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The relation between regionally consistent stretching lineations and plate motions

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Abstract—Regional stretching lineations in the Himalayas and southern Tibet coincide with the known relative motions of the converging plates. This relation is used to infer plate motions from the stretching lineations in the Variscan and Caledonian belts of Europe and in the late Proterozoic belts of northeast and east Africa. In each region, high-angle convergence of plates produced transverse stretching lineations, and high-angle collision was followed by, or combined with, large displacements, first by ductile shear and later by brittle fractures, parallel to the plate boundaries. Apparent stretching, as a result of volume loss (Ramsay & Wood 1973) is common but recognisable as such. Reorientation by later deformations must be allowed for (Ramsay 1967). In ductile shear zones, simple shearing leads to the progressive approach towards the shear plane of the XY plane of the strain ellipsoid and progressive approach of X, the extension axis, to the shear line (Ramsay & Graham 1970). Thus large extensions approximate to the shear line. Large extensions tend to produce sheath folds whose acutely curved axes approach the stretching direction (Quinquis *et al.* 1978). Also, the pre-existence of a strong linear fabric predisposes the development of later folds with axes parallel to the earlier linear fabric. For all of these reasons, the extension axis, X, rather than the fold axis, is of prime significance.

INTRODUCTION

THE VIEW is often expressed by geophysicists that the structures in the continental crust are so complex that they cannot be used to determine plate motions. The contrary view, that motions of crustal units can be inferred from the orientation of planar and linear structures, is not generally accepted by structural geologists, though strongly advocated by some, notably Mattauer (1975), Coward (1976), Mattauer & Mercier (1980) and Andrieux *et al.* (1981). Our aim is to show that, contrary to the pessimistic view of geophysicists, structures in rocks, in particular stretching lineations, can give a clear indication of plate motions. The basic reason for this is that destructive plate boundaries are shear zones on a crustal scale and shear zone geometry involves extension in or close to the direction of shear (Ramsay & Graham 1970).

There has in the past been much disagreement about the relation between the orientation of stretching lineations and the direction of movement. This largely arose from the assumption that folds develop with their axes normal to the direction of movement. The fold axis was used as a primary structural coordinate. Lineations normal to the fold axes were taken to be 'in *a*', parallel to the direction of movement and those parallel to the fold axes 'in *b*', normal to the movement direction. The latter were then explained by arcuation, rolling or superimposed deformation.

The finite strain ellipsoid cannot show the strain path or the incremental strains. Study of fibres grown during the strain (Durney & Ramsay 1973) can do so, or palaeomagnetism and seafloor magnetic anomalies on a different scale, as may the successively produced and spatially separated strain ellipsoids considered here.

In considering the structural indications of relative plate motions we are concerned with four stages: first, oceanic spreading, continuously occurring at an accreting boundary (mid-ocean ridge); second, the continuous subduction of oceanic crust at a destructive margin; third, collision, involving continents, island arcs or both, a process of limited duration; and fourth, post-collisional movements which are responsible for most of the observed structures in orogenic belts.

LINEAR FABRICS DEVELOPED AT CONSTRUCTIVE MARGINS

Evidence of the relation between accretion and fabric comes from seismic anisotropy of oceanic crust and study of fabrics and structures in ophiolites. The direction of maximum P wave velocity in oceanic crust is normal to the generative mid-ocean ridge (Raitt *et al.* 1971) and it coincides with the preferred orientation of olivine *a* axes (Christensen & Cross 1968) which is itself parallel, as in the Semail ophiolite, to a ubiquitous chromite lineation and to axes of tight folds. Thus the chromite lineation and the fold axes approximate to the flow line away from the spreading axis and the foliation to the flow plane (Nicolas *et al.* 1973). The folds are probably sheath folds. The stretching lineation indicates the direction of plate motion.

LINEAR FABRICS RELATED TO DESTRUCTIVE MARGINS

The Himalayas

The plate motions which have formed the Himalayas are accurately known from sea-floor magnetic anomalies, palaeomagnetism, orientations of transform

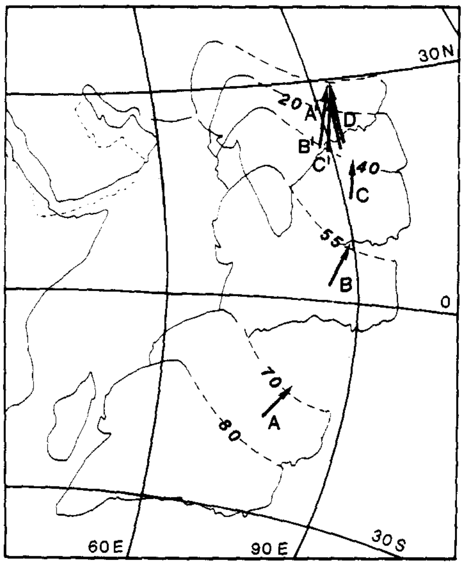

Fig. 1. Inferred motions of India, based on Powell (1979). A, B, C and D show motions at successive dates (shown by numerals in Ma); A', B' and C' are those directions rotated to the present orientation of the Indian plate.

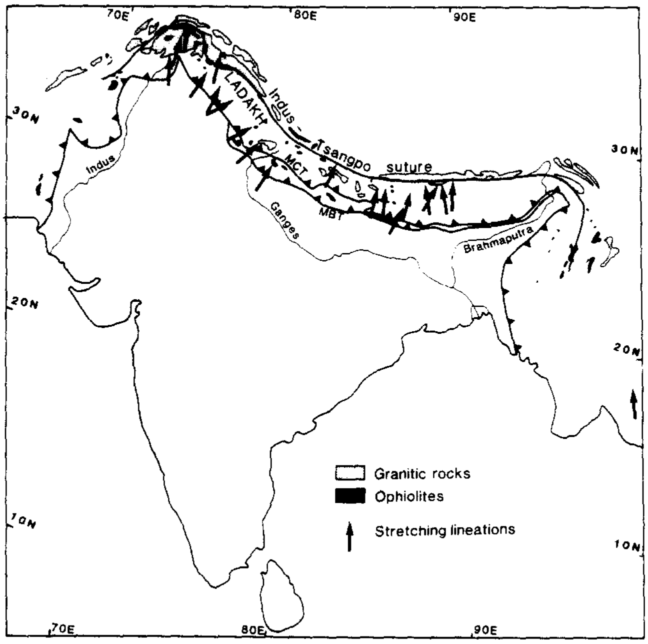

Fig. 2. Stretching lineations in the Himalayas, based on Acharya (1979), Andrieux *et al.* (1981), Bordet *et al.* (1980), Coward *et al.* (1982), Gangopadhyay (1979), Hashimoto (1973), Heim (1935), Mattauer (1975), Powell & Conaghan (1973), Roy (1979), Tapponnier *et al.* (1981), Thakur (1975) and observations by R. M. Shackleton. MCT, Main Central Thrust; MBT, Main Boundary Thrust.

faults and seismicity (Molnar & Tapponnier 1975, 1977, Powell 1979). The inferred motions are shown in Fig. 1.

The collision was essentially head-on, N–S, with no significant component parallel to the plate boundaries other than a small relative motion due to opposite rotations of the Indian and Eurasian plates. The collision is dated stratigraphically as Eocene (the time when the Tethyan ocean between the Indian and South Tibetan plates disappeared [Mattauer 1975]), by the time when sediment from the north was first deposited on the northern margin of the Indian plate (Eocene in Ladakh [Searle in press]), by the time when subduction of oceanic crust stopped, shown by the end of subduction-related calcalkaline magmatism north of the Indus suture to have been pre-Eocene, about 45 Ma (Searle, in press) and lastly by the time when the northward motion of the Indian plate slowed down, 38 Ma ago (Molnar & Tapponnier 1977) or 53 Ma ago (Powell 1979).

Considering that collision is not a sharply definable event, these estimates agree well, at about 50–40 Ma. This time refers to the collision of the Indian plate with the South Tibetan microplate which is probably of Gondwana origin (Tapponnier *et al.* 1981). From the scanty palaeomagnetic dates from this microplate (see Pozzi *et al.* 1982) it appears that the Indian plate moved about 5000 km north relative to the South Tibetan plate between 80 and 40 Ma ago, until collision about 40 Ma ago, after which the Indian plate, from then onward sutured to the South Tibetan plate, moved a further 2000 km north, and rotated about 35° counterclockwise relative to Eurasia. Much of this last northward motion was associated, according to the indentation model (Molnar & Tapponnier 1975) with the lateral expulsion of lithospheric wedges bounded by huge faults. Thus in the Himalayas and southern Tibet, the subduction stage lasted until about 50 Ma ago, the collisional phase from about 50 to 40 Ma ago and the post-collisional stage from 40 Ma onwards. The apparent relative motions, those which should be imprinted as structures on the Indian plate, would trend about 030–020°, 005° and 0° for the pre-collisional, collisional and post-collisional stages. These directions differ from the actual motions of India because the Indian plate was also rotating. Broadly speaking, the successive structures should be seen in zones progressively farther south because the focus of the main deformation shifted southwards, that associated with subduction and collision occurring along the Indus–Tsangpo suture while the main deformation in the Himalayas is post-collisional (Mattauer 1975) from about 30 Ma ago (Oligocene and Miocene) at the Main Central Thrust and still younger on the Main Boundary Thrust.

The orientations of stretching lineations are shown in Fig. 2. These lineations are contained in a schistosity which is often gently inclined so reorientation by later deformations is not in most places significant. There is a clear general coincidence between the orientation of the stretching lineation and the relative plate motions. It is premature to try to establish a more detailed relation to different stages. The motions recorded by the stretching lineation are independent of position on the arc; the motion recorded is to the NNE, although the fold axes curve with the arc. In some places, anomalous stretching lineations occur, for example in the Central Himalayan Gneiss near Manali (77°10′E, 32°15′N) where, after removing effects of later deformation, they trend E–W (Powell & Conaghan 1973). Such anomalies need further study.

The Variscan fold belt of Western Europe

Without the constraint of sea-floor magnetic

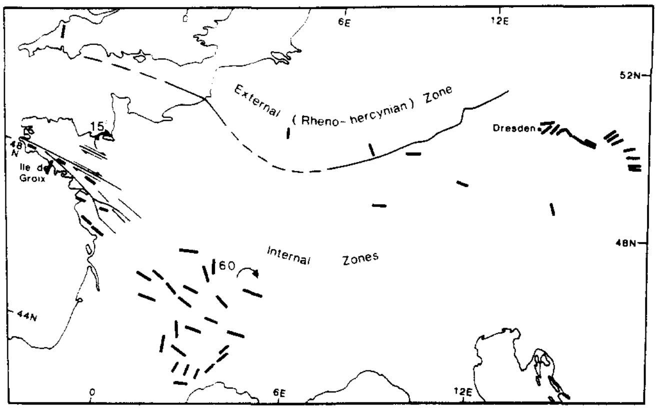

Fig. 3. Variscan stretching lineations, based on Audren *et al.* (1976), Brun & Burg (1982), Burg & Matte (1978), Guillot (1969), Keinow (1933), Korn (1929, 1933) Mehnert (1939), Mukhopadhyay (1973), Quinquis *et al.* (1978), Stenzel (1924) and the authors' observations. Variscan rotations based on Edel *et al.* (1981). Black dashes represent stretching lineations.

anomalies, knowledge of the plate motions which produced Variscan and older fold belts must depend on geological and palaeomagnetic evidence. In Western Europe, palaeomagnetic data are generally accepted to imply the existence of a wide (Rheic) ocean between Laurasia and Pangaea, which disappeared by the late Carboniferous (Irving 1979, Scotese *et al.* 1979), although motions involving very large relative sinistral motions of the two major plates have also been proposed (Morris 1976). A wide ocean between Pangaea and the Iberian–Armorican plate, until the late Carboniferous, is postulated by Perroud & Bonhommet (1981). Large Variscan rotations, of 110° (Ries *et al.* 1980) or 80° (Perroud & Bonhommet 1981), in the Iberian arc, imply sinistral motion of Pangaea relative to Armorica and Laurasia. Variscan rotation also occurred in the Massif Central (Edel *et al.* 1981) but the sense of rotation is uncertain.

The orientation of representative Variscan stretching lineations is shown in Fig. 3. The principal features are listed below.

(1) In the Ile de Groix, Southern Brittany, structures associated with a high P–low T glaucophane–lawsonite assemblage, dated at 380–420 Ma, show N–S stretching lineations and sheath folds and a recumbent schistosity (Quinquis *et al.* 1978). This zone is taken to represent an early Variscan or pre-Variscan subduction zone.

(2) An axial zone, including southern Armorica, the Massif Central, Central Germany and Czechoslovakia where the stretching lineations are predominantly longitudinal, roughly E–W. This E–W stretching has been studied in detail in southern Armorica and shown to be the result of dextral ductile shear (Audren *et al.* 1976). In Vendée, high-grade metamorphic rocks, including blue schists (at Bois de Cené) equivalent to those of Ile de Groix, are thrust southwards over fossiliferous lower Palaeozoic sediments. There is a marked E–W mineral and stretching lineation with elongation up to 300% but near the thrust zones these lineations are rotated towards a roughly N–S direction. Sheath fold axes are parallel to the E–W stretching lineation and nappe transport was westwards (Brun & Burg 1982). In the Massif Central there is a widespread WNW–ESE stretching lineation but near major southward-directed thrusts and also in the northern Montagne Noire N–S stretching lineations occur (Brun & Burg 1982, Burg & Matte 1978). In northern Germany and Czechoslovakia, roughly E–W stretching lineations seem to prevail (Fig. 3). Thus there is evidence of early (pre-Variscan?) southward ductile shear (the Ile de Groix), widespread longitudinal ductile shear, followed by southward ductile shear in thrust zones.

(3) In the northern, non-metamorphic Rhenohercynian zone, N–S or NNW–SSE stretching lineations clearly indicate the direction of movement. Sheath folds are seen in zones of high strain, as at Tintagel, Cornwall, but in general the folds are longitudinal, parallel to the fold belt. The Rhenohercynian zone represents thin-skinned deformation above a décollement surface (Shackleton *et al.* 1982). The shortening above this décollement, *c.* 150 km in southwest England, and the impressive along-strike extent of the décollement, from northern Germany to southwest England and beyond, suggests that it is related to plate motions rather than gravity spreading.

The age relations of transverse and longitudinal Variscan stretching lineations appear to be complex: the transverse motion seen in the Ile de Groix is the earliest but the transverse motion in the Rhenohercynian zone is late Carboniferous; the longitudinal motion in the axial zone is probably mid-Carboniferous. Finally, in the Stephanian or Permian, brittle fractures throughout the Variscan foldbelt represent a distributed transform zone with a dextral shift of about 500 km, connecting the southern Appalachians with the Urals (Arthaud & Matte 1977).

Caledonides of Northwest Europe

In the Scottish Highlands, from the Moine Thrust to the Highland Border, and in the Scandinavian Caledonides, transverse stretching lineations are widespread (Oftedahl 1948, 1950, Kvale 1953, Voll 1960, Hossack 1972, 1976) (see Fig. 4). The transverse motions indicated by these transverse structures persisted or recurred through successive phases of deformation (Voll 1960). However, there are also areas in the central parts of the orthotectonic Caledonides where longitudinal ductile shear, parallel to the belt, prevails. For example in Donegal, northern Mayo and Connemara in Eire, stretching lineations curve from NW to NE in a sense indicating sinistral shear (but dextral north of Clew Bay according to Sanderson *et al.* 1980). In Donegal, this is associated with the intrusion of the Main Donegal granite (Hutton 1982) dated at about 400 Ma. In Shetland, the Funzie Conglomerate in Unst, and conglomerate horizons in the Dalradian Muness Phyllites, show strong longitudinal (NNE–SSW) stretching (Read 1934, Flinn 1956). Progressively increasing strain towards a thrust contact, in the Funzie Conglomerate, was described in detail by Flinn. He attributed it to rolling (Flinn 1956) or to lateral flow of incompetent rock under a thrust nappe

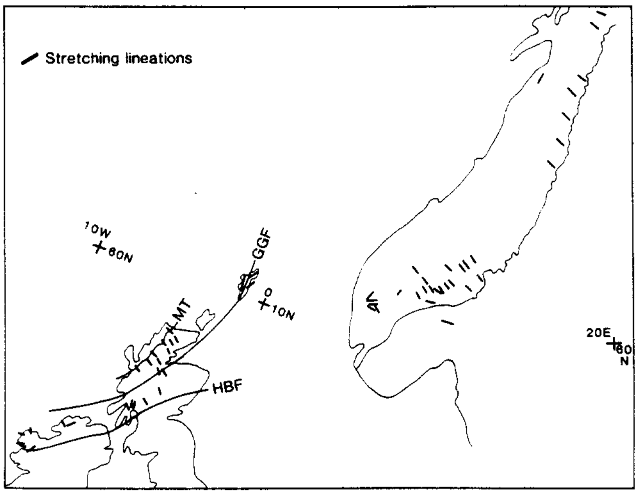

Fig. 4. Caledonian stretching lineations in the British Isles (orthotectonic zone) and Scandanavia, based principally on Flinn (1956), Flinn *et al.* (1979), Hossack (1972, 1976), Hutton (1982), Kvale (1953), Oftedahl (1948, 1950), Sanderson *et al.* (1980) and the authors' observations.

moving WNW (Flinn *et al.* 1979). The geometry described by Flinn coincides with that of a classic shear zone (Ramsay & Graham 1970) in which the overlying thrust mass was sheared towards the SSW relative to the Funzie Conglomerate (Fig. 5). The majority of the early (F_1) NNE–SSW folds which we have studied in the nearby Muness Phyllites of south Unst and Uyea face west but some face east, suggesting that they may be sheath folds. In the Norwegian Caledonides there are zones of longitudinal stretching lineations, and in Spitzbergen, intense longitudinal stretching of boulders in Eocambrian tillites has been described (Harland 1978, Gayer 1969) and attributed to longitudinal shear, consistently sinistral. The concept of transpression, or oblique relative motions of plates, was developed on the basis of this Spitzbergen evidence (Harland 1971). It is clear that within the Caledonides of Northwest Europe there are major zones of sinistral ductile shear. It has been suggested that this may be measured by an apparent 1500 km sinistral offset of the pre-Caledonian Grenville front (Shackleton 1979).

Finally, in the Caledonides there are brittle fractures parallel to the belt, with large sinistral displacement. On the Great Glen Fault there is thought to have been a sinistral displacement of about 160 km after the main

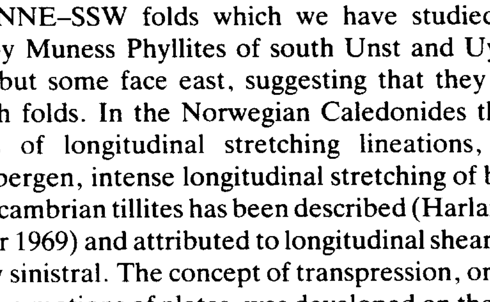

Fig. 5. Section of the Funzie conglomerate, Fetlar, Shetland, drawn parallel to the stretching direction (X). Ellipses represent XZ section of strain ellipsoid, and lines the attitude of the schistosity. Based on data given by Flinn (1956).

regional metamorphism (Winchester 1973) and of about 100 km after an early Devonian or late Silurian granodiorite (Kennedy 1946).

Thus in the Caledonides, transverse and longitudinal ductile shearing was followed by brittle shearing, the longitudinal displacements being consistently sinistral.

Upper Proterozoic: Northeast and East Africa, West Africa and Brazil

A very extensive region including Saudi Arabia, Eastern Egypt, Eastern Sudan, and most of Kenya, Eastern Tanzania and parts of Mozambique (the Mozambique Belt of Holmes 1948) is underlain by rocks which yield late Proterozoic (1000–450 Ma) radiometric dates, and so far, except near the western margin, no reliable Rb/Sr ages of more than 1000 Ma. This is part of the Pan-African domain of Kennedy (1961).

The western margin of this Pan-African fold belt is well-defined, as a N–S boundary against the Archaean Tanzanian and Zimbabwe cratons to the west, by structural, metamorphic and radiometric evidence (Johnson 1967, Hepworth *et al.* 1967, Hepworth & Kennerley 1970, Sanders 1965, Meinhold 1970, Pallister 1971). Farther north, this western margin has not been defined. In Egypt, it presumably runs through the Western Desert somewhere between Aswan and Uweinat where 2000 Ma granulites are known (Klerkx & Deutsch 1977).

Along the western front of the Mozambique belt, on the Mozambique–Zimbabwe frontier, and from northern Malawi through Tanzania and western Kenya, there is a pervasive and intense stretching lineation which persistently trends NW–SE to WNW–ESE, usually plunging SE. This trend makes a high but oblique angle with the overall N–S Mozambique front. Pebbles in conglomerates are drastically stretched, as at Magulilwa near Iringa in southern Tanzania. The NW–SE stretching lineation can be traced at least 200 km eastwards from the Mozambique front in central Tanzania. It is parallel to tight folds. The general eastward dip of the associated schistosity, and the eastward dip of the thrusts in the marginal zone of the belt (Sanders 1963, 1965) shows that the relative motion of the overlying plate was towards the northwest.

Diverging northwestwards from the Mozambique front are two, long, straight zones of intense deformation, the Ubendian belt and the Aswa zone (see Fig. 6). The Ubendian belt, like the Mozambique belt in southern Tanzania, is polycyclic. The structures and the rock units ('Usagaran' of the Mozambique belt) curve round the southeast margin of the Archaean Tanzanian craton (Meinhold 1970) and the NW–SE linear structures of the Mozambique belt appear continuous with, and indistinguishable from, the intense, gently plunging NW–SE lineations within the Ubendian belt. Similarly the stretching lineations at the margin of the Mozambique belt in western Kenya are continuous with those in the Aswa zone of Uganda. The lineations in the Ubendian and Aswa zones plunge at low angles and the planar

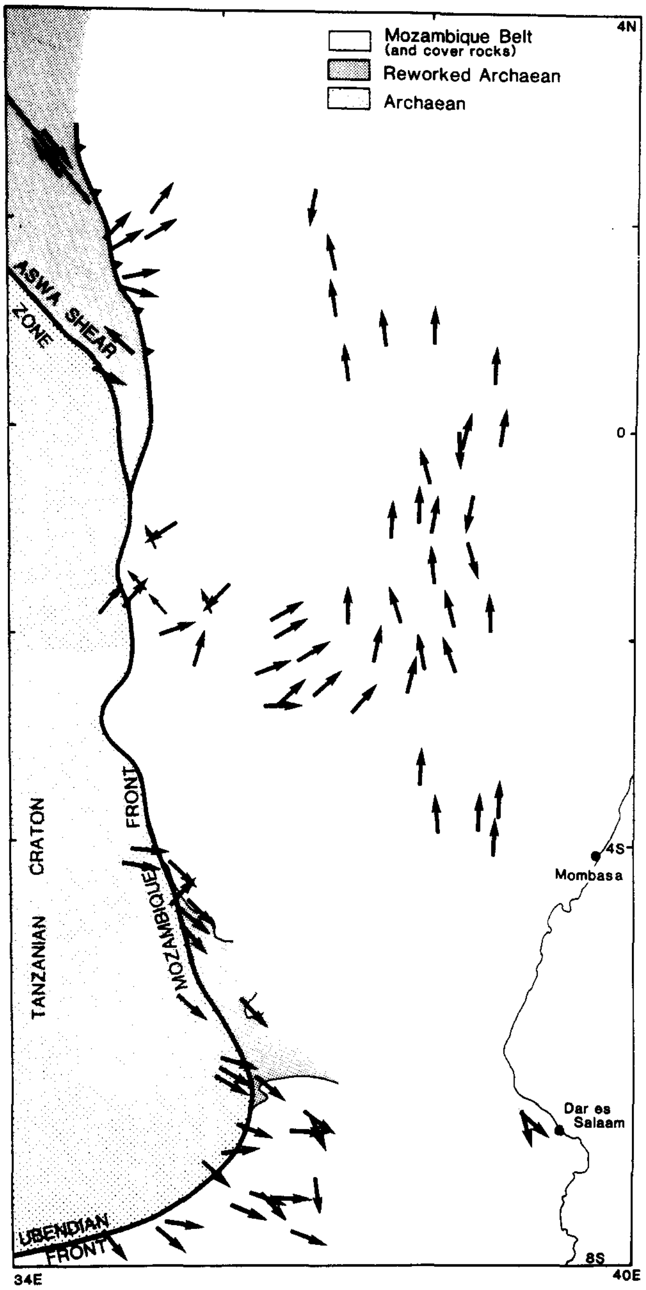

Fig. 6. Stretching lineations in the Mozambique belt of Tanzania and Kenya. Based on Hepworth (1967), Hepworth & Kennerley (1970), Meinhold (1970), Pallister (1971), Saggerson *et al.* (1960), Sanders (1963, 1965), Shackleton (1946) and the authors' observations.

fabrics are steep. These two zones are interpreted as zones of strike-slip movement, possibly transform faults. The Aswa zone appears from the strikes curving into it to be sinistral (Hepworth 1967).

A NW–SE trending stretching lineation, usually plunging SE, is also found in the Upper Proterozoic rocks in the Eastern Desert of Egypt. We interpret this array of consistently oriented stretching lineations, from the western border of Mozambique to Eastern Egypt, and in the Ubendian and Aswa zones, as indicating the directions of relative motion of large rigid lithospheric plates, although not necessarily a single pair of plates since there is evidence of several parallel sutures within the Pan-African domain. The motions cannot be attributed to gravity spreading because they are consistently oblique to the Mozambique front and extend to shear zones within the Archaean cratons.

In a large area of the Mozambique belt of central Kenya, the mineral and stretching lineations trend N–S or NNW–SSE parallel to the Mozambique belt (Shackleton 1946) (Fig. 6). These longitudinal lineations we interpret as indicating that there was N–S relative motion of the plates, parallel to the belt. The age relations of these longitudinal stretching lineations to the transverse ones nearer to the Mozambique front is not yet known, nor the sense of motion.

The Upper Proterozoic of West Africa (Algeria, Mali, Nigeria, Ghana) and the contemporary Brazilides in northeast Brazil show similar patterns of stretching lineations, with predominant longitudinal stretching, often very large, parallel to the fold axes in the interior of the belts but transverse near the margins, as in Ghana. Longitudinal (northward) thrusting of a granulite massif in the interior of the Pan-African of Mali has been postulated (Boullier *et al.* 1978). Finally, in West Africa there are many very large late Pan-African brittle longitudinal mylonite zones and faults, variable in sense of displacement.

CONCLUSIONS

(1) Data from older orogenic belts (Variscan, Caledonian, and especially the late Proterozoic of northeast Africa) supports the view that stretching lineations reflect the direction of relative motions of lithospheric plates.

(2) Transverse collision was often followed by ductile shear parallel to the plate boundary and this by brittle shear in the same direction and with the same sense.

(3) Shearing parallel to the plate boundaries may have been located and controlled by the higher ductility resulting from collision. Since these ductile interplate shear zones were, at least initially, gently inclined, the ductile shear parallel to the plate boundary often occurs on gently inclined surfaces. The later brittle shearing parallel to the plate boundaries is usually on steep planes.

We emphasise that this discussion is generalised and simplified. It is intended to show the validity of a method but it is only by very detailed work that the relative motions of crustal units will be understood.

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