al.* 1976a). Thus at low structural levels, associated with amphibolite facies metamorphism, the fabric was flat but in the higher structural levels, associated with greenschist facies metamorphism, the fabric is steeply dipping. The model for this variation in structural orientation is shown in Figs. 6 and 7(a). This suggests that the upright fabric, with associated crustal thickening, changes downwards into a flat lying shear zone. The upright fabric in NE Botswana and SW Rhodesia may be due to northward decrease in shear along this zone, causing longitudinal strain and hence thickening of the upper crust. In Botswana the displacement across this zone must be over 200 km.

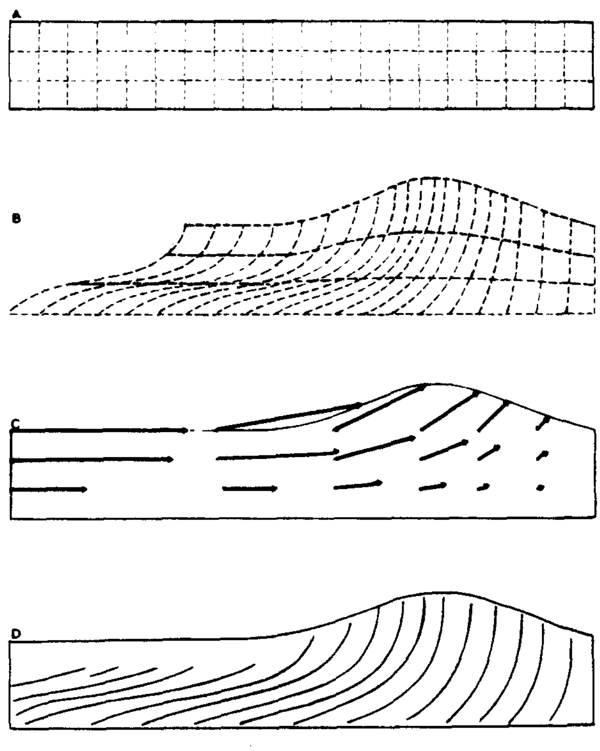

Fig. 6. Schematic model for the displacements across domain 1, Rhodesia and Botswana.
(A) Undeformed grid. (B) Deformed grid. (C) Displacement vectors. (D) Orientation of principal axes of the strain ellipsoid.

Similarly within Domain 3 in Rhodesia and South Africa the upright folds die out northwards into a gently dipping shear zone, suggesting shortening and thickening of the rocks above this shear zone (Fig. 7b). This major shear zone itself changes orientation. In Botswana it deforms amphibolite facies gneisses and is accompanied by amphibolite facies metamorphism (Hickman & Wakefield 1975) but to the east in Rhodesia it is gently dipping and is accompanied by granulite facies metamorphism (James 1975). Assuming that the metamorphic grade reflects the depth of the shear zone during deformation and that the change in metamorphic grade along the zone is due to later tilting then this major shear zone changes from a steeply dipping strike slip zone at moderate to high crustal levels to a gently inclined zone at greater depth.

Possible relationships between shear zones in the mobile belts and structures in the cratons are shown in

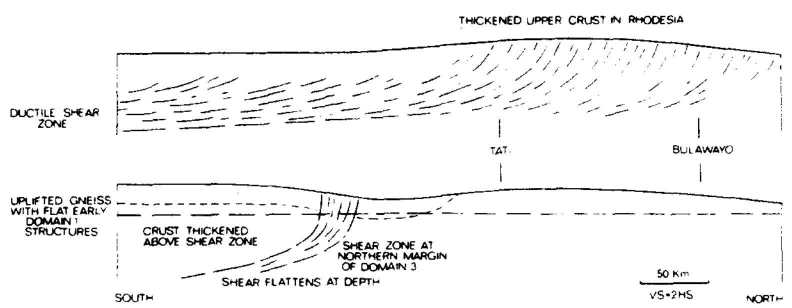

Fig. 7. (A) The preferred explanation for the change in fabric and strain across the section line A (Fig. 2) after the Domain 1 deformation. (B) as above, after Domain 3 deformation.

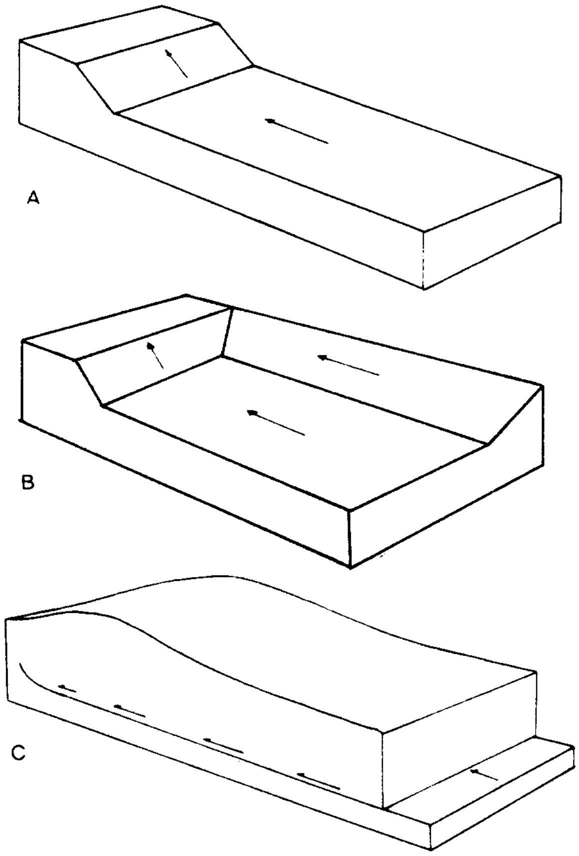

Fig. 8. Suggested shapes of the major shear zones, see text for discussion.

Fig. 8. This suggests that in the mobile belts the shear zones are flat lying to gently inclined and the junction between mobile belt and craton occurs where these zones:

(a) are gently inclined and overthrust high level rocks (Fig. 8A), so that the junction between mobile belt and craton is a thrust zone; or (b) become more steeply inclined but with strike slip sense (Fig. 8B), so that the junction is a transcurrent fault or shear zone; or (c) die out to produce a zone of shortening and thickening in the overlying rocks (Fig. 8C).

Note that this pattern of shear zones is different from that suggested by certain modern interpretations of studies on Himalayan tectonics where probably the whole lithosphere has been thickened (Dewey & Burke 1973, Molnar & Tapponier 1978). In such a model the thrust zones would be expected to steepen downwards, the fabric in the mobile belt should be steep associated with this thickening and there should be evidence of thickened lithosphere. This type of tectonics has not been recognised so far in southern Africa.

DISCUSSION

Fabric and folds in the shear zones

In the major shear zones of the Limpopo belt, the pre-existing fabrics, the cleavage and gneissic banding, curve and become further deformed. The fabric in the shear zones is usually a finite fabric, only rarely has a new fabric been produced, though there are often metamorphic changes associated with the intensification of the pre-existing fabric.

Within the steeply dipping shear zones the foliation is often folded and the fold hinges plunge nearly parallel to the mineral lineations in the zone. Also in the gently dipping to flat lying shear zones there are small to large scale recumbent folds and many of these have a characteristic sheath-fold shape (cf. Quinquis *et al.* 1978). These folds may be produced by a component of layer parallel shortening along the zones but also some may be produced by instabilities in the shear plane as described by Quinquis *et al.* (1978) from shear zones in the Ile de Groix and by Hudleston (1977) from shear zones in glacier ice. There need be no simple correlation between folds within a shear zone whether it be a small steeply dipping zone, or a major gently inclined zone, these folds could form at different times depending on the instability. Instabilities could cause the local development of several phases of folds and assuming that the nearly horizontal gneissic foliation in the mobile belts has been formed or modified by simple shear, it may be unwise to correlate structures across large areas of the mobile belts.

The shape of shear zones in three dimensions

On a small scale many shear zones are not straight but curved and wrap around lenses of less deformed rock and on a larger scale, in the mylonite zone at the northern margin of Domain 3 in Rhodesia, lozenge shaped blocks of undeformed material, up to 1 km wide, are separated by intensely sheared gneiss (Coward *et al.* 1976a). The foliation changes orientation around these blocks but the lineation maintains a constant trend. Thus the shear zones change from thrust sense to strike slip sense of displacement around these less deformed blocks.

The position and origin of the gently dipping shear zones

Evidence has been given for flat shear zones in the Limpopo belt. Similar flat lying shear zones have been described from other Precambrian areas. In the Namaqualand mobile belt in the NW Cape of South Africa (Fig. 1), the rocks are gengly dipping to flat lying intensely deformed amphibolite to granulite facies granites and gneisses, showing several local phases of recumbent folds. The shear zones are only steeply dipping where folded by later upright structures. Much of the intense Laxfordian deformation in the Lewisian complex of NW Scotland lies within originally flat-lying to gently dipping shear zones (Coward 1974) and Bridgewater *et al.* (1974) have described horizontal shears in the Precambrian rocks of West Greenland. Recent reflection seismic images of the lower crust in the U.S.A. (Phinney 1978, Smithson 1978) show nearly horizontal reflection surfaces which may be interpreted as nearly horizontal gneisses or shear zones.

Seismic images of the Wind River Thrust, Wyoming, obtained by the COCORP deep reflection programme (Smithson 1978) show that this thrust maintains a constant dip of nearly 30° to depths of over 24 km. The shear zone with thrust sense at the northern edge of Domain 2 in Rhodesia may be an ancient example of this type of thrust. Such a shear zone presumably must flatten in some low velocity zone in the lower crust or lower lithosphere. Low velocity layers within the crust have been suggested from seismic refraction and reflection studies in several areas of North America and Europe (Smith *et al.* 1975, Bâth 1978). These low velocity layers must represent weakness layers in the crust and could cause deflection of a shear zone. Such deflections may be similar to the flats and ramps in high level thrusts where the thrust passes through layered competent and incompetent rocks (Dahlstrom 1970).

Recent deep electrical resistivity measurements across southern Africa (van Zijl 1978) have suggested the position and possible origin of such low velocity zones. A moderately resistive zone in the central part of the Limpopo belt, attributed to the presence of intensely deformed amphibolites and granulites, can be traced beneath a more resistive zone attributed to the granites and greenstones of the Rhodesian craton. This suggests that similar deformed and metamorphosed rocks exist at a depth of about 10 km beneath the Rhodesian craton. This tendency of the amphibolite and granulite facies gneisses to show only moderate resistivity, compared to the granites and greenstones, must reflect an appreciable increase in porosity downwards in the crust. It suggests a free water content interconnected through cracks in these deformed rocks. van Zijl (1978) suggests that this water content may partly account for the low velocity layers found at depths beneath the present shields. Beneath this zone of moderate resistivity in the Limpopo belt, at a depth of between 30–40 km, there is a highly conductive zone. The preferred explanation for this zone (van Zijl 1978) is that it is formed by hydration and that serpentinite could account for its electrical properties. This zone, drawn nearly horizontal by van Zijl, could reflect the main slip plane for the major shear zones beneath the Limpopo belt. Beneath this zone the rocks are highly resistive, suggesting low water content and presumably therefore more rigid lithosphere.

Shear zones and crustal thickening and uplift

Bridgewater *et al.* (1974) have suggested that gently dipping Archaean shear zones with thrust sense caused localised crustal thickening and hence an increase in crustal rigidity. Variable displacement along a flat or gently dipping shear zone as shown in Figs. 6, 7(a) and 8(a) would also cause localised crustal thickening. Similar layer parallel shortening and hence thickening, occurs above high level gently inclined thrust planes (Coward & Kim in press).

This analogy to high level thrusts may be taken further to explain crustal thickening and uplift away from the present outcrop of the shear zone. Where a thrust plane changes dip and climbs a ramp, a structurally necessary fold must be produced in the rocks above the ramp (Dahlstrom 1970). If the gently inclined shear zones in the lower crust change dip then this change must be accompanied by uplift of the overlying crust. This model may account for the uplift of granulite facies rock in the southern part of the Limpopo belt (Figs. 2 and 9). An interpretation of the gravity data (Coward & Fairhead in press) suggests that the granulite–amphibolite boundary dips gently to the south beneath the Kaapvaal craton. These granulites may have been uplifted by a shear zone of thrust sense along their northern margin (Fig. 9a). However the shear zones at this northern margin have dominantly a strike slip sense of displacement and some may post-date the uplift of the granulites. The alternative explanation is that the granulites were uplifted by a change in dip of the underlying shear zone, as shown in Fig. 9(b). The depth of this underlying zone is unknown but immaterial. This second model which suggests that the granulite–amphibolite boundary dips beneath the rocks of Domain 3 gives the better fit to the gravity data (Coward & Fairhead in press).

To the south of the granulites lies the Witwatersrand basin, a zone of mid-plate sedimentation which formed at approximately the same time as the deformation in

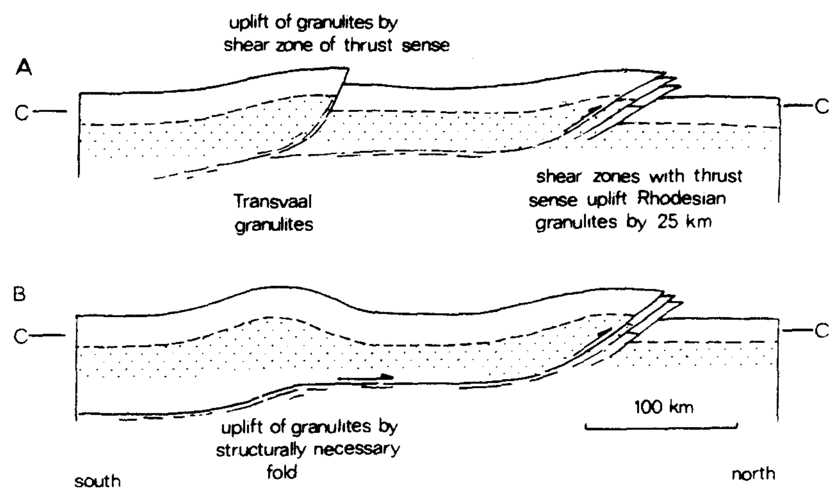

Fig. 9. Alternative sections to explain the uplift of granulite facies rocks in the Limpopo belts. Section line B on Fig. 2.

the Limpopo belt to the north. This basin need not have formed by only vertical crustal movement but possibly by horizontal movements on an underlying irregular shear zone.

The relationship of the major shear zones to plate collision

An analysis of present day plate motions by Minster & Jordan (1978) has suggested that there may be internal shortening within the Indian plate at a rate of about 1 cm a^{-1}. Thus the present day plates are not necessarily rigid. Such movements must give rise to anomalous stress conditions within the plates and from a study of the major earthquake zones within the Asian plate, north of Tibet, Molnar & Tapponier (1975, 1978) have suggested that deformation in Asia is analogous to that in a plain rigid/plastic medium (Asia) indented by a rigid indenter (India). The major strike slip faults in Asia correspond to slip lines in the plastic region. Watterson (1978) has applied this theory to explain the presence of steeply dipping shear zones of strike slip sense which formed in the early Proterozoic rocks of the North Atlantic craton. It is difficult to apply this theory to the steeply dipping shear zones of Africa. It is not yet known how the Pofadder shear zone relates to any plate collision and the later steep strike slip shear zones within and parallel to the strike of the Pan African belts, may reflect late strike-slip lateral movement of the cratons after any collision.

The gently dipping shear zones in southern Africa may have resulted from plate collision but no collision zone has been found which could account for the Limpopo belt and Namaqualand shear zones. The displacements associated with the Limpopo belt post-date the formation of the granites and greenstones of the Kaapvaal craton and must have been transmitted beneath the present craton. Whether these displacements were transmitted at the base of the lithosphere or within the Archaean crust is unknown. These shear zones represent decoupling of the upper from lower part of the plates.

Two possible models, relating the shear zones to plate collision, are shown in Fig. 10. Model 10A suggests decoupling of the plate from its downgoing slab; the shear zone would propagate in front of the over-ridden plate. Model 10B, based on the flake tectonic model of Oxburgh (1972) suggests that the shear zones propagate in front of the over-riding flake, above the subducting slab. Any calc-alkali magmatism associated with this subduction would be involved in the shear zone. This model could account for the large amount of syntectonic and post-tectonic granite within the flat lying gneisses of Namaqualand, and may also account for the granites of the Limpopo belt.

CONCLUSIONS

1. Some shear zones are structurally necessary structures formed to maintain compatibility between zones of different deformation intensity. Others are discrete shear zones unrelated to the local tectonics. These two types of shear zone have different strain patterns.
2. The shear zones may change strike and dip but maintain a constant movement direction.
3. Many of the structures associated with major ductile shear zones are analogous to structures in high-level thrust zones.
4. As many of the mobile belts contain gently dipping shear zones and much of the gently dipping fabric was caused by simple shear on a large scale, many of the folds may have been produced by instabilities in the shear planes and so there may be no correlation between fold phases in different parts of the mobile belts.
5. No evidence has been found for the major shear zones steepening at depth. Crustal thickening may take place above flat or gently dipping shear zones, or by lower crust being uplifted by gently dipping shear zones.
6. The flat shear zones may be caused by deflections along weak (low velocity) zones within or at the base of the crust.
7. Precambrian plates were not rigid. The flat or gently dipping zones represent decoupling of the crust from the lower lithosphere. These decoupling zones have formed at a considerable distance from an old plate margin.

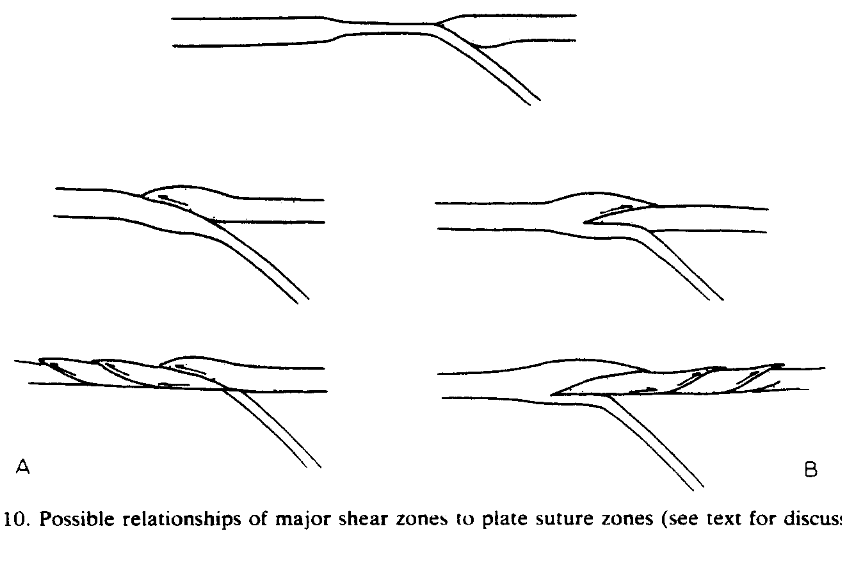

Fig. 10. Possible relationships of major shear zones to plate suture zones (see text for discussion).

REFERENCES

Allsopp, H. L. 1965. Rb–Sr and K–Ar age measurements on the Great Dyke of southern Rhodesia. *J. geophys. Res.* **70**, 977–984.

Baker, B. H. 1962. *Outline of the Geology of Kenya*. University of Nairobi publication.

Båth, M. 1979. Focal depth distribution of Swedish earthquakes. *Tectonophysics* **53**, T29–T31.

Bear, L. M. 1952. A geological reconnaissance of the area south-east of Embu. *Rep. Geol. Surv. Kenya.* **23**.

Bickle, M. J. & Coward, M. P. 1977. A major thrust in the southern Damaran belt, Namibia. *20th ann. Rep. res. Inst. afr. Geol., Univ. Leeds*, 8–13.

Blaine, J. L. 1978. Tectonic evolution of the Waldau ridge structure and the Okahandja lineament in part of the central Damara orogen, west of Okahandja, South West Africa. *Bull. Precambr. Res. Unit, Univ. Cape Town.* **21**.

Briden, J. C. 1976. Application of palaeomagnetism to Proterozoic tectonics. *Phil. Trans. R. Soc.* **A280**, 405–416.

Bridgewater, D., Escher, A & Watterson, J. 1973. Tectonic displacements and thermal activity in two contrasting Proterozoic mobile belts from Greenland. *Phil. Trans. R. Soc.* **A273**, 513–533.

Bridgewater, D., McGregor, V. & Myers, J. S. 1974. A horizontal tectonic regime in the Archaean of Greenland and its implications for early crustal thickening. *Precambrian Res.* **1**, 179–197.

Burke, K., Dewey, J. F. & Kidd, W. S. F. 1976. Dominance of horizontal movements, arc and microcontinental collision during the permobile regime. In: *Early History of the Earth* (edited by Windley, B. F.) Wiley, London, 113–130.

Cobbold, P. R. 1977. Description and origin of banded deformation structures. I. Regional strain, local perturbations, and deformation bands. *Can. J. Earth Sci.* **14**, 1721–1731.

Coward, M. P. 1974. Flat lying structures within the Lewisian Basement Gneiss Complex of N. W. Scotland. *Proc. Geol. Ass.* **85**, 459–472.

Coward, M. P. 1976. Archaean deformation patterns in southern Africa. *Phil. Trans. R. Soc.* **A280**, 313–331.

Coward, M. P. & Fairhead, J. D. in press. Gravity and structural evidence for the deep structure of the Limpopo belt, southern Africa. *Tectonophysics.*

Coward, M. P., Graham, R. H. James, P. R. & Wakefield, J. 1973. A structural interpretation of the northern margin of the Limpopo orogenic belt, southern Africa. *Phil. Trans. R. Soc.* **A273**, 487–492.

Coward, M. P. & James, P. R. 1974. The deformation patterns of two Archaean greenstone belts in Rhodesia and Botswana. *Precambrian Res.* **1**, 235–258.

Coward, M. P., James, P. R. & Wright, L. I. 1976a. The movement pattern across the northern margin of the Limpopo mobile belt, Southern Africa. *Bull. geol. Soc. Am.* **87**, 601–611.

Coward, M. P., & Kim, J. H. in press. Strain within thrust sheets. In: *Thrust and Nappe tectonics* (edited by McClay, K. & Price, N. J.) *Spec. Publ. geol. Soc. Lond.*

Coward, M. P., Lintern, B. C. & Wright, L. I. 1976b. The pre-cleavage deformation of the sediments and gneisses of the northern part of the Limpopo belt, southern Africa. In: *Early History of the Earth* (edited by Windley, B. F.) Wiley, London, 323–330.

Cox, K. G., Johnson, R. L., Monkman, L. J., Stillman, C. J., Vail, J. R. & Wood, D. N. 1965. The geology of the Nuanetsi igeneous province. *Phil. Trans. R. Soc.* **A257**, 71–218.

Clifford, T. N. 1970. The structural framework of Africa. In: *African magmatism and Tectonics* (edited by Clifford, T. N. & Gass, I. G.) Oliver and Boyd, Edinburgh, 1–26.

Dahlstrom, C. D. A. 1970. Structural evidence of the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petrol. Geol.*, **18**, 332–406.

Dewey, J. F. & Burke, K. C. A. 1973. Tibetan, Variscan and Precambrian basement reactivation: products of continental collision. *J. Geol.* **81**, 683–692.

Gass, I. G. 1977. The evolution of the Pan African crystalline basement in N.E. Africa and Arabia. *J. geol. Soc. Lond.* **134**, 129–138.

Halbich, I. W. 1977. Structure and tectonics along the southern margin of the Damara mobile belt, South West Africa. *Anal. Univ. Stell. Ser. Al(Geol)* **2**, 149–247.

Hawkesworth, C. J., Moorbath, S., O'Nions, R. K. & Wilson, J. F. 1975. Age relationships between greenstone belts and 'granites' in the Rhodesian Archaean Craton. *Earth Planet. Sci. Lett.* **25**, 251–262.

Hepworth, J. V. 1965. The charnockites of southern West Nile, Uganda and their paragenesis. *Int. geol. Congr.* **22**.

Hepworth, J. V. 1967. The photogeological recognition of ancient orogenic belts in Africa. *Q. Jl geol. Soc. Lond.* **123**, 253–292.

Hickman, M. H. & Wakefield, J. 1975. Tectonic implications of new geochronological data for the Limpopo belt at Pikwe, Botswana. *Bull. geol. Soc. Am* **86**, 1468–1472.

Holmes, A. 1951. The sequence of Precambrian orogenic belts in south and central Africa. *Int. geol. Congr.* **18**, 254–269.

Hossack, J. R. 1978. The correction of stratigraphic sections for tectonic strain in the Bygdin area, Norway. *J. geol. Soc. Lond.* **135**, 229–241.

Hudleston, P. J. 1977. Similar folds, recumbent folds and gravity tectonics in ice and rocks. *J. Geol.* **85**, 113–122.

Jackson, M. P. A. 1979. A major charnockitic-granolite province in southwestern Africa. *Geology* **7**, 22–26.

James, P. R. 1975. A deformation study across the northern margin of the Limpopo belt, Rhodesia. Unpublished Ph.D. thesis. University of Leeds, 369pp.

Joubert, P. 1974. Wrench-fault tectonics in the Namaqualand Metamorphic Complex. *Bull. Precamb. Res. Unit, Univ. Cape Town* **15**, 17–28.

Kennedy, W. Q. 1964. The structural differentiation of Africa in the Pan-African (± 500 m.y.) tectonic episode. *8th ann. Rep. res. Inst. afr. Geol., Univ. Leeds.*, p.48.

Kroner, A. 1977. The Precambrian geotectonic evolution of Africa: plate accretion versus plate deformation. *Precambrian Res.* **4**, 163–213.

McDaid, J. M. 1978. The geology of the northern part of Diamond Area No. 1, South West Africa. *Ann. Rep. Precambr. Res. Unit, Univ. Cape Town*, **14/15**, 124–140.

Mason, R. 1969. Transcurrent dislocation in the Limpopo orogenic belt. *Proc. geol. Soc. Lond.* **1655**, 93–96.

Minister, J. B. & Jordan, T. H. 1978. Present day plate motions. *J. geophys. Res.* **83**, 5331–5334.

Molnar, P. & Tapponier, P. 1975. Cenozoic tectonics of Asia: Effects of a continental collision. *Science* **189**, 419–426.

Molnar, P. & Tapponier, P. 1978. Active tectonics of Tibet. *J. geophys. Res.* **83**, 5361–5374.

Oxburgh, E. R. 1972. Flake tectonics and continental collision. *Nature, Lond.* **239**, 202–204.

Piper, J. D. A. 1976. Palaeomagnetic evidence for a Proterozoic super-continent. *Phil. Trans. R. Soc.* **A280**, 469–490.

Phinney, R. A. 1978. Interpretation of reflection seismic images of the lower continental crust. *EOS Trans. A.G.U.* **59**, 389.

Quinquis, H., Audren, Cl., Brun, J. P. & Cobbold, P. R. 1978. Intense progressive shear in Ile de Groix blueschists and compatibility with subduction of obduction. *Nature, Lond.* **273**, 43–44.

Reid, D. L. 1977. Geochemistry of Precambrian igneous rocks in the lower Orange River region. *Bull. Precamb. Res. Unit, Univ. Cape Town* **22**.

Sanders, L. D. 1965. Geology of the Broderick Falls area. *Bull. Geol. Surv. Kenya.* **7**.

Shackleton, R. M. 1951. A contribution to the geology of the Kavirondo Rift valley. *Q. Jl geol. Soc. Lond.* **106**, 342–392.

Shackleton, R. M. 1976. Pan African structures. *Phil. Trans. R. Soc.* **A280**, 491–497.

Shackleton, R. M. 1977. Possible late Precambrian ophiolites in Africa and Brazil. *20th ann. Rep. res. Inst. afr. Geol., Univ. Leeds*, 3–7.

Smith, R. B., Braile, L. W. & Keller, G. R. 1975. Upper crustal low-velocity layers: possible effect of high temperatures over a mantle upwarp at the basin range—Colorado plateau transition. *Earth Planet. Sci. Lett.* **28**, 197–204.

Smithson, S. B. 1978. Modelling continental crust: structural and chemical constraints. *Geophys. Res. Lett.* **5**, 749–752.

Swardt, A. M. J. de., Garrard, P. & Simpson, J. G. 1965. Major zones of transcurrent dislocation and superposition of orogenic belts in part of central Africa. *Bull. geol. Soc. Am.* **76**, 89–102.

Sutton, J., Watson, J. V. & James, T. C. 1954. A study of the metamorphic rocks of Karrama and Kungwe Bay, Western Tanganyika. *Bull. Geol. Surv. Tanganyika* **22**.

Toogood, D. J. 1976. The structural and metamorphic evolution of a gneiss terrain in the Namaqua belt near Onseepkans, South West Africa. *Bull. Precambr. Res. Unit, Univ. Cape Town*, **19**.

van Zijl, J. S. 1978. The relationship between the deep electrical resistivity, structure and tectonic provinces in southern Africa. *Trans. geol. Soc. S. Afr.* **81**, 129–142.

Watterson, J. 1978. Proterozoic intraplate deformation in the light of south-east Asian neotectonics. *Nature, Lond.* **273**, 636–640.

Wright, L. I. 1977. A structural section across the central and north marginal zones of the Limpopo belt, southern Africa. Unpublished Ph.D. thesis, University of Leeds.