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

The Limpopo Mobile Belt - Southern Africa

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The Limpopo mobile belt – southern Africa

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The Limpopo belt is an extensive ENE-trending linear zone of high-grade metamorphic tectonites which separates the Archaean nucleii of the Rhodesian craton to the north from the Kaapvaal craton to the south. The belt consists of reworked Archaean granite–greenstone terrain with an early Proterozoic cover sequence, the Messina Formation, infolded and metamorphosed with the basement. Two major zones of shearing and transcurrent dislocation separate marginal granulite zones from a central zone which consists of complexly infolded cover rocks and reworked basement. The northern granulite zone appears to grade transitionally into the Rhodesian craton to the north, whereas there is some evidence that the southern granulite zone is faulted against the Kaapvaal craton to the south. The whole belt has behaved as a zone of crustal weakness throughout geological time, and is characterized by repeated shear deformation, igneous intrusion and extrusion, despite the cessation of major regional tectono–thermal reactivation about 1900 Ma ago.

1. Introduction

During his travels in southern Africa between 1865 and 1872, the German explorer Carl Mauch noted a belt of gneissic rocks situated between latitudes 21° S and 23° S, in what is now eastern Botswana. Mauch recorded these rocks on his 'Geological Map of the Transvaal and Rhodesia' (Harger 1934), and recognized the distinction between this gneiss belt and the 'gold belt' (granite-greenstone) terrain of Rhodesia to the north. It was not until 1953 that Macgregor referred to the 'Limpopo orogenic belt' as an entity, following Holmes's paper on orogenic belts in Africa (Holmes 1951), and Cox *et al.* (1965) stimulated further interest in the Limpopo belt by recognizing a megascopic tectonic zonation within the belt. The first systematic geochronological study of the Limpopo belt was undertaken by Van Breemen (1968), who revealed a complex sequence of tectono-thermal events with major age groupings at around 2000 and 2650 Ma.

Much of the present knowledge of the Limpopo belt is derived from the work of the South African, Botswana and Rhodesian Geological Surveys, and a comprehensive list of references to this work, and other geological investigations within the belt, is contained in the references. This present paper is an attempt to present a unified account of the Limpopo mobile belt as a geotectonic entity, and as such it incorporates previous work together with extensive geological and photogeological investigations over most of the belt by the author.

2. Outline of the geology

The Limpopo belt is a major zone of high-grade metamorphic tectonites which separates the Archaean granite–greenstone terrains of the Rhodesian and Kaapvaal cratons (figure 1). It extends along an ENE axis for over 560 km and varies in width between 240 and 320 km. At the eastern extremities of the mobile belt younger sedimentary and volcanic formations conceal the junction between the Limpopo belt and the Mozambique belt, which is however manifested by the Nuanetsi igneous province and the spectacular westward inflexion of the Lebombo monocline. The western extremities of the Limpopo belt are covered by the Kalahari Basin.

In its simplest form the Limpopo belt consists of a reworked strip of Archaean granite–greenstone terrain, together with the remnants of an early Proterozoic cover sequence (the Messina Formation) which has been metamorphosed and deformed with the pre-cover basement. There is a distinct zonation of the mobile belt (figure 2) with marginal granulite zones consisting mainly of reworked basement, separated from a central zone by major shear belts or 'straight zones'. The central zone is characterized by complex cover–basement relationships between the Messina Formation and the reworked basement and includes major anorthosite complexes which have been reworked with the cover rocks.

Tectonically controlled basins (clastic wedge-type) of Waterberg system (Proterozoic) and Karroo system sedimentary and volcanic rocks have been developed subsequent to the cessation of any major tectono-thermal reactivation within the belt, but there is evidence of continued instability along the belt between 2000 Ma and the present day (e.g. faulting and igneous activity).

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2.1. *The northern marginal zone*

The northern marginal zone falls entirely within Rhodesia and Botswana and is subdivided into a granulite subzone and the Tuli-Sabi shear belt (figure 2). The granulite subzone grades transitionally into the Rhodesian craton to the north and can itself be further divided into a well-preserved granulite–charnockite terrain adjacent to the craton and a more sheared retrograde granulite terrain adjacent to the Tuli-Sabi shear belt. In eastern Botswana the granulite–charnockite terrain dies out laterally into retrograde granulite terrain, and the southwestern corner of the Rhodesian craton is completely flanked by reworked terrain extending into the Shashe mobile belt (Crockett 1968).

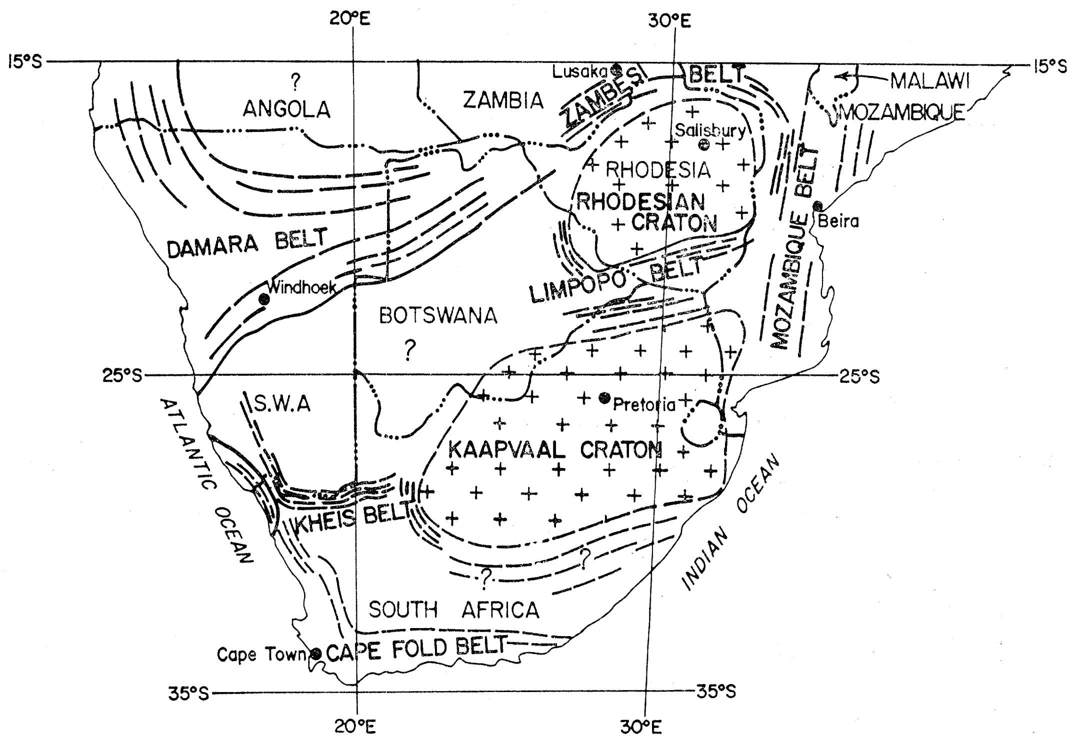

FIGURE 1. Tectonic setting of the Limpopo mobile belt, southern Africa.

(*a*) *Granulite subzone*

The granulite terrain of the northern marginal zone consists mainly of reworked cratonic granite–greenstone material of Archaean age. Satellite dykes of the Great Dyke system intrude the granulite terrain and these have been dated at 2600 ± 120 Ma by Van Breemen (Robertson & Van Breemen 1970), which thus sets a minimum age for the granulite metamorphism. MacGregor (1951) appears to have been the first to suggest an Archaean origin for the granulite terrain, and Worst (1962) and Robertson (1972*a*) have presented convincing evidence of the Archaean (granite–greenstone) origin of the granulites and related rocks. The reworked nature of the northern marginal zone as developed in Botswana was first suggested by Gerrard (1965), later supported by Mason (1970) and Bennett (1971*a*).

Over much of the granulite terrain in Rhodesia there is remarkable preservation of granulite facies mineral assemblages and textures, although the shearing associated with the Tuli-Sabi shear belt (see later) has caused widespread retrogression in the southern parts of the zone.

As the granulites extend westwards into Botswana, shearing has once more caused retrogression, leaving enclaves of granulite rocks surrounded by retrograde granulites.

A full range of granulite rocks has been identified in the zone by Worst (1962), Swift, White, Wiles & Worst (1953) and Robertson (1968, 1970, 1972*a*), ranging from acid gneisses to basic and ultrabasic rocks. Charnockites are developed locally (usually as cores of major domal structures) and charnockitic gneisses are widespread. Pyroxene granulites and associated basic granulites, together with interlayered gneissic granulites and magnetite quartzites, represent equivalent lower grade greenstone belt material of the Rhodesian craton. Despite the intense metamorphism, the magnetite quartzites retain the finely banded nature and intricate fold structures of their cratonic equivalents, the banded ironstones. The gneissic granulites are characterized by honey-brown coloured felspars, bluish quartz crystals, the development of pyroxenes and predominantly granoblastic textures.

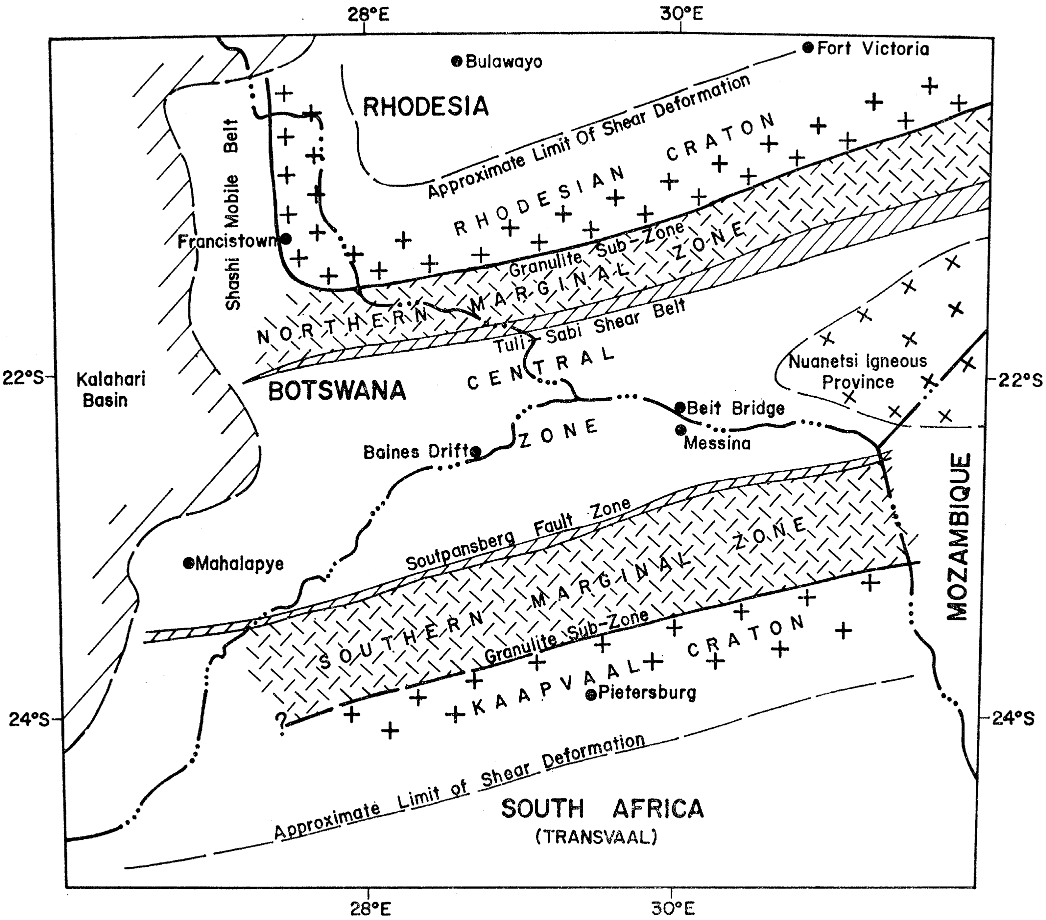

Figure 2. Tectonic subdivision of the Limpopo mobile belt.

In a large area south of Buhwa, Worst (1962), Mason (1970) and Robertson (1970, 1972*a*) have outlined a series of major ultrabasic complexes which predate the granulite metamorphism and contain large podiform chromite bodies. These complexes are intimately associated with recognizable greenstone belt material, and probably represent layered differentiated bodies emplaced within the Archaean greenstone sequence.

The tectonic style of the granulite terrain emphasizes the differences between this terrain and the cratonic terrain to the north. Early isoclinal folds are refolded by major open folds which often close into domal structures. The most prominent structural element is a variably developed coarse foliation which presents a marked contrast to the fine schistosity and crenulation cleavage of the cratonic greenstone belts. There is strong ENE structural grain throughout the zone, with regional foliation trends and major fold axes parallel or subparallel to the grain.

An interesting feature of the northern margins of the granulite subzone is the development of an intrusive porphyritic granite (Worst 1962; Robertson 1972), which occurs as narrow, elongated bodies subparallel to the margins of the mobile belt. The granite is frequently sheared and often assumes the character of a prophyroblastic granitic gneiss. Similar developments of this granite or porphyroblastic gneiss have been mapped along the northern margins of the mobile belt in Botswana (Mason 1970; Bennett 1971). Worst (1962) suggested that the granite was intruded along a major thrust plane which marked the edge of the mobile belt, but subsequent work has shown this hypothesis to be doubtful and, in fact, there is little evidence of a major dislocation anywhere along the northern edge of the mobile belt. However, the porphyritic granite is significantly located at or close to the craton–mobile belt interface.

(*b*) *Fracture systems*

The structural grain of the subzone lies parallel to the major ENE trend developed throughout the mobile belt, and is emphasized by fractures with that trend. A complementary WNW trending set of fractures is developed, and many of these fractures are filled by basic dykes, as are many of the ENE set of fractures. In addition to these fractures, a major fracture system with NNE and NNW trends, is developed in the granulite subzone. This system is related to the Great Dyke fracture system (Robertson & Van Breemen 1970; Mason 1970) and extends through the Rhodesian craton to the north. Significantly, this fracture system is not developed in the central zone of the mobile belt, but it is again developed in the southern marginal zone and on the Kaapvaal craton (see later).

(*c*) *Monachane and Towla fragments*

The Monachane and Towla fragments occur in Botswana and Rhodesia respectively, and are considered to represent tectonic fragments of Messina Formation cover rocks resting on a granulite basement. Mason (1970) has described the Monachane fragment, which constitutes a complexly refolded and overturned anticlinal structure developed in metaquartzites, magnetite quartzites, marbles, amphibolites and quartzo–felspathic gneisses of the Messina Formation. Along the southern margins of the fragment, basement gneisses occur in juxtaposition with the cover rocks and thin slivers of basement have been pushed into the adjacent cover rocks. Although the cover rocks are characterized by cataclastic fabrics, they have clearly not suffered the complex deformational history of the surrounding basement gneisses.

The Towla fragment is poorly exposed, but nevertheless shows many of the features characteristic of the Monachane fragment. There is a remarkable similarity in the sequence of rock types exposed here and sections through the Messina Formation in the Beitbridge area to the south (central zone), although Mackie & Oosthuizen (1972) and Robertson (1972*a*) maintain that these rocks are related to the Archaean greenstone belt sequences developed to the north. The tectonic style of the Towla fragment is certainly not typical of reworked basement, and the amphibolite facies assemblages here appear to be pro-grade, despite the surrounding granulite terrain. Another curious feature of the Towla fragment is that it interrupts the Tuli-Sabi shear

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belt, which is very strongly developed on either side of the fragment. This again suggests that the fragment is unlikely to represent Archaean greenstone belt material.

(*d*) *Shashe mobile belt*

The Shashe mobile belt is a continuation of the main Limpopo belt, which wraps around the southwestern margins of the Rhodesian craton (Crockett 1968; Bennett 1970). This appendage of the Limpopo belt consists of a reworked basement (cratonic material) with infolded remnants of Messina Formation cover rocks, including metaquartzites, marbles and associated quartzo–felspathic gneisses. The Matsitamma schist belt (Bennett 1970*b*, *c*) appears to represent a large area of Messina Formation rocks, preserved at a lower grade of metamorphism than the mobile belt proper, resting on and infolded with a gneissic basement. The southeastern parts of the schist belt grade into typical high-grade Messina Formation rocks into the Shashe area (Crockett 1968). The tectonic style of his belt is more akin to that of the central zone of the Limpopo belt rather than that of the northern marginal zone. The Shashe belt is covered to the west and north by younger formations.

Appendages of the Tati schist belt (Archaean/cratonic) extend as tectonically thinned zones into the Shashe mobile belt, where basic greenschists have been upgraded to two-pyroxene amphibolites within relatively short lateral distances from the edge of the Rhodesian craton (Mason 1970). The Messina Formation cover rocks, although complexly folded, have clearly not suffered the extremely complex deformational history of the reworked basement with which they are infolded, and this can be seen very well in the areas immediately south and southwest of the Tati schist belt.

(*e*) *Tuli-Sabi shear belt*

Extending almost the entire exposed length of the mobile belt and varying in width between 5 and 50 km, the Tuli-Sabi shear belt constitutes one of the largest known shear belts ('straight zones') on the African continent. Within the belt there is evidence of repeated movement over long periods of time both during and after major tectono–thermal reactivation. The shear belt is characterized by thin flinty crush bands which occur as curvilinear features separating strongly sheared segments of the country rocks (mainly flaser gneisses and mylonitic gneisses) thus imparting a distinctly braided character to the belt. These cataclasites dip consistently southwards towards the central zone of the mobile belt with generally steep dip angles (> 50 and often > 70°). However, local 'flat' zones have been observed in some areas. Both basement and cover rocks can be identified in the shear belt, but refoliation is often so intense as to transform the original granitic rocks into flaggy, dark-grey gneisses and individual layers in the shear belt show extreme attenuation and fragmentation, especially cover rocks such as metaquartzites and marbles, and the anorthositic rocks.

Early isoclinal folds have been largely obliterated by subsequent refoliation, but there is still evidence of repeated isoclinal folding within the shear belt and the development of flattened 'eye' folds is common. Strongly developed linear fabrics, fold mullions and rod structures tend to plunge gently either way along the shear belt and rapid plunge reversals are commonly encountered.

There is evidence of a complex history of recrystallization and retrogression of existing high-grade metamorphic assemblages, with the generation of large quantities of hydrous minerals (especially micas and chlorites) (Mason 1970).

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One of the features of the Tuli-Sabi shear belt are the numerous small and large shear splays which penetrate the granulite terrain north of the main shear belt. To the south of the shear belt, cataclastic textures characterize rocks of the central zone up to 10 km south of the Tuli-Sabi fault system, but refoliation is only developed immediately adjacent to the fault zone.

Movements along the shear belt have involved vertical, horizontal and rotational components at various stages in its evolution. There is an obvious dextral transcurrent component, and the reactivation of the Tuli-Sabi fault zone in Post-Karoo system times involved major movements with downthrows to the south of over 1000 m (Mason 1969).

In a ductile shear belt of this magnitude it is difficult to prove quantitative displacement of the marginal zone and craton, relative to the central zone of the mobile belt, although very great displacement of fragments caught up in the shear belt may have taken place. It is suggested that displacements of major crustal segments on either side of the shear belt here may be illusory, and that most of the strain and displacement of material was taken up in the shear belt itself. The complex isoclinal folding and associated deformation could also be explained by flowage while the shear belt was behaving in a ductile manner. Theoretical premises for the calculation of strain and displacement in small-scale shear belts (Ramsey & Graham 1970) will probably be invalid when applied to ductile shear belts of the magnitude of the Tuli-Sabi shear belt, because (*a*) the belt has behaved in a ductile manner and there is evidence of cataclastic flowage and (*b*) it is unlikely that the belt has formed by simple shear.

2.2. *The central zone*

Flanked to the north and south by major shear belts and granulite terrains, the central zone of the mobile belt has emerged as a distinct entity – in effect a mobile belt within a mobile belt. The zone is characterized by complex relationships between reworked basement and a cover sequence known as the Messina Formation. Major anorthosite complexes occur within the central zone, and in each case these complexes appear to have intruded the Messina Formation. Metamorphism and deformation of the zone during at least two major tectono–thermal events, has resulted in complex basement-cover relationships, and the development of a distinctive tectonic style.

The large-scale drag of the formations in the central zone into the Tuli-Sabi shear belt was noted by Mason (1969) who used this as evidence for dextral movement along the shear belt. Although not as pronounced, there is a similar drag of the formations into the Soutpansberg fault which again indicates a dextral sense of movement.

(*a*) *Reworked basement*

Recognition of possible reworked Archaean granite-greenstone terrain in the northern Transvaal was first made by Sohnge (1946). Recently Bahnemann (1971, 1972) has pointed out the existence of distinctive reworked basement zones in the Messina area, and Mason (1970) and Bennett (1971 *b*) recognized the existence of similar areas within the central zone in eastern Botswana. The reworked basement within this zone is typically retrograded granulite terrain, characterized by extremely complexly deformed gneisses, by homogenization of large areas through partial melting, and by the development of basic dykes which pre-date the Messina Formation. Bahnemann (1971) states that the basement gneisses in the Messina area are ‘generally devoid of garnets’ and he uses this evidence as part of his criterion for the distinction between basement and Messina Formation. This criterion cannot be applied throughout the mobile belt, and the highly garnetiferous nature of the reworked basement of the granulite terrains in the belt suggests that the presence or absence of garnet in the central zone can hardly be used as a valid distinction of cover and basement.

Several major areas of basement can be distinguished within the central zone, and these contain no recognizable cover rocks (Messina Formation metaquartzites, marbles and associated amphibolites, anorthositic gneisses and quartzo–felspathic gneisses). The development of massive porphyroblastic granitic gneisses and migmatites with cores of homogeneous post-tectonic granite are typical of these areas (the Bulai granite in the Messina area and the Mahalapye granite–migmatite complex in eastern Botswana are good examples). Sheared basic and ultrabasic relics (probably ex-greenstone belt material) occur within the basement gneiss areas, and several generations of basic dykes can be recognized, although the earliest dykes are only represented by intensively deformed mafic scheiren within the gneiss. Detailed studies of these basement areas are needed to establish a more precise sequence of dyke intrusion and deformation.

Many areas within the central zone include strips and fragments of basement infolded with the Messina Formation, but large areas occur where no basement protrudes through the cover. Obviously such variations within the central zone depend on the degree of basement reactivation in any given area and on the present level of erosion in such areas.

(*b*) *The Messina Formation*

The introduction of the term 'Messina Formation' was made by Sohnge, le Roex & Nel (1948) who took the type area around Messina in the northern Transvaal. Originally the Formation was restricted to a sequence of meta-quartzites, magnetite quartzites, dolomite and marble. Associated quartzo–felspathic gneisses and amphibolites were mostly interpreted as 'older granites and gneisses' and 'basic and ultrabasic intrusives', the latter also including serpentinites, anorthositic gneisses, hornblendites, pyroxenites, peridotites and hornblende gneisses. Tynedale-Biscoe (1949) remarked on the close similarities of the rocks exposed in the Beitbridge area of Rhodesia to those around Messina, and McConnell (1956) was first to recognize the similarities of rocks in eastern Botswana (then Bechuanaland) to the Messina Formation as developed in the northern Transvaal. The Messina Formation in eastern Botswana was re-defined (Mason 1965*a*, 1967) to include quartzo–felspathic gneisses, amphibolites and other rocks intimately associated with the metaquartzites, marbles and calc-silicate rocks as a distinctive cover sequence, and the term 'Baines drift formation' was introduced. The basement areas (not recognized as such at that time) were mapped within a 'gneissic formation'. Bennett (1971*b*) has attempted a subdivision of what he calls the 'Tectono–metamorphic complex of eastern Botswana', but the recognition of basement and cover has been neglected and the result is rather confusing.

The Messina Formation can be regarded as a predominantly metasedimentary cover sequence of semi-pelitic, psammitic and calcareous rocks deposited in shallow water within the trough-like confines of the Limpopo belt. There is some evidence to suggest that volcanic rocks of acid to intermediate composition occur within the Messina Formation, but their distinction is subtle and at the present stage of knowledge these rocks have certainly never been mapped separately. As defined in eastern Botswana, the Messina Formation should include certain gneissic rocks and amphibolites as well, and criteria for distinguishing these gneisses from the basement gneisses include less complex deformational history, less homogenization through partial

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melting, the presence of less deformed basic dykes of post-basement age, and the frequent and conformable association of these younger gneisses with metaquartzites, magnetite quartzites and marbles. The quartzo–felspathic gneisses may or may not be banded, but where they are banded the banding is often clearly due to original stratification, as opposed to the tectonic banding which much of the refoliated and reworked basement gneisses possess. This distinction cannot always be made, but it is important to realize that it exists. Most of the amphibolites and some of the interlayered serpentinized ultrabasic horizons (Messina and Beitbridge areas) represent sill-like basic and ultrabasic intrusions within the Messina Formation, which may or may not be related to major anorthosite complexes (see below). The best developments of the Messina Formation in the central zone are exposed around Messina (Sohnge *et al.* 1948) to the east and northeast of Beitbridge (Rhodesia), and in the Baines drift area of eastern Botswana (Mason 1967).

No direct link between the Messina Formation and any of the non-metamorphic, pre-2000 Ma cratonic sedimentary sequences (Transvaal system, Witwatersrand system, etc.) has been established. Nicolaysen (1962) has suggested the possibility that the Messina Formation may represent a metamorphosed equivalent of the Transvaal system (Kaapvaal craton), a suggestion which assumes some validity, if the 2000 Ma tectono–thermal event in the mobile belt can be related to the close of the depositional phase of the Messina Formation. Furthermore, the edges of the Transvaal basin as preserved today, are certainly not the limits of the original sedimentary depositary, which could well have extended over the Limpopo mobile belt.

(*c*) *The Anorthosite complexes*

Major anorthosite complexes have been identified in the Messina area (Sohnge *et al.* 1948; Van Eeden *et al.* 1955) and in the Pikwe–Selibe area of eastern Botswana (Mason 1970; Thomas 1970). Anorthositic rocks are developed in many other areas in the northern Transvaal and in Botswana, but they do not appear to be well developed away from the Beitbridge area in Rhodesia. The anorthosite complexes appear to be semi-concordant intrusions within the Messina Formation, and are distinctly layered. Both Sohnge *et al.* (1948) and Van Eeden *et al.* (1955) recorded chromitite layers in serpentinites, hornblende gneisses and anorthositic hornblende gneisses in the area around Messina, but in attempting to show that the anorthositic rocks in this area were derived from calcareous metasedimentary rocks, Bahnemann (1970) has not considered the chromitite layers. Mason (1970) has demonstrated the existence of a layered sequence of anorthositic rocks in the Pikwe area of eastern Botswana, and this includes anorthosite, anorthositic hornblende gneiss (gabbroidal anorthosite), amphibolite and hornblendite, with every variation between these main types. Anorthositic rocks are also well developed and intimately associated with amphibolites interlayered with the Messina Formation in the Baines drift area of eastern Botswana (Mason, 1967). The anorthosite complexes are intensely folded with the Messina Formation, and compositional variations within the various rock types of these complexes have induced very different modes of deformation, with the purer anorthosites tending to resist deformation and the amphibolitic material yielding in a more mobile fashion. However, in some of the less deformed areas, distinctive relict igneous textures can be recognized, especially in the gabbroidal anorthosites. The composition of the plagioclase in the anorthositic rocks is variable but commonly within the range bytownite–labradorite. Bahnemann's explanation of variation in the anorthite content of the plagioclase as being due to contamination (Bahnemann 1970) is only partly acceptable and probably of very local application, as there must have been some original variation in the anorthite content from igneous differentiation. However, the main cause of fluctuation of anorthite content in the deformed anorthosites is probably metamorphic. (B. F. Windley, personal communication, 1972).

(d) *Cover–basement relationships*

The relationships between the basement and the Messina Formation vary from situations where the boundaries between the two are distinct enough to map out on a regional scale, to those cases where cover and basement are so complexly infolded and partially melted, that distinction between the two becomes difficult and in extreme cases impossible. The most common relationship throughout the central zone is where the remobilized basement has pushed upwards into the cover-producing major domal structures separated by tightly folded synclinal areas of cover rocks. The cores of the domal structures are thus usually occupied by basement material, and in many of these cores fragments of more resistant cover rocks (usually metaquartzite or marble) occur as relics. This type of occurrence is well illustrated by the migmatitic complexes associated with the Bulai granite west of Messina and the Mahalapye complex in eastern Botswana. On a more local scale, slices of basement may be pushed into the cover and infolded with it around the flanks of the major structures and, where intense shearing has occurred, the basement fragments may form resistant blocks around which the cover rocks become more deformed and mobile. This latter situation can be observed at outcrop scale in many parts of the central zone, and it is becoming apparent that the larger areas of basement within the zone may well have developed and behaved in exactly the same way.

(e) *Fracture systems*

The central zone is dominated by a series of major ENE trending faults and shear zones, and the only other major fracture trend developed in the zone is a complementary WNW trend. Many of these fractures are dyke-filled (basic dykes of post-Waterberg system and post-Karroo system age), and some parts of the zone are infested with dyke swarms parallel to the major fracture trends (one of the best examples occurs in the axial parts of the zone in eastern Botswana). There is no evidence from any part of the central zone of the development of the Great Dyke fracture system, which extends on both sides of the central zone into the respective cratonic areas.

Most of the major faults display evidence of transcurrent and vertical movements, and of repeated rejuvenation, with post-Karroo system vertical adjustment being the latest phase of movement in most cases. The Dowe–Tokwe fault (Sohnge *et al.* 1948) at Messina is perhaps best known of the major dislocations in the central zone because of its bearing on the location of the Messina copper deposits. The major faults are further characterized by sinuous fault traces and the development of complex splays and associated en echelon fractures.

2.3. *The southern marginal zone*

The southern marginal zone is largely confined to the northern Transvaal and is virtually a mirror image of the northern marginal zone. It consists of a granulite subzone adjacent to the Kaapvaal craton, and the Soutpansberg fault zone, which separates the granulite terrain from the central zone. Much of this zone is covered by sedimentary and volcanic rocks of the Waterberg and Karroo systems, but extensive tracts of granulite terrain occur to the north and northeast of Pietersburg and south of the Soutpansberg. Two outlying portions of the Bushveld igneous complex intrude the zone in the areas northwest of Pietersburg, and alkaline complexes occur in the areas northwest of the Sutherland Range. The alkaline complexes include syenites and nepheline syenites.

(*a*) *Granulite sub-zone*

The granulite terrain is very similar in most respects to that of the northern granulite subzone, but it is on the whole poorly exposed and there is very little known of this subzone apart from Hall's general description (1920). The reworked cratonic nature of this terrain can be clearly established in the areas north of, and adjacent to, the Sutherland Range greenstone belt, where granulite facies equivalents of the cratonic granite–greenstone material can be recognized, and in some instances can be traced laterally across the craton-mobile belt interface. A major shear belt extends along an ENE trend along the edge of the craton in the northern Transvaal, but patches of granulite material occur to the south of the shear belt (F. W. Solesbury, personal communication 1971) and in fact a transitional relationship still exists between craton and mobile belt, which is very similar to that established between the Rhodesian craton and the northern marginal zone.

A wide variety of basic and acid granulites and granulite gneisses occurs in the granulite subzone and again these are characterized by honey-brown felspars, bluish quartz and granoblastic textures. The development of orthopyroxene, clinopyroxene, or both pyroxenes in any one rock type, and the preservation of original mineralogy as opposed to retrograde mineral assemblages, depends on local circumstances and appears to be mainly controlled by compositional and tectonic factors (degree of shearing, etc.). Many of the gneissic granulites are highly garnetiferous and some are sillimanite-rich. Corundum occurs in some of the granulite gneisses as well as in pegmatites (Hall 1920). Bodies of homogeneous granitic rocks occur within the granulite zone, and although some of these are charnockitic, others appear to be post-granulite metamorphism intrusions.

A series of basic and ultrabasic bodies occur in the zone and appear to represent pre-tectonic intrusions very similar to those found in the northern marginal zone. Corundum-bearing pegmatites are commonly associated with the ultrabasic rocks, a factor which led Hall (1920) to postulate a genetic link between the two.

Burger (1959) obtained an age of 2550 ± 100 Ma from a monazite in a pegmatite cutting granulite gneisses near Bandolier Kop (Pb Th method) which is all the more significant now in the light of the 2600 Ma age obtained from the Great Dyke extensions cutting the northern marginal zone (Robertson & Van Breemen 1970).

The geological map of the Pietersburg sheet (1:250000 scale) shows fragments of Messina Formation rocks (marble, metaquartzite and magnetite quartzite) close to the edge of the mobile belt and partly overlying the Kaapvaal craton to the north of Pietersburg. The provisional 1:1000000 geological map of South Africa has consistently shown these fragments as Archaean, so that some doubt exists as to their correlation. If the fragments are Messina Formation, they will assume importance in any attempt to find a cratonic equivalent of that Formation.

(*b*) *Fracture systems*

The ENE trending fracture system is strongly developed across the southern marginal zone and extends southwards well into the Kaapvaal craton (up to 100 km at least). WNW trending

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fractures are developed but are very much subordinate to the ENE trending fractures. Major ENE trending shear zones occur immediately south of the Soutpansberg (east of Louis Trichardt) and along the contact zone between the granulite subzone and the Kaapvaal craton (north of the Sutherland Range).

A NNE, NNW trending fracture system is also developed in the southern marginal zone, which mirrors the Great Dyke fracture system developed in the northern marginal zone. An interesting feature of the adjacent cratonic area (Sutherland Range area) is the development of a very strong northeasterly trending fracture system which is invaded by swarms of basic dykes. Northwestwards from the Phalaborwa alkaline complexes, a major warp has disrupted the drainages of the lowveld areas and in fact follows a line between Phalaborwa and the two syenite complexes immediately northwest of the Sutherland Range (F. W. Solesbury, personal communication, 1971). Northwest trending faults and shears are developed through the area but are very much subordinate to the northeast trending set of fractures.

Recently released airborne magnetic maps of the southern marginal zone and adjacent craton show an overwhelming predominance of ENE structures through the southern marginal zone. The magnetic maps further indicate intense folding with axial traces parallel to the ENE trend, and suggest that the granulite terrain is sliced into extremely flattened and stretched fragments alined parallel to the edge of the mobile belt. However, further detailed study of the magnetic data is required because much of the obvious magnetic expression is a reflexion of basic dyke swarms. There is a conspicuous swing of the dyke trends in the Sutherland Range area to the northeast trend, but the ENE trend is still preserved as well.

(*c*) *Soutpansberg fault zone*

The major faults which trend ENE through the Soutpansberg region almost certainly represent reactivated Basement fractures, akin to the Tuli-Sabi fault system along the northern edges of the central zone of the Limpopo belt. Post-Waterberg system and post-Karroo system movements along these faults have been predominantly normal with downthrown blocks to the south. There is no evidence for the reverse faulting suggested by Van Eeden *et al.* (1955), Van Zyl (1950) and O. R. Van Eeden (personal communication, 1969). The fault zone marks the southern limit of the main Messina Formation developments within the central zone. Much of the zone is covered by Waterberg system and Karroo system strata, but small pre-Waterberg system inliers expose highly sheared cataclasites similar to those exposed in the Tuli-Sabi shear belt to the north. The largest of these inliers occurs adjacent to the fault system north of Vila Nora, where mylonite and flaser gneisses are reinforced by sinuous crush zones. Thus there are strong similarities between the Soutpansberg fault zone and the Tuli-Sabi shear belt, and both occur as bounding limits to the central zone of the mobile belt.

Cox *et al.* (1965) underlined the problem of antithetic faulting in the Soutpansberg fault zone, whereby the northerly dips of the Karroo system stata are offset by the southward downthrow of the faults. They pointed out that this problem also crops up in the Zambezi mobile belt (north of the Rhodesian craton) and suggested that one explanation of the problem might be relative lateral movements between the crust and its substratum. One of the factors to be considered here is that the Karroo system strata appear to have been deposited against existing fault scarps, which not only controlled the sedimentation and volcanicity, but disrupted the Karroo strata on reactivation of the faults. There is evidence that the deposition of the Waterberg system rocks in this zone were similarly controlled and affected.

2.4. *Craton–mobile belt relationships*

Any boundary drawn as a limit to the mobile belt, must of necessity be somewhat arbitrary, but the boundaries shown on the accompanying map are taken as the limits of thorough basement reconstitution as far as possible. The relationships between the mobile belt and the granite–greenstone terrains of the Rhodesian and Kaapvaal cratons are transitional, although there is evidence of intense shearing and dislocation along parts of the Kaapvaal craton contact. Shear deformation related to the major Limpopo mobilization events has affected the fringes of the Rhodesian craton for distances of up to 60 km north of the mobile belt, and within these limits swarms of basic dykes occur along ENE and WNW trends, which are the typical post-metamorphic basic dyke trends within the mobile belt itself. This shearing and dyke intrusion is particularly strongly developed in the extreme southwestern corner of the Rhodesian craton in the areas east and ENE of Francistown (Botswana). The effects of shear deformation on the Kaapvaal craton extend at least as far south of the mobile belt boundary as the so-called 'Murchison line', again some 60 km to the south of the mobile belt, with some major ENE-trending shear zones developed in this zone, and swarms of basic dykes developed along the edges of the craton. The 'Murchison line' itself has long been recognized as a fundamental crustal weakness in South Africa (Hall 1912) and marks on ENE-trending line of carbonatites and alkaline complexes in addition to being a prominent line of faulting and shearing. There is increasing evidence that igneous activity on both cratons took place during the major periods of tectono-thermal reactivation in the mobile belt at about 2700 and 2000 Ma.

Along the entire exposed lengths of the mobile belt–craton boundaries, nowhere does the Messina Formation cover sequence overstep these boundaries, except for the possible Messina Formation fragments mapped to the north of Pietersburg. Reworked basement adjoins both cratons. Any traverse from the cratons into the mobile belt will start in granite–greenstone terrain, with a prevalence of tonalitic rocks surrounding greenstone fragments in which are preserved all the relict textures of the original volcanic rocks, together with schistosity and crenulation cleavage as the dominant structural elements, and the development of mineral assemblages characteristic of the greenschist facies. Within a few kilometres (usually less than 6 km) of the mobile belt there is a marked coarsening of grain size, assemblages typical of the amphibolite facies are developed, and the fine structural elements give way to coarser foliation. The appearance of mineral assemblages typical of the granulite facies marks the beginning of the mobile belt proper, and within 1 or 2 km typical granulite terrain is usually developed, with coarse foliation, generally denser rocks, often intense shearing and the development of rod and mullion structures.

Extreme tectonic thinning of greenstone belt material into the fringes of the mobile belt occurs at several places along the northern contact (e.g. the narrow 'arms' of the Tati schist belt in eastern Botswana, and a narrow appendage of the Mweza schist belt southwest of Buhwa in Rhodesia). At these localities the grade of metamorphism changes laterally from lower amphibolite facies to extreme granulite facies in the sheared out arms of the greenstone belts over distances of 2 or 3 km. Similar relationships can be observed in the Sutherland Range greenstone belt near the southern margins of the mobile belt. The edge of the Rhodesian craton is further marked by a series of elongated bodies of sheared porphyritic granite, which are particularly well developed south of the Mweza-Buhwa schist belt in Rhodesia (Robertson 1972*b*) and south and southeast of the Tati schist belt in eastern Botswana (Mason 1970).

Although the field evidence points to transitional relationships between the mobile belt and its confining cratons, the transition from thoroughly reworked cratonic material in the mobile belt to the amphibolite facies fringes of the cratons is remarkably rapid in most areas. However, the lateral transitions along lithological units which transgress the mobile belt/craton boundaries are such that any interpretation of changes in the grade of metamorphism must ignore the usual explanations of such phenomena in terms of different crustal levels. In some way, uniformly high heat flow coupled with shearing, occurred along a prescribed strip of crust, leaving the cratonic areas intact.

2.5. *Post-metamorphic cover sequences*

(*a*) *The Waterberg system*

Following the cessation of major tectono-thermal activity in the mobile belt, volcanic and sedimentary rocks of the Waterberg system were deposited on an uneven floor of crystalline rocks in fault-bounded yoked basins. Andesitic lavas and associated tuffs were first deposited and followed by arkosic sandstones, grits and quartzites with thinly interlayered shales. In the Soutpansberg region, an upper sequence of andesitic lavas was followed by a sequence of sandstones, shales and waterlain acid tuffs (O. Barker, personal communication, 1971).

In eastern Botswana the Waterberg system includes minor volcanic developments in a dominantly arenaceous and argillaceous sequence. The Shushong series which crops out southwest and west of Mahalapye, includes calcareous rocks, and was at one time correlated with the Transvaal system (Boocock 1961). In fact this sequence rests on the crystalline rocks of the mobile belt, is not metamorphosed, and is more logically correlated with the Waterberg system with which it appears to be almost continuously linked in the areas south of Mahalapye and southwest of Palapye.

The Umkondo system, which overlies the Limpopo belt tectonites at the eastern end of the mobile belt, represents the Waterberg system as developed in Rhodesia, and the succession of arenaceous, argillaceous and calcareous rocks, with some interlayered andesitic volcanic rocks is very similar to the Waterberg system sequences developed elsewhere, in and adjacent to the Limpopo belt. The Waterberg system was deposited after cessation of major tectono–thermal activity in the Limpopo belt ($\pm$ 1900 Ma) and before the intrusion of basic dykes and sills more than 1600 Ma old.

(*b*) *The Karroo system*

The Phanerozoic Karroo system was again deposited on an uneven floor of crystalline rocks and Waterberg system strata, between 200 and 160 Ma ago, well over 1000 Ma after the deposition of the Waterberg system. In most of the Karroo system basins which overlie the mobile belt, a succession of coarse arkosic sedimentary rocks, sandstones and shales of the Ecca series, are followed upwards by reddish buff coloured sandstones and mudstones (Bushveld or cave sandstone stage of the Stormberg series). Above the cave sandstone stage, and partly interlayered with it, are thick sequences of basaltic lavas, and in the Nuanetsi Igneous Province rhyolites occur as well (Cox *et al.* 1965).

Apart from the major developments of Karroo system strata in the Kalahari basin, which covers the western end of the mobile belt, most of the depositaries within the mobile belt are fault-bounded (usually on the north side), yoked basins or clastic wedges, which were controlled at the outset and influenced by, pre-Karroo System structures. Cox *et al.* (1965) have drawn attention to the pre-existing structural controls of sedimentation and volcanicity in the Nuanetsi Province. Perhaps the best example of one of the fault-bounded depositaries is the Tuli-Sabi trough in the central parts of the mobile belt. There is considerable thickening of the sedimentary-volcanic pile in this depositary towards the Tuli-Sabi fault system, which bounds it on the north side and which was reactivated in post-Karroo system times with downthrow to the south.

2.6 *Post-metamorphic intrusive igneous activity*

Small peridotite and pyroxenite plugs intrude parts of the mobile belt, mainly the central zone, and these are post-2000 Ma but pre-Waterberg system in age. Numerous basic dykes and sills intrude the Waterberg system rocks and appear to be related to the volcanic activity associated with the deposition of the Waterberg system. The basic dykes follow the major fracture trends of the mobile belt with few exceptions.

Although as yet there are no recorded igneous episodes between the Waterberg and post-Waterberg system intrusions mentioned above and the Karroo system igneous activity, it is likely that some evidence of basic dyke intrusion, or alkaline igneous intrusion, will be found in the region as further work is carried out, but it is unlikely that any major igneous event within the mobile belt remains undiscovered.

Karroo system and post-Karroo system igneous activity is rife throughout the mobile belt, with the intrusion of dolerite dykes, sills and sheets. The dykes tend to follow the major ENE and WNW fracture trends, and many of the faults in the belt were wholly or partly filled by these dolerites. Important as it is in any synthesis of the geology of the Limpopo belt, the post-Karroo system dolerite dyke intrusions are overshadowed by the spectacular intrusion of post-Karroo System ring complexes near the eastern end of the belt (Cox *et al.* 1965). These consist of granite, granophyre, gabbro, and to a less extent nordmarkite, with nepheline syenite developed in the Marangudzi Complex. The ring complexes are clearly alined along the fundamental ENE line of the Limpopo belt.

The recent discovery of post-Karroo System kimberlite pipes near the western end of the mobile belt (covered by the Kalahari basin) prompted Crockett & Mason (1968) to equate these and the Nuanetsi ring complexes with mantle disturbance at foci of major tectonic intersections. The thermal springs of the Soutpansberg area, and sporadic seismic activity in parts of the belt, are testimony to the continuance of weak activity within the Limpopo belt at the present day.

3. Evolutionary aspects

3.1. *Sequence of events*

Table 1 is an attempt to illustrate the evolution of the Limpopo mobile belt from its initiation to the present day. The initial siting of the mobile belt must have occurred at least during, if not before, the evolution of the Archaean granite–greenstone terrain, along a predetermined crustal weakness. As yet there is no evidence that the early formation of a granulite basement in the mobile belt actually postdated the intrusion of the syntectonic tonalite plutons which surround the cratonic greenstone belts, and which were instrumental in creating the distinctive tectonic and metamorphic style of the greenstone belts. The 'reworked cratonic material' which makes up the granulite terrain could, in fact, be an extreme end product of granite–greenstone evolution related to more intense shearing and heat flow along the mobile belt as opposed to the craton. In this case the only reworked material would be pre-greenstone belt

Table 1. Suggested sequence of events in the Limpopo mobile belt

provisional time scale		
		thermal springs still active along Soutpansberg fault zone. Weak seisic mactivity
		deposition of *Cretaceous System* strata. Intrusion of kimberlites
		intrusion of ring complexes and kimberlites (?) in zones of maximum structual disturbance (Nuanetsi Province).
		extension intrusion of dolerite dykes, sills and sheets
		renewed activity along existing faults, again with dominantly vertical movement
200 Ma		deposition of *Karroo system* sediments and lavas in yoked basins on an uneven pre-Karroo System surface
		time interval of over 1000 Ma during which no record has yet been found of any significant event in the region
		reactivation of old fault lines with dominantly vertical movements, followed by intrusion of diabase sills and dykes
1700 Ma		deposition of *Waterberg system* lavas and sediments in yoked basins on an uneven floor of crystalline rocks
1900 Ma	span of active tectono–thermal life of Limpopo mobile belt	intrusion of small ultrabasic complexes. Renewed transcurrent dislocation throughout the mobile belt under brittle fracture conditions
		cessation of major tectono–thermal mobilization in the Limpopo belt
		intrustion of *Bushveld complex* on Kaapvaal craton ±1950 Ma. Renewed movement at margins of central zone as mobility decreases – formation of cataclastic zones. Refoliation accompanied by recrystallization under greenschist facies conditions in cataclastic zones
2000 Ma		widespread Rb and Sr mineral ages
2300 Ma		major tectono–thermal reactivation – Mahalapye granite-migmatite complexes, ±2240 Ma. Extrusion of Ventersdorp system lavas on Kaapvaal craton at ±2300 Ma.
2600 Ma		emplacement of the *great Dyke system* – generation of fracture system normal to ENE–WNW trending fracture system
2690 Ma		regional metamorphism under amphibolite and granulite facies conditions reaching maximum intensity in the central zone. Transcurrent movement, reactivation of basement resulting in complex refolded structures in cover, often bearing no relationship to ENE linearity of the mobile belt as a whole – typical cover/basement relationships developed
		Rb–Sr whole rock ages for Bulai and Singelele granites near Messina
		deposition of *Messina Formation* sediments and lavas over the granulite terrain, accompanied by mafic and ultramafic intrusive phases and emplacement of the anorthosite complexes. The major area of deposition appears to have been the central zone of the mobile belt, and the Shashe mobile belt
		intrusion of ultrabasic complexes? Regional metamorphism under extreme granulite facies conditions, along an ENE trending strip separating the Rhodesian and Kaapvaal cratons, widespread reworking of cratonic granite–greenstone material
		generation of ENE–WNW fracture system throughout the mobile belt, and development of related shear belts
		initiation of mobile belt
3400 Ma		development of Archaean granite–greenstone terrain. (Intrusion of ultrabasic complexes.) Inherent weakness between the present Rhodesian and Kaapvaal cratons established at the outset
		Siting of mobile belt determined by establishment of fundamental fracture systems in the early crust

N.B. The deposition of the Messina Formation may alternatively be placed after the emplacement of the Great Dyke system, and the major tectono–thermal event which caused the deformation and metamorphism of the Messina Formation would then probably extend between 2300 and 2000 Ma.

basement. However, at present it is thought that the granulite metamorphism was younger than the tectonic and metamorphic evolution of the cratonic greenstone belts, and a minimum age of 2600 ± 120 Ma is placed on the granulite metamorphism by ages determined for satellite dykes of the Great Dyke system in Rhodesia (Robertson & Van Breemen 1970).

Craton–mobile belt relationships indicate that the granulite terrain evolved at similar crustal levels to the granite–greenstone terrain of the cratonic areas, and the absence of wet cover rocks ensured a relatively dry metamorphism to produce the granulite. The major difference between the craton and the mobile belt at this stage appears to have been that the latter was the site of major shear belt generation. Whether this itself was sufficient to convert a substantial strip of crust into granulite terrain via frictional heat and extreme pressures caused by the shearing, or whether any substantial accession of heat from the mantle at the same time aided the process of granulite formation, cannot at present be determined, although the latter possibly seems likely. In any case the generation of the Tuli-Sabi shear belt and the Soutpansberg fault zone probably date back to the earliest events in the mobile belt, and probably had a strong influence on the evolution of the granulite terrain.

The Messina Formation cover sequence was laid down on the granulite basement after the consolidation of the latter, and the widespread 2000 Ma Rb–Sr mineral ages throughout the mobile belt place a minimum age on the Messina Formation. The possibility of correlating the Messina Formation with the Transvaal system has been discussed already, and the author has considered elsewhere (Mason 1970) the possibility of correlating the Messina Formation with the Pongola system (the oldest post-Archaean cover sequence in southern Africa). If the whole rock isochron of 2690 ± 60 Ma for the Bulai and Singelele granites (Van Breemen 1970) represents the basement reactivation episode which deformed the Messina Formation and produced the present cover–basement relationships within the belt, then obviously the deposition of the Messina Formation would predate this event, and in this case a correlation with the Pongola system would become a probability. The whole rock isochron of 2240 ± 420 Ma for four samples of basic and acid material from migmatites of the Mahalapye complex in Botswana (Van Breemen 1970) further indicates an older age for the Messina Formation.

Van Breemen & Dodson (1972) point out that the age of a metasedimentary rock in the Messina Formation close to the Bulai granite near Messina may be significantly older (± 3100 Ma) than the reactivation of the basement and intrusion of the Bulai granite at *ca.* 2690 Ma. This would suggest a possible correlation of the Messina Formation with the Archaean Swaziland system, but the geological evidence shows the Messina Formation as younger than the Swaziland system greenstone belts. An interpretation of the Messina Formation as a shelf sedimentary facies equivalent to the Shavaian system (Archaean) of Rhodesia cannot at this stage be disproved, but such a correlation is considered unlikely.

Another important aspect of the age of the deposition of the Messina Formation is the relationship of the anorthosite complexes to basic and ultrabasic intrusive complexes in the cratonic areas. The obvious inference if a Transvaal system age is accepted, is that the anorthosite complexes would be coeval with the Bushveld complex, whereas an older age for the Messina Formation would suggest possible correlation of the anorthosite complexes with the Great Dyke or even the Rooiwater complex of the northern Transvaal.

The 2000 ± Ma Rb–Sr mineral ages obtained from various parts of the mobile belt (Van Breemen 1970) may indicate a third major tectono-thermal event in the belt or they may represent isotopic readjustments caused by widespread shearing. Contemporaneous igneous activity (granite intrusion and the emplacement of the Bushveld complex) about this time, suggests that the 2000 Ma event was more significant than just an isotopic readjustment. Whatever the case, the shear belts which flank the central zone of the mobile belt appear to mark the edges of the main mobilized area during the deformation and metamorphism of the Messina Formation. Van Breemen (1968) suggested three explanations for the 2000 Ma event as follows: (*a*) a stage in the cooling of the belt after the high grade metamorphism, (*b*) mild re-heating of the belt around 2000 Ma, and (*c*) a stage in uplift and erosion accompanied by cooling of the belt, and he favoured explanation (*c*) at that time. If the major tectono–thermal mobilization of the Messina Formation occurred before 2600 Ma, then (*b*) would seem the more likely explanation of the 2000 Ma event, and it is suggested that the mild re-heating would have been largely due to shearing throughout the belt, for which there is plenty of evidence. Van Breeman & Dodson (1972) have recently considered the 2000 Ma event as a significant thermal event related to igneous intrusions such as the Bushveld complex, on the cratons. Obviously there is ample scope for some more detailed geochronological studies to resolve the complex history of the Limpopo belt before 2000 Ma and to establish more precisely what was the significance of the 2000 Ma event.

The stabilization of the Limpopo belt after the 2000 Ma event, in effect welded the Rhodesian and Kaapvaal cratons together in readiness for the major tectono–thermal events which affected other mobile belts in southern and central Africa between 900 and 1300 Ma and between 450 and 700 Ma. However, the mobile belt has remained a zone of instability, albeit on a diminished scale, and with the emphasis more on localized faulting and igneous intrusions, rather than tectono–thermal mobilization.

3.2. *Tectonic aspects*

Perhaps the most striking aspect of the tectonic evolution of the Limpopo belt is the zonation of the belt and its overall symmetry, with granulite zones flanking a central zone and separated from it by major shear belts. There is no doubt that the granulite zones represent on older tectono–thermal event than that which produced the central zone in its present form. Thus the mobile belt became constricted after its initiation, and finally ceased tectono–thermal mobility about 2000 Ma ago. Another tectonic aspect of the belt is the overwhelming role of shear deformation in its evolution, especially with regard to the ENE/WNW fracture systems. The separate identity of the central zone of the belt is emphasized by the absence of major fracture trends related to the Great Dyke fracture system of the flanking marginal zones and cratons, which is evidence against the Messina Formation begin pre-Great Dyke in age.

The deformation of the mobile belt appears to have been controlled by shear belt systems throughout its history. As deformation progressed in any one major event, the shear belts separated more stable crustal segments from one another, with intense shearing in between which led to partial melting of material in extreme cases and to the rheidic behaviour of this material between the stable segments. The deformation of the Messina Formation with the reactivated basement appears to have followed much the same pattern, with the cover rocks tightly infolded between and surrounding cores of basement rocks. Partial melting of the cover rocks and the basement cores in some cases, led to the production of complexly refolded structures and peculiar types of eye-folds and circular structures. Talbot (1971) recently tried to demonstrate that the latter could have formed by convective mechanisms while the rocks were rheidic, and Wynne-Edwards (1967, 1969) has suggested that the major mechanism of fold

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generation in the mobile belt environment (with specific reference to the Grenville belt in Canada) was nonuniform laminar rheid flow.

The consistency of attitudes of structural elements over large parts of the Limpopo belt (Sohnge *et al.* 1948; Van Eeden *et al.* 1955; Jacobsen 1967; Bahnemann 1971; Mason 1970; Thomas 1970; Bennett 1970*a*; Wakefield, personal communication, 1971) suggests that a compromise explanation is required for the complex fold deformation within the belt. Certainly within the central zone, the cover-basement relationships suggest an overriding shear element modified by local partial melting and flowage of rocks under rheidic conditions. Even in these circumstances the flow folding would be controlled by confining stable blocks and ultimately governed by the major shear belts. In the case of the Limpopo belt, a sequence of deformation elucidated in the granulite terrain might correspond closely to a sequence of deformation elucidated in the central zone although the sequences belong to separate events. Because of the repeated shearing along the ENE/WNW fracture trends, any deformation in the belt is likely to conform to a set pattern, so that establishment of a sequence in one area may not automatically lead to recognition of the same sequence in other areas. Deformation of the Messina Formation usually involved the generation of intrafolial isoclinal folds within major isoclinal folds, which are in turn coaxially refolded by more open folds which are themselves flexed around basement cores. The refolding appears to have been progressive and in some cases almost synchronous, with apparent maintenance of stable high grade metamorphic conditions throughout.

The shear belts or 'straight zones' which flank the central zone can be regarded as the scars of the post-granulite mobile belt 'fronts' which formed along old existing weaknesses separating the new mobile belt from the older granulite terrains, which behaved as stable blocks. It is envisaged that the central zone was mobilized separately and ground against the stabilized granulite zones to the north and south throughout this period of mobilization and even subsequent to it. There is some evidence that material from the central zone was thrust over the marginal zones along the shear belts. How far these shear belts represented sites of major vertical movements during tectono–thermal mobilization is not certain, but they are obviously important features in the evolution of the belt and would no doubt yield more pertinent information from detailed study.

3.3. *Metamorphic aspects*

The variability of shear deformation, the polyphase nature of deformation related to more than one major period of tectono–thermal reactivation, the presence or absence of cover rocks and the extreme compositional variations of rock types within the Limpopo belt, are factors which ensure the generation of complex and variable metamorphic mineral assemblages within the mobile belt. Each deformational sequence is characterized by a sequence of mineral development related to it. The dry granulite metamorphis of the marginal zones contrasts strongly with the wet metamorphism of the central zone, which varies from amphibolite facies to granulite facies depending on the presence or absence of cover rocks, the degree of shearing and partial melting, and the level at which the rocks are presently exposed. Retrogression of granulite terrain appears to have been accelerated where cover rocks have been infolded with the basement and more especially where shear deformation has been most intense. Recrystallization of mineral assemblages in rocks within the major shear belts during the final post-metamorphic period occurred under greenshcist facies conditions.

Partial melting of pre-existing basement granitic rocks has led to the formation of regenerated intrusive granites such as those developed in the Mahalapye complex. Redistribution of potash appears to have occurred particularly in zones where the cover rocks have been infolded and partially melted with the basement, as in the Messina and Beitbridge areas.

One aspect of the Shashe mobile belt which should be mentioned is the change of tectonic and metamorphis style in the Messina Formation rocks between the Matsitamma schist belt and the mobile belt proper. The change is typical of a situation where the cover rocks ovesstep a scraton/mobile belt interface, with a gradation from fine schistosity and cleavage elements and associated low grade metamorphism over the cratonic area to coarsely foliated, high grade metamorphic tectonites in the mobile belt.

Any classification of the type of polyphase metamorphism such as that encountered in the Limpopo belt, must consider the complex interplay of prograde and retrograde metamorphic sequences developed during the various major tectono–thermal events which have affected the belt. The facies series classification (Miyashiro 1961) does not take these factors into account, so that to assign a facies series to an entity like the Limpopo mobile belt would probably be misleading and, without qualification, it would be meaningless.

3.4. *Some implications*

The Limpopo mobile belt shows promise as a geotectonic entity which could yield critical evidence in the present attempts to gain an understanding of the evolution of the Archaean and early Proterozoic crust. The granulite terrains of the marginal zones in particular offer opportunities for the more detailed study of tectonic and metamorphic phenomena which will bear directly on the early history of the mobile belt and provide a sensible framework for systematic geochronological studies to be undertaken.

Anhaeusser, Mason, Viljoen & Viljoen (1969) pointed out that Precambrian mobile belts in general did not fit into models of 'orogenesis' based on the Alpine model (tectogene hypotheses), and Mason (1970) pointed out that many of the features which did not fit the Alpine model are equally difficult to explain by a model based on Phanerozoic ocean floor spreading and 'plate tectonics' (Hess 1962; Vine 1966). Dewey & Horsfield (1970) stated their belief that 'ocean-driven plate mechanisms have been responsible for the growth and evolution of continents since at least 3×10^9 yr. ago' – a bold statement to make in our present state of knowledge.

The granulite terrains and their transitional relationships with the cratonic areas, the predominantly shallow water, basin-type sedimentary rocks of the cover sequences, the uninterrupted extension of cover sequences from mobile belt onto craton, and the repeated tectono–thermal reactivation of mobile belts with or without cover sequences, are all major facets of Precambrian mobile belts which suggest that these belts were never sites where physical crustal separation occurred in the sense of present-day plate tectonics. Shackleton (1969), in a summary of some of the problems involved in understanding crustal evolution, suggested that before Phanerozoic ocean floor spreading, zones of high heat flow occurred along crustal strips arranged in polygonal patterns around the Archaean cratonic nucleii. Mantle convection was the mechanism invoked to explain the high heat flow and Shackleton further suggested that hot ascending currents rose beneath the mobile belts and that cool currents flowed nodally downwards beneath the cratons.

Clifford (1968) argued for an increase in the size of convection cells through the Earth's history, although as he pointed out this is not necessarily a progressive increase. Clifford's arguments were based on the recognition of increasing stability on the African continent with respect to tectono–thermal reactivation since the Archaean. Taking this argument together with Shackleton's, and assuming that mantle convection is the fundamental control over abnormal heat flow in the crust, it is logical to envisage a situation where increasing forces, generated by larger convective units through Precambrian time, reached a point where actual dismembering of the continental crust occurred. The Precambrian mobile belts may thus represent subcontinental 'ridges' of mantle-spreading, along which high heat flow was generated without actually separating crustal fragments. Cyclic renewal of heating and mobilization continued, with increasing stability developing in many belts until the Late Precambrian, when the convective forces approached a point where actual continental splitting became possible, and in most cases predictably achieved along the tectonic weaknesses emphasized by the 500 Ma mobilization related to the largest pre-drift cells developed. The translation of heat flow and the convective mechanisms envisaged above, in terms of shear deformation and the development of major shear belts which display transcurrent dislocation within the mobile belts, appear entirely feasible. In fact the vision of a craton surrounded by mobile belts simultaneously in the full throes of tectono–thermal reactivation makes the idea of possible craton rotation with accompanying transcurrent dislocation in the mobile belts a very real possibility. The Limpopo belt, with its flanking cratons and its development of shear belts of subcontinental dimensions, provides a superb subject for further study, and offers the possibility of a meaningful contribution to the reconciliation of geologically provable facts and attractive hypotheses with regard to the evolution of Precambrian mobile belts.

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