PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Archaean Deformation Patterns in Southern Africa

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Archaean deformation patterns in southern Africa

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Strain measurements have been made to help quantify the intensity of deformation and amount of displacement across Archaean greenstone belts in Rhodesia and Botswana and across the gneisses of the Limpopo mobile belt. The area has been divided into three domains based on the orientation of the finite strain fabric and the orientation of the maximum extension direction in associated shear zones. The domains are considered to have different movement patterns and to be similar to small orogenic belts.

Early deformation within the greenstone belts accompanied the intrusion of the diaipric granites, but there was also bulk translation and rotation of greenstone belt and gneiss leading to imbrication of the stratigraphic pile and the formation of large nappes of overturned rock. This was followed by regional phases of deformation which affected all the greenstone belts and the gneisses of the Limpopo belt. Detailed strain measurements show a variation in amount of shortening during this phase, from under 30 % across the Shabani–Bellingwe belt in central Rhodesia, to over 60 % across the Tati and Matsitama belts in northern Botswana. Many local variations in intensity of deformation occur within large ductile shear zones and deviations from plane strain may be partly due to such rotational deformation. The regional deformation pattern suggests that there was movement of the Rhodesian craton approximately 200 km to the southwest relative to the gneisses of the Limpopo belt, producing a dominantly flattening deformation in the southwest of Rhodesia, but dominantly simple shear with a nearly horizontal sinistral movement, in the southeast.

Introduction

In the Archaean rocks of southern Africa, measurements of natural strain have previously been confined to small areas or to individual problems. Ramsay (1963) reported two strain measurements from the Fig Tree and Moodies formations of the Barberton greenstone belt in the eastern Transvaal and Gay (1969) and Anhaeusser (1966) give further strain measurements from the Fig Tree formation in this belt. Wood (1973) gives results of ten strain measurements from the conglomerates of the Umvuma greenstone belt in central Rhodesia.

In 1970 the Research Institute of African Geology at Leeds University began a project to examine the geological structure and quantify the strain across the Limpopo orogenic belt and the greenstone belts and granites in the adjacent Rhodesian and Kaapvaal cratons (figure 1). Some results have been given elsewhere, in Coward & James (1974), Coward, James & Wright (in the Press), James (1975), Graham (1974) and Wakefield (1974). The aim of this paper is to summarize these results and discuss the geological structure of the Archaean rocks of southern Rhodesia and northeast Botswana in the light of these strain measurements.

From geochemical evidence, Anhaeusser *et al.* (1969) consider that the greenstone belts of Rhodesia and northeastern Botswana (figure 1) were originally deposited on ocean floor without any basement granite crust. However, Oldham (1970) and Bickle *et al.* (1975) report a clear unconformity of greenstone belt sediments resting on a basement gneissic granite, and Coward, James & Wright (in the Press) outline several areas of this gneissic basement.

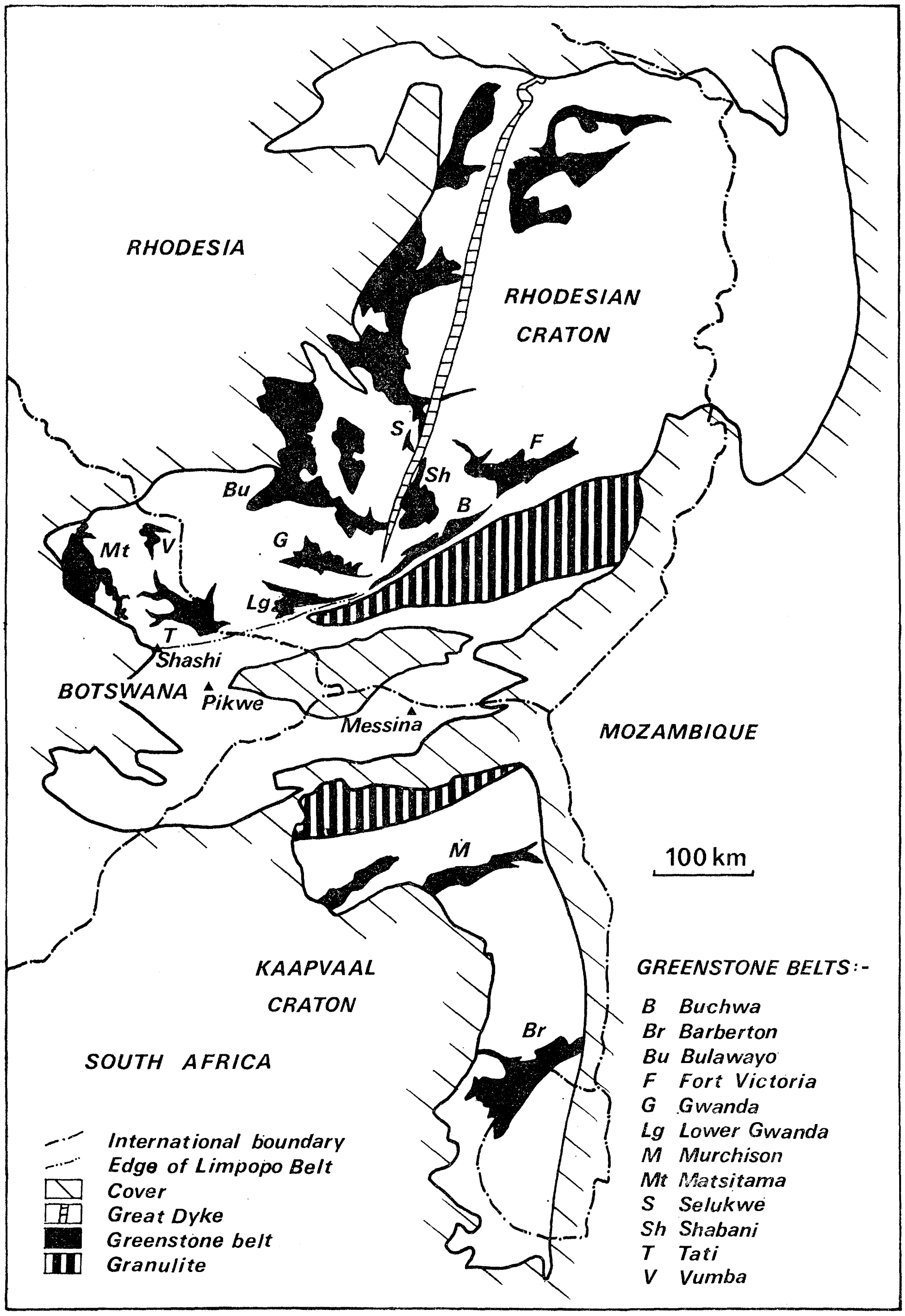

FIGURE 1. Simplified location map of part of southern Africa showing Rhodesian and Kaapvaal cratons, the greenstone belts and the position of the granulites at the northern edge of the Limpopo Belt. The position of the northern margin of the Limpopo belt is diffuse and depends upon the bias of the author. The line of this figure is after Mason (1973). The southern boundary is equally diffuse. Van Breeman & Dodson (1972) place it north of the granulites in the Transvaal; Mason (1973) places it south of the granulites. The 'Cover' is the Witwatersrand sediments and younger rocks (less than 2600 Ma).

Hickman (1974*a*) and Hawksworth *et al.* (1975) give ages of approximately 3600 Ma for these old gneisses.

The greenstone belts show a lithostratigraphic sequence from ultrabasic and basic volcanics to acid volcanics and sediments (Wilson 1973). Coward, Lintern & Wright (in the Press) consider that all the greenstone belts in southwest Rhodesia were once part of a large basin in which there were lateral facies changes. In the east, shelf sediments, intercalated with basic volcanics, rest on basement gneiss while in the central part there are dominantly ultrabasic and basic rocks. In the west there is a wide zone of thick shelf sediments which crop out in the Matsitama belt in northeast Botswana and consist of current-bedded sandstones, limestones and some basic volcanics. These are correlated with similar sediments in the Limpopo belt at Shashi and Madinare in eastern Botswana. Thus the metasediments of the Limpopo belt, which are normally termed the Messina Formation (Sohnge *et al.* 1949) from their type occurrence at Messina in the northern Transvaal, have been through the same deformation sequence as the main Rhodesian greenstone belts and appear to be of similar age but different sedimentary facies.

The greenstone belts and the sediments of the Messina formation have been intruded by several phases of diapiric granite, tonalite and diorite. The earliest granites are deformed together with the greenstone belts but the later granites form large irregular-shaped batholiths. Hawksworth *et al.* (1975) obtained ages from four of the greenstone belts and three of the granites from central Rhodesia of between 2800 and 2500 Ma, and similarly Hickman (1974*b*) has dated three of the large batholiths at 2600 Ma.

In the northeastern part of the Limpopo belt there is an area of granulite facies rocks (figure 1) (Cox *et al.* 1965; Mason 1973). Satellites of the Great Dyke, which intrude the granulites, have been dated at 2580 Ma, giving a minimum age for this granulite facies metamorphism (Robertson & van Breeman 1970). Deformation in the central part of the Limpopo belt may post-date the Great Dyke; gneisses in eastern Botswana, give a thin slice, whole rock isochron of 2100 Ma and Hickman & Wakefield (1975) consider that this indicates the age of the latest phase of ductile deformation in this zone.

The finite strain fabric and strain measurement

The granites and greenstones of Rhodesia and the gneisses and metasediments of the Limpopo belt show a penetrative finite strain fabric which can be traced from Fort Victoria south to Messina and from Shabani west to Matsitama. In the greenstone belts this fabric is shown by the preferred orientation of hornblendes, chlorites and micas, by pressure solution stripes and by the shapes of deformed objects (Coward & James 1974). In the granites and gneisses it is shown by the preferred orientation of minerals and the shape and orientation of deformed grains and inclusions. Measurements of the shape of the deformation ellipsoid were made from ellipsoidal or sub-ellipsoidal objects using methods described and Ramsay (1967), Dunnet (1969), Dunnet & Siddans (1971) and Roberts & Siddans (1971). Some quartz-rich grits and quartz grains within deformed granites were measured by the centre-to-centre method described by Ramsay (1967, pp. 195–197), that is by measuring the distance between centres of adjacent grains, this distance being proportional to the longitudinal strain in that direction.

The shape of the strain ellipsoid was also measured using the method described by Talbot (1970) which involves the determination of the shape of the surface of no finite elongation from a stereographic plot of poles to folded and non-folded or boudinaged veins, the boundary between the two fields being the trace of the normal to the surface of no finite elongation. A simple modification of the Talbot method was used in which the orientation of the surface of no finite elongation was measured directly on a surface, preferably parallel to a principal plane, by measuring the pitch on this surface of the line which separates those parts of the veins which are folded from parts which are boudinaged. Where there were insufficient veins for such determinations, measurements were made of the shortening along folded veins by using the method described by Hudleston (1973). This involves measuring the amount of post-buckle flattening and the shortening due to fold amplification and estimating the amount of shortening prior to fold amplification from the characteristic fold wavelength/thickness ratio. There may be considerable errors in such estimates; Cobbold (this volume) has shown that natural folds need not develop with a characteristic wavelength and hence estimates of the pre-buckle shortening may be little short of guess-work.

The strain parameters used in this paper are the ratios of the three principal axes, $X > Y > Z$, the percentage shortening or elongation in these directions, assuming constant volume, the natural logarithmic octahedral strain, E_s (a function of the octahedral unit shear γ_0) and the Lodes unit V where

$$\gamma_0 = \tfrac{4}{3}\,(\ln x/y + \ln y/z + \ln x/z)^{\frac{1}{2}},$$

$$E_s = (\tfrac{1}{2}\sqrt{3})\,\gamma_0$$

and

$$V = (\ln y/z - \ln x/y)/\ln x/z$$

(Hossack 1968; Gay 1969). Values of V range from $+1$ for uniaxial oblate ellipsoids to -1 for uniaxial prolate ellipsoids.

Subdivision of the region into structural domains

The fabric in the granites, greenstones and gneisses is a finite fabric, a result of several deformation stages, not necessarily synchronous throughout the whole area. The orientation of the finite fabric varies in orientation as does the mineral lineation which is normally parallel to the maximum extension direction of the deformation ellipsoid. The area has been divided into different domains based on the orientation of this fabric and on the orientation of the maximum extension direction in associated shear zones. The domains are considered to have different movement patterns similar to small orogenic belts, and are numbered 1 to 3, domain 3 being the youngest.

Domain 1 is characterized by a steep foliation and a down-dip lineation in the west, but in the south and east, the foliation curves form a northwest to a northeast trend, while the mineral lineation and maximum extension direction plunges northeast or southwest throughout (figure 2). There are localized zones of more intense deformation, major shear zones, which cut across the regional foliation. In the southern part of the domain, the shear zones have a northeast trend, a sinistral sense of movement and a nearly horizontal maximum extension direction. In the north and west of the domain, the shear zones strike north–south and have a dextral sense of movement but the lineation plunges in the same northeast–southwest plane (figure 2). Coward, James & Wright (in the Press) consider this arcuation of foliation and change in sense of shear zones to indicate movement of the Rhodesian craton to the southwest relative

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to South Africa, producing shortening across the foliation in the southwest part of the domain and some shortening but also considerable simple shear with near-horizontal movement in the southeast of the domain (figure 3). Much of this paper will be concerned with the state of strain within this domain.

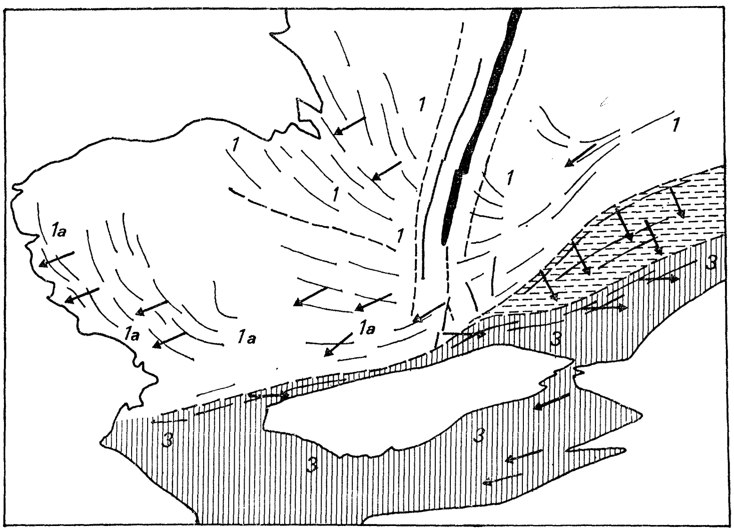

FIGURE 2. Map showing the trend of the foliation and lineation within the three structural domains. Domain 1*a* is subdivided from domain 1 on the presence of later structures which fold the cleavage. Cover is shown clear.

The northern margin of domain 2 is marked by a major shear zone with a down-dip lineation. Within this domain, the foliation has a similar trend to that in the adjacent parts of domain 1, however the fold hinges, the lineation and the maximum extension direction are near-normal to those in domain 1 and plunge to the south-southeast (figure 2). The rocks are granulite-facies gneisses and metasediments which are deformed by several major shear zones, each with the same strike of foliation, down-dip lineation and sense of movement. James (1975) has described the deformation of this domain in terms of a regional but heterogeneous simple shear deformation, the granulites being uplifted and thrust from the south-southeast.

The northern margin of domain 3 crosses the boundary of domains 1 and 2 (figure 2). In the east, the junction between domains 3 and 2 is a zone of gently dipping mylonites, while in the west the junction between domains 3 and 1 is a steep shear zone (Coward *et al.* 1973). In the east where this shear zone cuts relatively undeformed granites and granulites, it forms a new, near-horizontal fabric, but elsewhere, the earlier fabrics are rotated to produce an intense, generally mylonitic, finite strain fabric. On the mylonite foliation surfaces there is a well developed lineation shown by elongate quartz leaves and feldspar augen which presumably marks the movement direction in the shear zone. From this lineation and the curvature of the foliation into shear zones, the sense of movement is inferred to be dextral, southern and upper gneisses moving west, parallel to the strike of the shear zone (figure 3). In the east, the mylonitic gneisses grade upwards and southwards into less intensely sheared but still dominantly flat-lying gneisses which are interbanded with sediments of the Messina formation. These flat-lying rocks are folded and locally intensely flattened into a series of upright periclinal folds with axes trending north-northeast. These folds cannot be recognized in domain 2 or in the southeastern part of domain 1 and hence the marginal shear zone has acted as a plane of décollement between folded rocks to the south and the granulites, granites and greenstones to the north. Strain measurements made in the mylonites with the centre-to-centre method on the shape of deformed quartz grains gave an estimate of up to 50 km displacement across this zone in southern Rhodesia (Coward, James & Wright, in the Press; James 1975). As much of the deformation in the more cataclastically deformed gneisses presumably took place by grain boundary sliding or shear along discrete planes, strain measurements made by any conventional method are likely to be much too low, and as much of the flattening in the gneisses with domain 3 to the south may also have been accompanied by simple shear, this figure of 50 km displacement is probably a considerable underestimate.

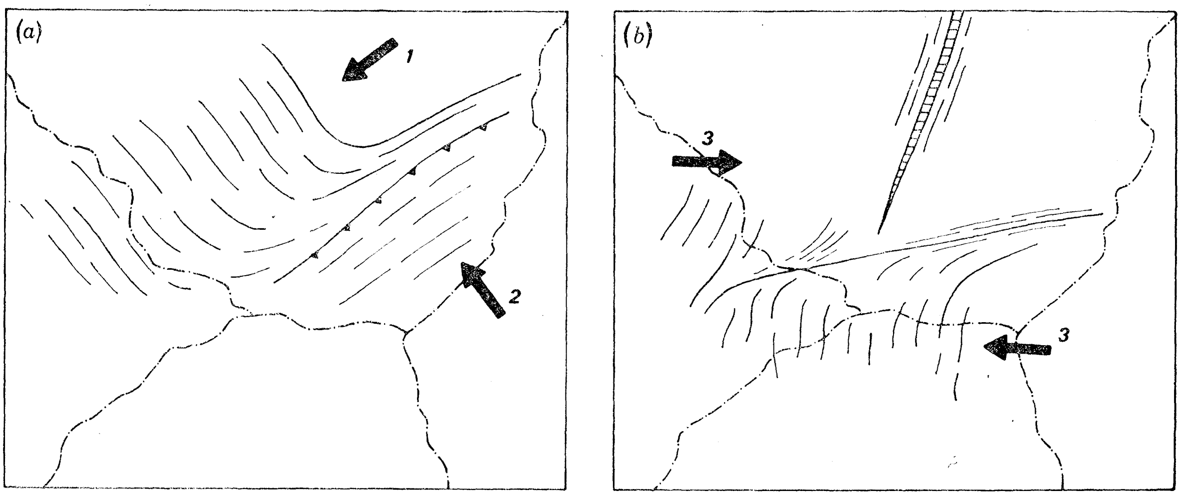

FIGURE 3. Diagrammatic models to show the directions of relative movement envisaged for the deformation of the three domains.

DEFORMATION HISTORY OF DOMAIN 1

The deformation state of the rocks in domain 1 is a result of several deformation events. Though the main fabric in this domain is a finite fabric and not necessarily synchronous throughout the area, it can be used as a form of time marker for the deformation sequence. The deformation history can be separated into:

1. A pre-cleavage regional deformation, prior to the intrusion of the diapiric granites.
2. Deformation associated with granite intrusion.
3. Regional deformation producing the main fabric.
4. Late phases of deformation which fold and whose fabrics cross-cut the regional fabric.

1. *The pre-cleavage regional deformation*

The rocks of the Tati greenstone belt in northeastern Botswana dip to the southwest but are overturned and young to the northeast (Mason 1968; Coward & James 1974). Litherland & Key (1974) have shown that the Tati greenstone belt can be joined to smaller fragments of greenstone belt material to the north (figure 4), to make an extensive sheet of overturned northeastward facing rocks. This overturning pre-dated the intrusion of the diapiric granites and the development of the main cleavage (figure 5). Similar overturned rocks occur in the core of the Matsitama belt, west of Tati (figure 4).

In southern Rhodesia, the Antelope and Lower Gwanda greenstone belts young upwards to a layer of gneiss which shows several phases of complex early deformation and which may well be allochthonous basement gneiss (Coward, James & Wright, in the Press). In southern Rhodesia, the Selukwe greenstone belt was overturned and thrust into its present position

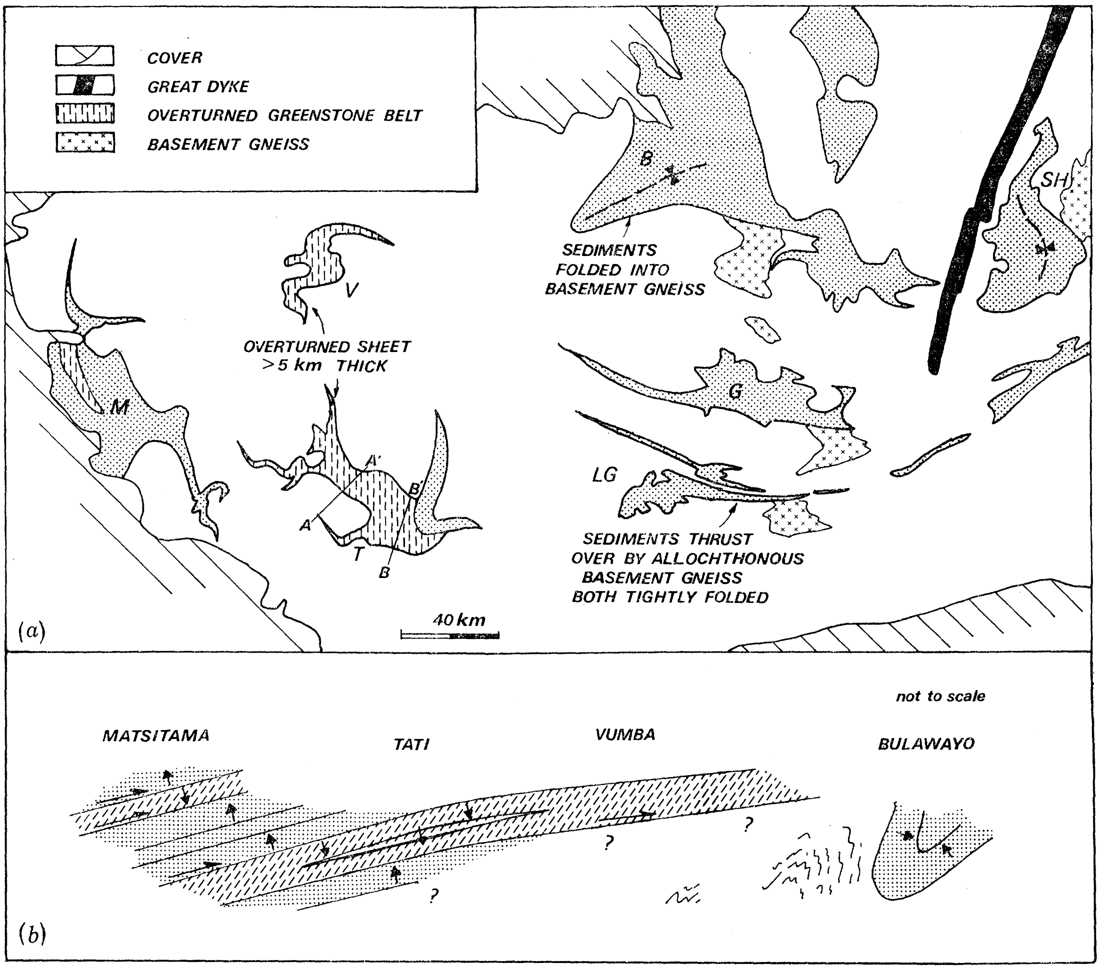

FIGURE 4. (a) Map of the southwest part of domain 1 showing the autochthonous and allochthonous greenstone belts and gneisses Greenstone belts: M, Matsitama; V, Vuma; T, Tati; B, Bulawayo; G, Gwanda; LG, Lower Gwanda; SH, Shabani.

(b) Schematic section through the greenstone belts between Matsitama and Bulawayo, before the intrusion of the diapiric granites. Arrows indicate younging direction.

before the intrusion of the diapiric granites (Stowe 1974). Thus much of the greenstone belt material and the gneiss in the south-west part of this domain is allochthonous and much of this early deformation took place by large scale bulk translation and rotation. The Selukwe greenstone belt may have been transported some considerable distance, the nearest comparable autochthonous ultrabasic and basic rocks are more than 50 km to the south or southwest, south of Belingwe and at Filabusi.

The greenstone belts at Bulawayo, Shabani and Fort Victoria appear autochthonous and that at Shabani rests directly on older basement, but these autochthonous greenstones also suffered folding before being intruded by granites and were then deformed by the cleavage-producing episode (figure 4). Although this folding was locally tight, no penetrative fabric was produced.

No thrust planes are seen except in the Matsitama belt where the junction between uninverted and inverted beds is marked by a zone of intense stylolitization.

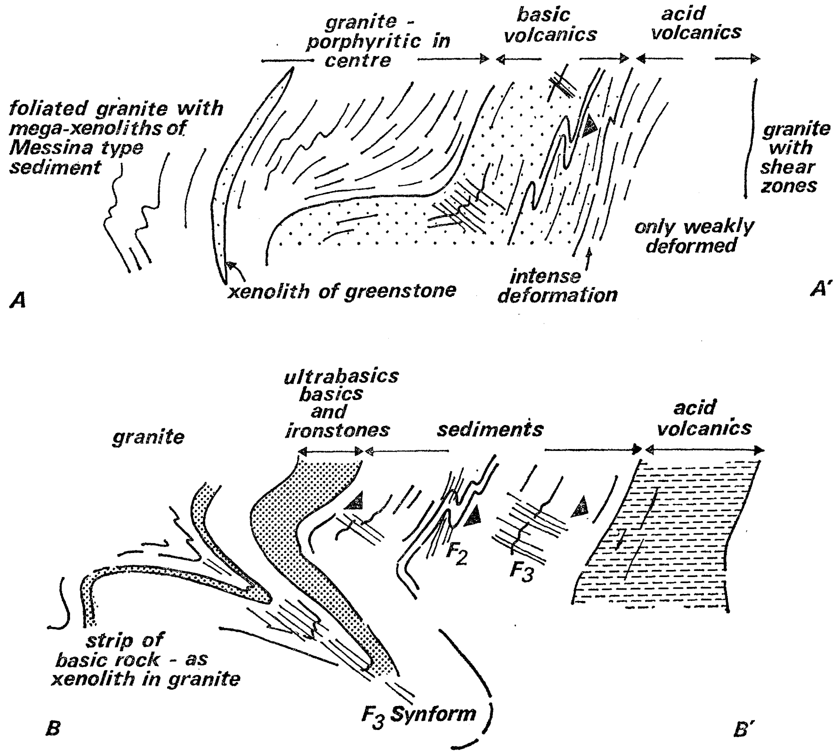

Figure 5. Two cross sections through the Tati greenstone belt. Section lines are shown in figure 4.

2. *Deformation associated with the intrusion of granite*

Anhaeusser *et al.* (1969), Anhaeusser (1973) and Mason (1973) considered that much of the deformation in Archaean greenstone belts was caused by the forceful intrusion of diapiric granites. However the granites in Rhodesia and northeast Botswana clearly post-date the over-turning of the Tati and Vumba greenstone belts (Litherland & Key 1974; Coward & James 1974) and at Bulawayo and Fort Victoria the granites post-date the folds in authochthonous greenstones.

Macgregor (1951) introduced the term 'gregarious batholith' for the large masses of 'granite' which surround the greenstones, and Fyfe (1973) suggested that the size of these batholiths reflected the size of the convection cells in the Archaean upper mantle. He suggested that the batholiths represent granite domes which grew by addition of new material at the centre, while the edge expanded outwards as a balloon.

However, most batholiths cannot be considered as simple 'balloons'. They consist of several ages of granite and the latest granites intrude not only the earlier granites at the centres of the batholiths, but also greenstones at the edges. Much of the 'granite' within the batholiths is basement gneiss, and this maintains its early tectonic trend throughout large areas and even has the same trend in adjacent batholiths (Coward, James & Wright, in the Press). There seems to have been very little reorientation of this gneissic material during the formation of the batholith, and most 'gregarious batholiths' should not be considered as batholiths at all.

The intrusive granites occur in two main forms:

(i) As distinct bodies, circular or elliptical in plan. These bodies vary from 1 to over 30 km in diameter. They are probably sections through diapiric 'bubbles' as described by Fyfe (1973).

(ii) As large, often irregular bodies, whose edge may be partly fault-controlled. The Chibi batholith south of Shabani forms one such body, over 30 km wide and over 140 km long.

The granites are of various ages. Many carry a foliation throughout, not just at the margin, and this foliation appears continuous with the tectonic foliation in the adjacent greenstones (Coward & James 1974). Many of the diapirs are elliptical in plan, their long axes being parallel to the foliation trend. Thus many of the granites have suffered a regional deformation along with adjacent greenstones. Other granites clearly post-date all the deformation; they cross-cut the structures and have irregular unfoliated margins.

Faulting and stoping were important intrusion mechanisms. Many of the early granites, such as those north of the Tati greenstone belt (Litherland 1973) contain large masses, 'mega-xenoliths', of greenstone-belt material. Some of the early gneissic material was reactivated and uplifted by faulting. West of the Fort Victoria greenstone belt, the Chibi 'batholith' which consists of late post-tectonic granite and also early, possibly pre-greenstone gneiss, cuts across all the structures in the Fort Victoria belt, but produces no new visible fabric in granite, gneiss or greenstone. The contact must be an arcuate fault or a series of faults.

Some tonalites and diorites carry deformed xenoliths which show a high strain ratio, much higher than that in the adjacent greenstone belts. The deformation shown by these xenoliths consists of two components, the regional cleavage producing deformation and an earlier deformation due to the intrusion of the igneous body. This is shown by the skewed distribution of points on the plot of strain ration (R_f) against ϕ, the pitch of the long axis of the strain ellipse. By using the programme 'STRANE' (Dunnet & Siddans 1971) and assuming that the cleavage represents the principal plane of the later deformation, values have been assigned to the two components of the finite strain in a tonalite southwest of Gwanda (figure 9). Though these tonalites show internal deformation, it is difficult to know how much of this deformation was transmitted to the adjacent country rock.

Some diapirs may have been formed by forceful intrusion; a rim syncline surrounds a granite which intrudes the Mont d'Or area of Selukwe, though this deformation may have been accentuated by later deformation. Similarly the greenstones in the southern part of the Tati belt were deformed to a distance of 1 km away from the contact of a large tonalite body before the main cleavage-producing deformation (Coward & James 1974). At distances greater than 1 or 2 km from the contacts, there seems to have been very little deformation due to granite intrusion.

Within the Limpopo belt, the cleavage-producing deformation was preceded by the segregation of quartzo-feldspathic material, granite and pegmatite and the intrusion of large sheets of granite. Coward, James & Wright (in the Press) suggest that many of these migmatites may be the deeper level equivalents of the diapiric granites seen in central Rhodesia. However, recent isotope work on the Rhodesian granites (Hawksworth *et al.* 1975, Moorbath, in the Press) shows that much of the Rhodesian granite is probably mantle-derived and cannot have been formed by partial melting of much older basement gneiss.

In the area between Bulawayo, Fort Victoria and the northern edge of the granulites, greenstone belts make up 28 % of the rock, older basement gneiss 18 %, early intrusive granite 25 % and younger unfoliated intrusive granite 29 %. Thus if the implications of Hawksworth *et al.* (1975) are correct, over 50 % of the rock now exposed at the surface in southern Rhodesia was new intrusive material added to the crust and not derived from partial melting of basement gneiss. This raises a severe space problem which has not been solved. Stoping and fault-controlled intrusion must have been the most important mechanisms for granite emplacement, but a considerable amount of material must have either sunk within the magma or must have been uplifted by the incoming granite. A large part of Rhodesia must have been underplated and possibly uplifted during this period and the tectonics may have been similar to those at the western margin of South America at the present day.

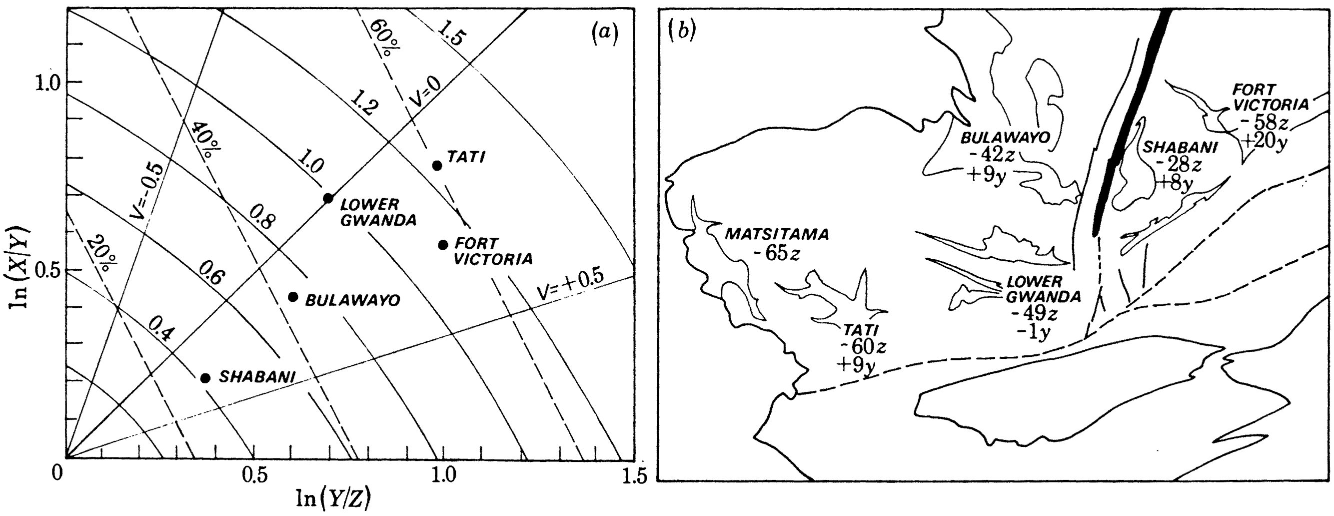

FIGURE 6. (*a*) Mean strains from five greenstone belts in domain 1, on logarithmic deformation plot (data from Coward, James & Wright, in the Press). Lines of equal value of V (Lode's unit), E_s (solid lines) and percentage shortening in the z direction (dashed lines) are shown.
(*b*) Map showing the mean strain expressed in terms of percentage shortening in the z direction and percentage shortening or elongation in the y direction for the greenstone belts. Figures from the Matsitama belt were taken from a smaller amount of data, and are not represented in figure 6*a*.

3. *Regional deformation producing the main fabric*

Few major structures were produced during this phase of deformation, but the earlier structures and the diapiric granites were deformed to produce the regional finite fabric. Strain measurements have been made on greenstone belts (Coward & James 1974; Coward, James & Wright, in the Press), and the results are summarized in figure 6. The intensity of deformation varies from over 60 % shortening across the Tati greenstone belt to under 30 % across the greenstone belt at Shabani. East of Shabani the deformation is negligible; the rocks are tilted but show no cleavage, the unconformity of sediments underlying greenstones resting on basement granite is preserved intact and sedimentary structures such as ripple marks and stromatolites are tilted but otherwise undisturbed.

Measurements of the amount of shortening along buckled veins are given in figure 7 for 14 localities in the Tati greenstone belt, 7 localities in the gneisses south of Gwanda and 5 localities in the gneisses southeast of the Lower Gwanda–Antelope belt. These values are

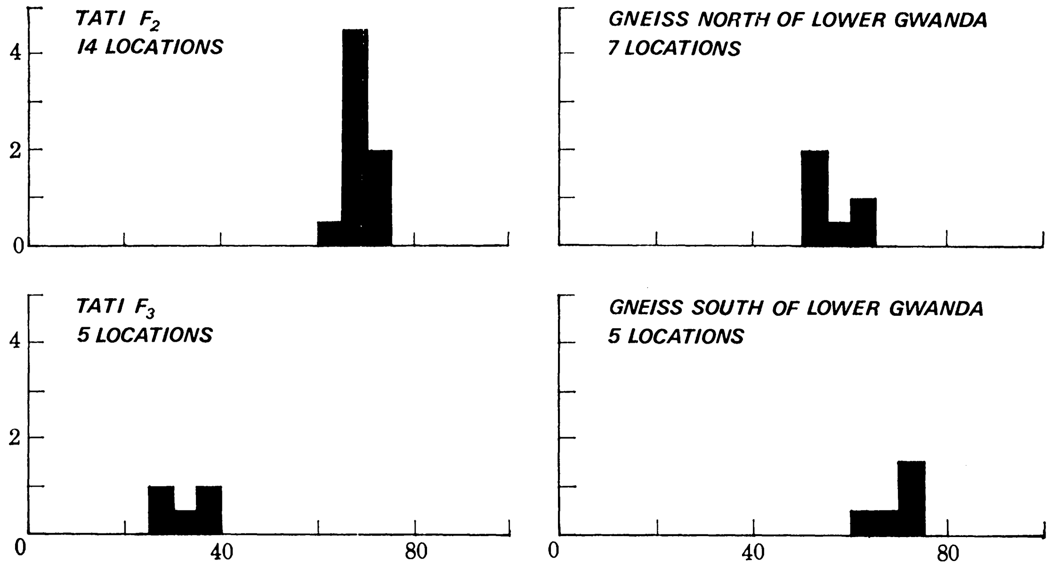

FIGURE 7. Histograms showing the amount of shortening along buckled veins.

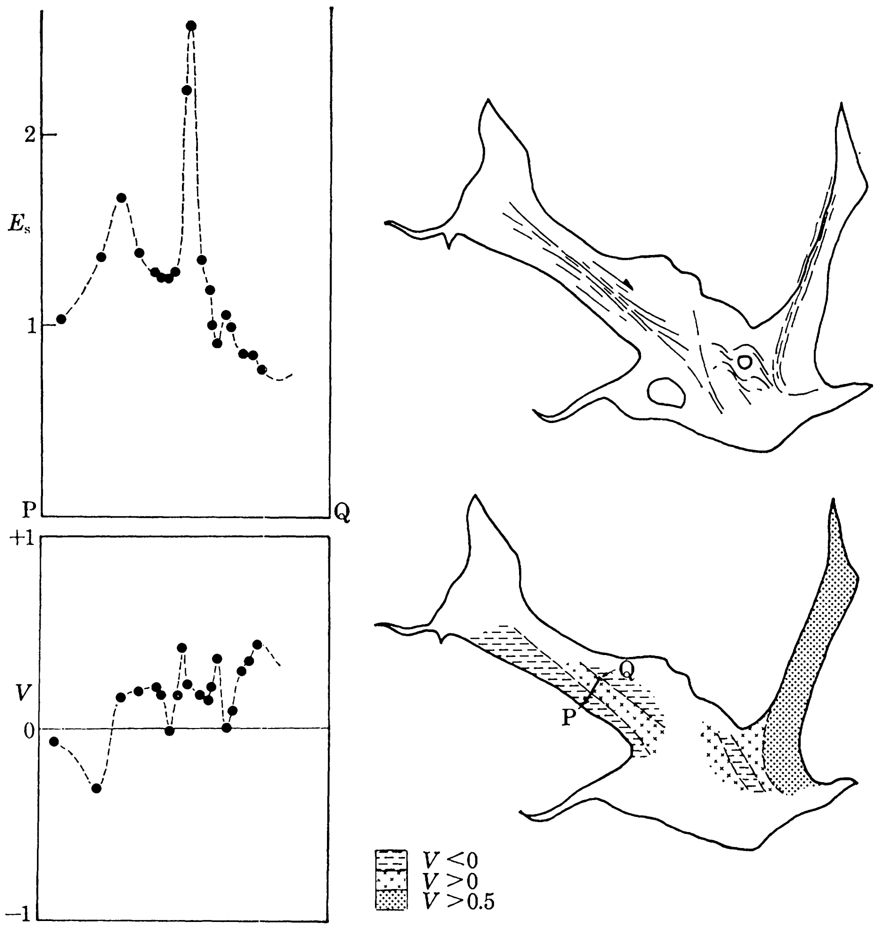

FIGURE 8. Map of the Tati greenstone belt showing the main shear zones, the variation in oblateness of the strain ellipsoid (variation in Lode's unit, V) and a strain profile through part of the northwest arm of the belt.

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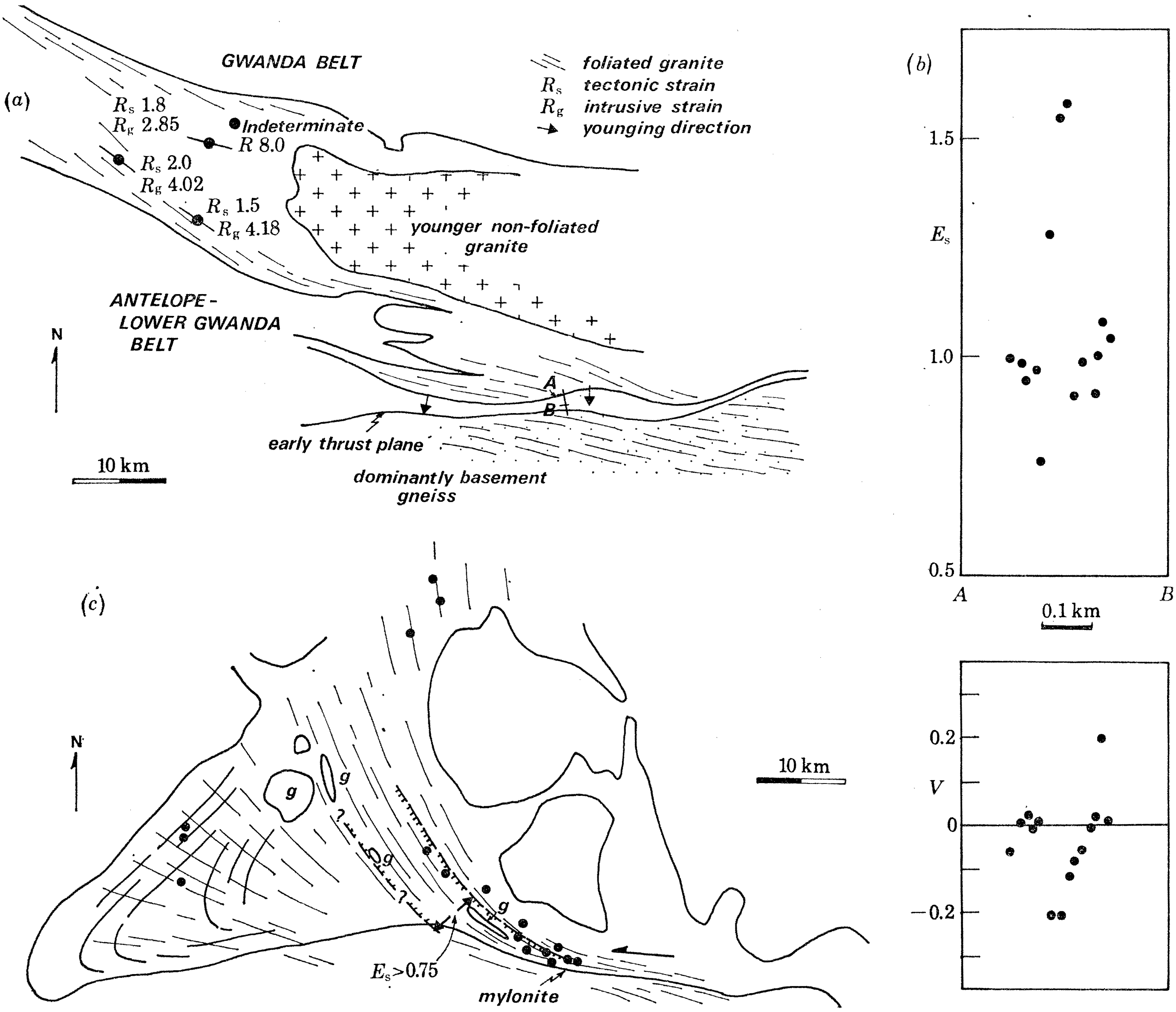

Figure 9. (*a*) Map of the northern part of the Lower Gwanda–Antelope greenstone belt showing how the strain ratios measured from deformed xenoliths on horizontal surfaces in the older foliated granite can be separated into two components, a tectonic strain (R_s) and an earlier strain due to the granite intrusion (R_g). The line indicates the orientation of the principal axis of the R_g strain ellipse. Insufficient outcrop prohibited measurements in three dimensions. In the eastern part of the map, the basement gneiss is shown stippled. Cleavage is shown by narrow lines.

(*b*) Strain profile through the eastern arm of the Lower Gwanda belt. Location of profile section shown in figure 9*a*.

(*c*) Map of the Bulawayo greenstone belt showing the zone of more intense deformation at the front of a shear zone. Small syenitic granites in the Bulawayan greenstones are shown (*g*). Locations of strain measurements shown by the solid circles. (Partly after Amm 1940.)

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a little higher than the values of shortening calculated from measurements of deformation ellipsoids assuming no volume change. This discrepancy may be due to volume change but is more likely to be due to errors involved in measuring the buckled veins and estimating the amount of shortening before fold amplification using a characteristic wavelength/thickness ratio.

At Tati, Lower Gwanda and the southern part of the Belingwe belt, measurements of the deviatoric strain made from deformed objects were combined with measurements of the surface of no finite elongation made from the orientation of buckled and boudinaged veins. These measurements enabled the volume change to be calculated (Barr & Coward 1974). Measurements from the basic rocks at Lower Gwanda and Belingwe give a volume loss of under 5 %, locally up to 12 %.

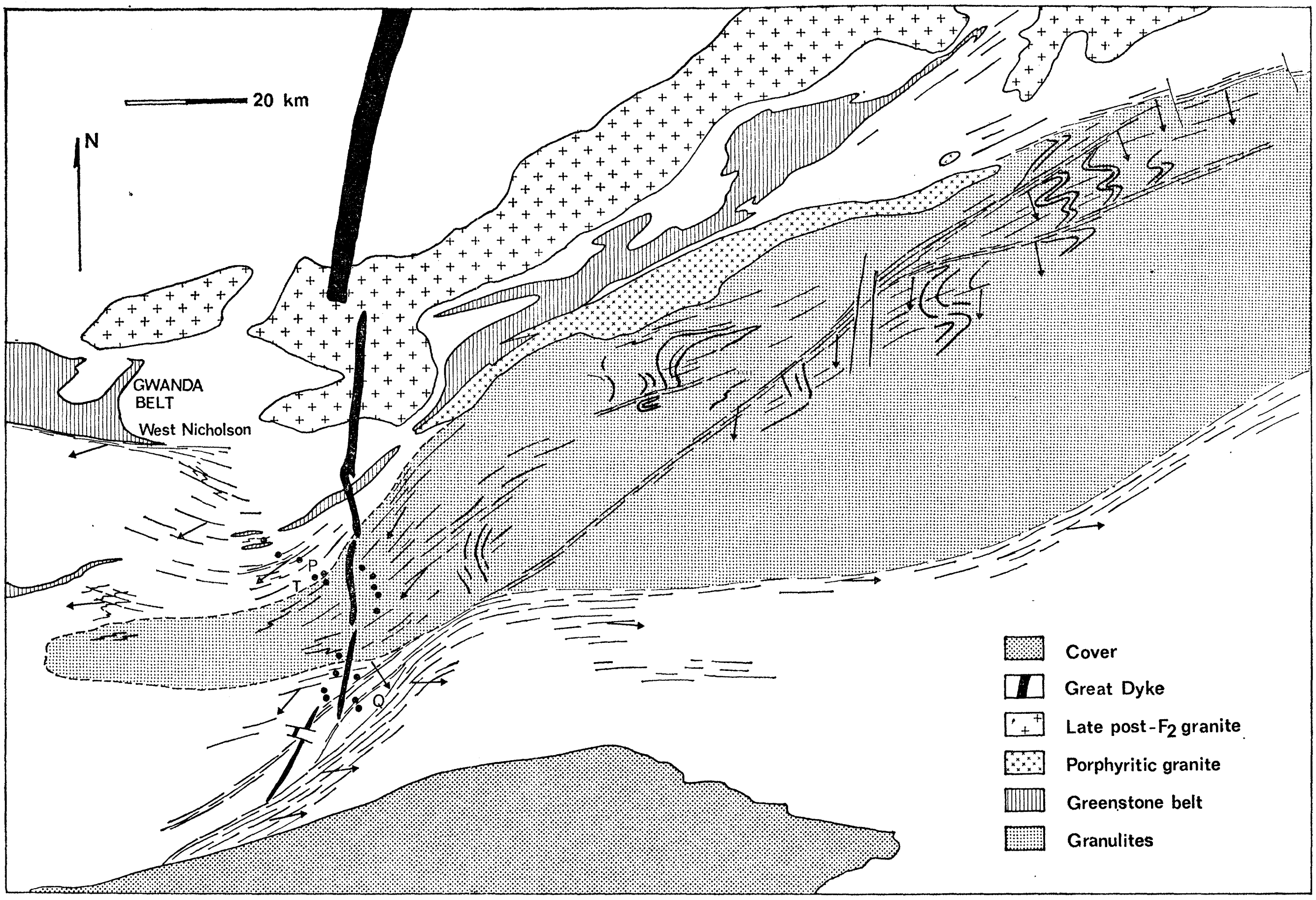

FIGURE 10. Map of the granulites in the southern part of domain 1 and the western part of domain 2. Lithological layering is shown by heavy lines, cleavage by narrow lines. The circles are localities of strain measurements. T, Tod's Hotel. Eastern portion of map after James (1974).

Apart from the regional variation in strain intensity shown in figure 7, there are local variations within greenstone belts. In the northwest arm of the Tati belt (figure 8), a shear zone can be seen from the curvature of the foliation and from the localized more intense strain. Similar variations occur in the Lower Gwanda belt (figure 9*a*, *b*). There is intense deformation along the southern margin of the Bulawayo greenstone belt, with the production of a mylonite fabric. This shear zone dies out along strike to the west and the deformation is taken up by extra flattening within the Bulawayan greenstones (figure 9*c*). This extra flattening can be measured from the shape of deformed agglomerate fragments and also from the shape of small deformed granite diapirs which are elliptical in plan in the zone of more intense deformation (figure 9*c*).

There are also variations in intensity of deformation in the granulites in the southern part of the domain (figure 10). Much of this rock is high-grade, sometimes charnockitic granite with very little tectonite fabric and much of the granite contains undeformed xenoliths of basic or earlier gneissic material. These relatively undeformed granites are cut by steeply dipping shear

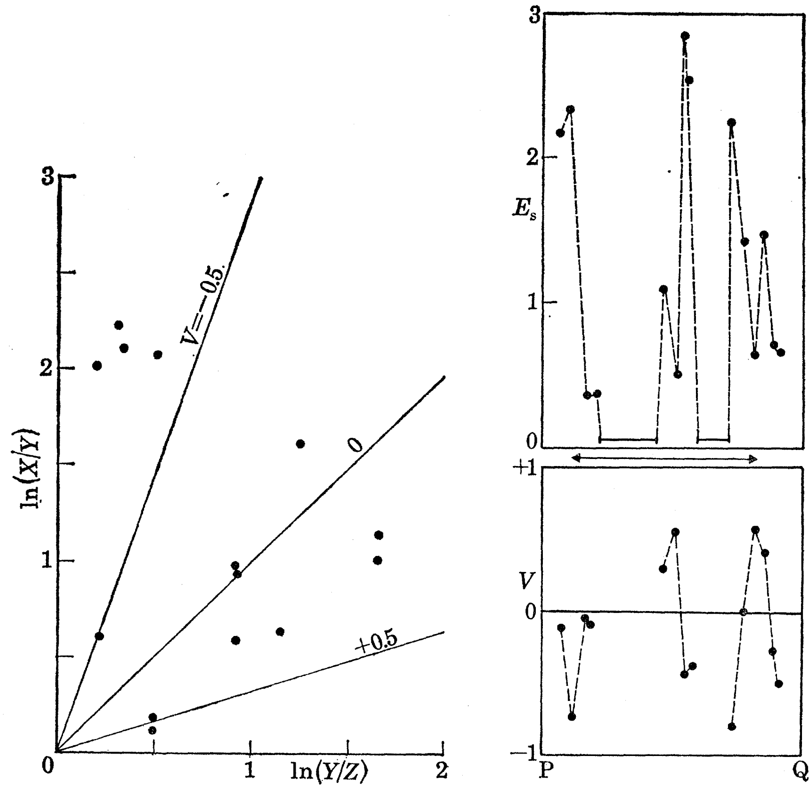

FIGURE 11. (*a*) Strains measured from the granulites and adjacent gneisses on logarithmic deformation plot. (*b*) Strain profile through the southern part of domain 1, section line on figure 10. The horizontal bars represent areas of low-intensity deformation as seen from the paucity of tectonic fabric and relatively undeformed xenoliths in the charnockitic granites. The horizontal arrowed line is 20 km.

zones with an intense tectonite fabric. The strain has been measured from the shape of the xenoliths and from the surface of no finite elongation from folded and boudinaged veins, and estimated from measurements of the shape of quartz leaves; the shape of the original quartz grains changes and becomes more ellipsoidal within the shear zones. The results are plotted in figure 11. In the eastern part of the granulite outcrop there is less granite; the rocks are more gneissic and are folded with low-intensity deformation in the hinges but intense deformation in the limbs (figure 10). Shearing has been concentrated in the limbs of these folds, producing a 'mega-lithon' type of structure like that seen on a small scale in crenulation cleavage.

In the greenstone belt at Lower Gwanda, the finite deformation is close to that of plane strain, but in the other belts the ellipsoids are more oblate, while in the granulites in the south they are more prolate. This deviation from plane strain may be partly due to rotational deformation.

In the Tati belt, increments of simple shear have been added to the regional deformation and the movement direction in the shear zone is normal to the maximum extension direction of the regional deformation, but nearly parallel to the intermediate extension direction. The result is an increase in oblateness with increase in intensity of simple shear (Coward & James 1974).

There do not appear to be any significant regional variations in elongation in the intermediate extension direction which might indicate that the arcuate foliation was due to later bending. There is no shortening in this intermediate extension direction on the inner part of the arc, nor any major increase in extension in this direction in the outer part of the arc. The least deformed rocks occur in the inner part of the arc at Shabani, while the most intensely deformed rocks are in the southwest of the domain. Thus the strain measurements support the contention that the arcuation is a result of an undeformed 'protocraton' in central Rhodesia moving southwest relative to adjacent areas, producing extensive flattening in the southwest of the domain but dominantly simple shear with a sinistral sense of movement in the southeast (figure 3).

Assuming that the anomalously intense deformation in the shear zones can be ignored, and assuming irrotational deformation between the shear zones, a mean shortening of 49 % has been computed for the section between Matsitama and Shabani. This computation assumes that the deformation in the granites and gneisses is similar to that in the adjacent greenstone belts, as it is only in these belts that abundant reliable strain markers have been found. Support for this assumption is given by the agreement between the amount of shortening shown by buckled veins in gneisses adjacent to the Lower Gwanda belt and the amount of shortening shown by deformed objects within the belt (figures 6 and 7).

If all the above assumptions are correct, there must have been movement of the Rhodesian craton some 200 km to the southwest relative to the gneisses of the Limpopo belt. However much more of the deformation may be rotational than has been assumed and this figure probably represents a maximum for the shortening during this phase.

4. *Strains due to late phases of deformation*

Late phases of deformation are important in the west of domain 1 where the main foliation is often crenulated and locally tightly folded (figure 5). The amount of shortening measured along veins folded during this phase, from five locations at Tati, are shown in figure 8. Late folds which fold the cleavage are important in the gneisses south of Tati, where they become the dominant structures in the northwest part of the Limpopo belt. They are also important in the Matsitama belt, where strain measurements indicate intense deformation, with X–Z ratios of over 30:1, partly as a result of this late deformation.

In central Rhodesia, late folds are important in the Selukwe and Shabani areas. At Selukwe, the cleavage-producing deformation is intense in the south, but north of Selukwe township it is very weak and conglomerates in the Selukwe greenstone belt are undeformed. The later folding took place on the north-northeast trending axial planes and deformation is accentuated in shear zones with a similar trend. There are two major shear zones. The Wanderer shear zone in the east has a down-dip lineation. This intense shearing deformed conglomerate pebbles into ellipsoids with the same down-dip maximum extension direction and where the shear zone cuts the ultrabasic rocks south-east of Selukwe, chrome-bearing ore bodies are prolate with the same down-dip extension (cf. Cotterill 1969). The shear zone wraps around the granites producing

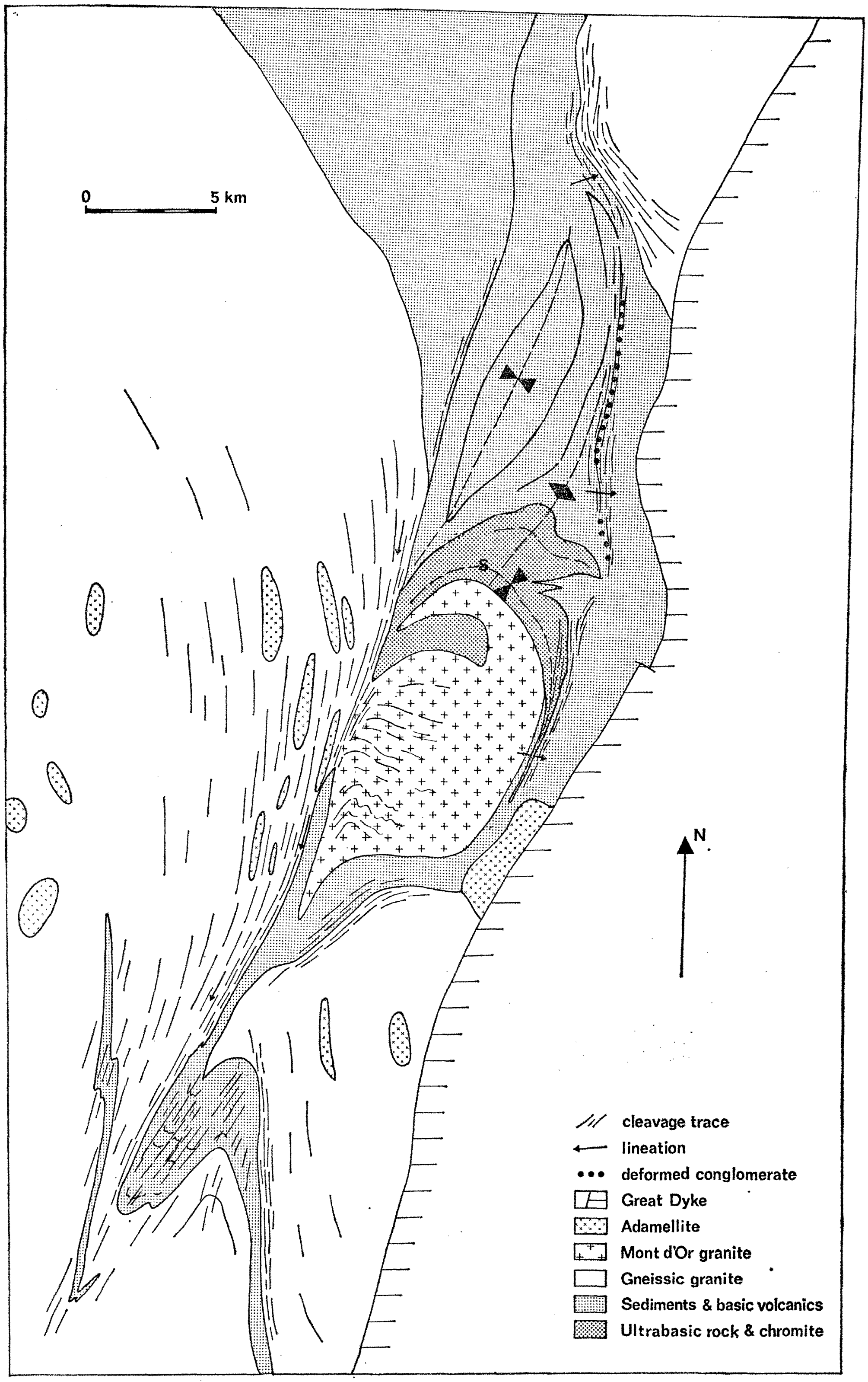

FIGURE 12. Map of the Selukwe greenstone belt, partly after Stowe (1968).

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the most intense deformation at the eastern and western margins of the granites, but leaving areas of low-intensity deformation, pressure shadows, at the northern and southern margins (figure 13).

West of Selukwe, the Surprise shear zone forms a zone of intensely deformed rock more than 1 km wide (figures 12 and 13), which cuts across the granite bodies. The maximum extension direction is nearly horizontal and the sense of movement is dextral. Small adamellitic bodies have been mapped by Stowe (1968) west of Selukwe. Away from the shear zone, these bodies are irregular in shape and orientation, but they become elliptical in plan towards the shear zone, and show a strain ratio of up to 10:1, similar to that given by the shape of deformed quartz grains in the granites and gneisses within the shear zone.

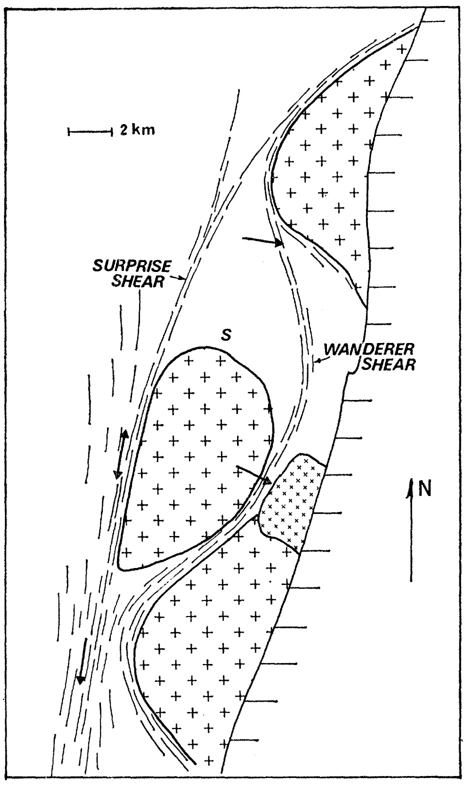

Figure 13. Map showing the two principal shear zones. S, Selukwe.

South of Selukwe, all the structures and fabrics described above are deformed by a ductile shear zone trending east–west, with a sinistral displacement of some 8 to 10 km (Coward, James & Wright, in the Press). North of Selukwe, another east–west trending zone deforms the Umvuma greenstone belt east of the Great Dyke. Wood (1973) recorded intense strains wth 75 % shortening in the z direction and 400 % extension in the x direction from deformed conglomerates within this belt.

The Great Dyke and its satellites cut these east–west shear zones but lie parallel to the north-northeast trending structures; the satellites of the Great Dyke mark the eastern and western limits of this deformation. The Great Dyke has been dated at 2580 Ma (Allsopp 1965; Robertson & van Breemen 1970) and must have closely followed this ductile deformation. This coincidence suggests that the north-northeast trending structures in the greenstones and gneisses may have influenced the position and orientation of the Great Dyke, or that the Great Dyke may have been intruded during a period of stress relaxation following this deformation.

Discussion

Some deformation of the greenstone belts in Rhodesia accompanied the intrusion of the diapiric granites, but important early phases of deformation pre-dated their intrusion. During this early deformation there was imbrication and overturning of parts of the stratigraphic pile and there may have been considerable crustal shortening. However, as intrusive granites obscure much of this early structure and as there has been intense later deformation, it has not been possible to place quantitative estimates on the amount of shortening during this early phase.

This early deformation was followed by regional deformation which produced the main cleavage in the greenstone belts and granites. Strain measurements indicate a mean shortening of 49 % across the cleavage in the section between Matsitama and Shabani and the deformation in the southeastern part of domain 1 was largely by associated simple shear with sinistral sense of movement.

The northern edges of domains 2 and 3 are also shear zones, of different ages but similar trend. Together they give the general impression of the east-northeast 'Limpopo trend'. Shear zones of similar trend occur in central Rhodesia and also in South Africa, where the northern edge of the zone of granulites is a major fault zone, possibly superimposed on an earlier shear zone (Mason 1973). Shear zones with an east-northeast trend occur in the Transvaal within the granulites and also at the northern edge of the Murchison greenstone belt (Graham 1974). These shear zones were formed by different phases of transcurrent movement, where the relative motions of the different crustal segments were generally parallel to the strike of the deformation zones and not across them.

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