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THE LIMPOPO MOBILE BELT: A RESULT OF CONTINENTAL COLLISION

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Abstract. The 600-km-long Limpopo Mobile Belt is discussed within the frame of a Proterozoic supercontinent model [Piper, 1976]. Evidence is presented that the Rhodesian and Kaapvaal cratons may have been separated by distances of more than 1000 km of oceanic crust. From about 3350 Ma ago the Kaapvaal Craton appears to have been driven intermittently N, NW, and then WNW against the Rhodesian Craton forming the NE-SW trending collision zone, the Limpopo Mobile Belt, and all the major fold and fracture patterns found. This movement would be similar to the oblique movement of the Pacific plate into the Aleutian trench. When collision ceased around 2500 Ma ago, it is likely that the Great Dyke and other complexes intruded along release fractures formed at right angles to the compression.

INTRODUCTION

The 600-km-long Limpopo Mobile Belt (LMB) forms a NE trending zone 300 km wide lying between the Rhodesian and Kaapvaal Cratons and comprises some of the most ancient high-grade gneisses and granulites on earth. The LMB has been subdivided structurally into north and south marginal zones with ENE and NE trends and a central zone with predominantly N trends. The intial division of the LMB into these zones was made with boundaries represented by the Tuli-Sabi shear belt (Figure 1, C3, F2) and the Soutpansberg fault zone [Cox et al., 1965; Mason, 1973]. However, recent mapping in Zimbabwe has demonstrated that the trend of the fault lines does not define the trend of the Archaean LMB but that these faults are younger events [Light, 1980]. A complete reevaluation of the units of the LMB should therefore be seriously considered as has been suggested previously by Du Toit and Van Reenen [1977]. The behavior of the major bounding faults defined by Cox et al. and Mason is consistent with their being younger Riedel-type shears [De Sitter, 1964; Coward, 1976; Coward et al., 1976a; Key and Hutton, 1976]. These shears would form oblique to the trend of the mobile belt after the deformation had begun and thus would not represent primary continental rift-type faults.

The rocks within the LMB may be broadly subdivided into a garnet-free basement (Sand River Gneisses, R.S.A.) overlain by highly metamorphosed and complexly deformed garnet-bearing cover gneisses, the Beitbridge Group (Gumbu, Malala Drift, and Mount Dowe Groups, R.S.A.). These are intruded by the Ultramafic and Anorthosite Suites (Undifferentiated Messina Suite in R.S.A.) and the charnockitic enderbitic, Singelele, and Bulai gneisses (porphyritic granites) (Figure 1).

Burke and Dewey [1972], Dewey and Burke [1973], and Burke et al. [1976] believe that the Archaean and Proterozoic mobile belts of Africa have formed as a consequence of continental collision of the Tibetan type, while Van Biljon [1977] has suggested a similar origin for the

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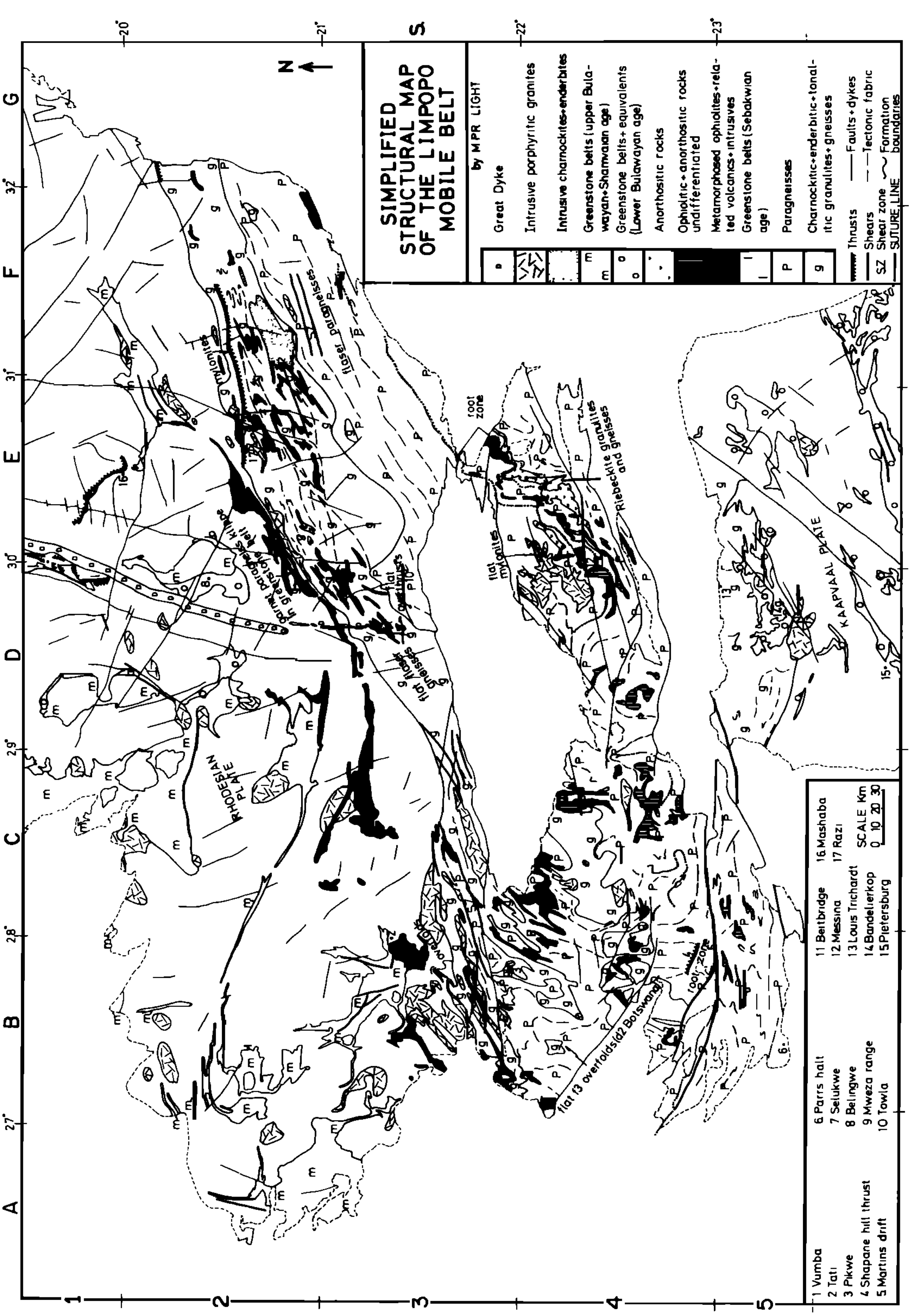


Fig. 1. Simplified structural map of the Limpopo Mobile Belt.

LMB. One of the major difficulties in interpreting Archaean terrains by using a plate tectonic concept is the accepted andesite model of continental growth and island arc accretion formulated by Taylor and White [1966]. However, Tarney and Windley [1977] argue that though crustal processes may have changed with time, the basic processes of plate tectonics should not be discounted from the Archaean simply because Archaean rocks cannot be moulded into the andesite model, as spreading is the main mechanism by which the earth's internal heat is dissipated [Langford and Morin, 1976; Burke et al., 1976; Drury, 1978; Bickle, 1978]. Evidence, which is summarized in Table 1, will be presented to support the contention that the LMB formed as the result of a major continental collision event. This has led to the construction of a generalized cross-section of the LMB based on a continental collision model (Figure 2).

EARLY ARCHAEAN

Green [1972] and Glickson [1980] suggest that the original crust of the earth formed from impact origin around 4000 Ma and has an ultrabasic to mafic volcanic composition, while Green [1972] believes that the Yilgarn Shield in Australia is a terrestrial mare.

The central zone of the LMB is underlain by a slightly younger basement complex (±3800 Ma old) which is composed largely of metamorphosed greywacke [Barton, 1979]. This sequence of dirty clastics is believed to have been derived from the erosion of an island arc or Andean type orogenic belt with the rocks forming as a result of subduction and partial melting of the basaltic crust [Barton and Key, 1981]. There is thus an indication in the LMB that plate tectonic activity began at least 1000 Ma earlier than the presently recognized starting time. This sequence of greywackes underwent folding, metamorphism and faulting [Light, 1980] and though older is compositionally similar to the Archaean gneiss terrains of both the Rhodesian and Kaapvaal Cratons [Barton and Key, 1981].

Mafic dykes of tholeiitic composition intruded into the basement complex of the LMB around 3550 Ma at about the same time as mafic magmatism in the Kaapvaal Craton (Barberton Mountain Land) and the Rhodesian Craton (Sebakwian Group) [Jahn and Shih, 1974; Barton et al., 1977a; Wilson et al., 1978; Hamilton , 1979]. These tholeiitic dykes, though they have a higher initial $^{87}Sr/^{86}Sr$ ratio and K_2O content than oceanic tholeiites, have an average composition similar to the parental magma of the Messina Suite (Anorthosite and Ultramafic Suites) which is similar to modern oceanic island arc tholeiites and to Archaean basalts from the Canadian greenstone belts [Barton et al., 1977a]. The high initial $^{87}Sr/^{86}Sr$ ratio and K_2O content of these tholeiites is very likely the result of crustal contamination of oceanic source material [Light, 1980].

Paleomagnetic data indicate that the Rhodesian and Kaapvaal Cratons could have formed a stable unit since about 2300 Ma [Piper et al., 1973; McElhinny and McWilliams, 1977], but simatic gaps may have existed prior to that time in the continental crust. McElhinny and McWilliams [1977] suggest that simatic gaps may also have transiently opened later in many parts of the supracontinental crust as the tolerance of polar wander curves does not allow for the detection of plate separations of less than 1000 km. They say, however, that the continuity of the polar wander paths requires that such gaps close through the return of the cratons to their original positions.

The formation of the basin into which the Beitbridge Group cover rocks were deposited occurred at approximately the same time as the formation of the early granite-greenstone terrains on both cratons [Hawkesworth and O'Nions, 1977; Wilson et al., 1978; Hamilton et al., 1979].

Although the high grade of metamorphism of the LMB has led to the

TABLE 1. Archaen Tectonic and Metamorphic Events in the Limpopo Mobile Belt and Adjacent Cratons

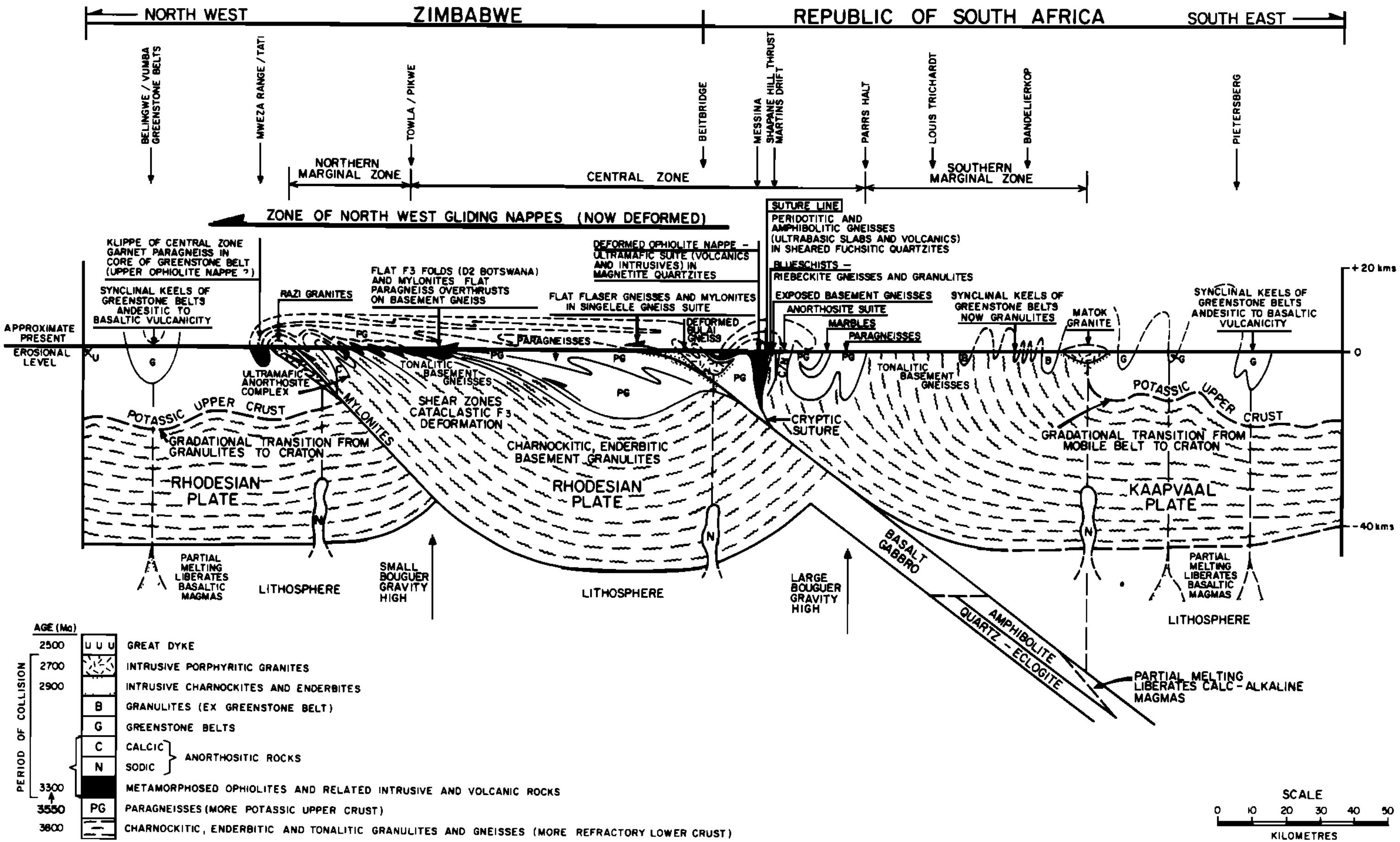
Event Number	Approx. Age Ma	Metamorphic Grade and Depth of Burial, km	Event Description
(15)	2500	Medium 19-21	WNW collisional movement of Kaapvaal Craton against Rhodesian Craton ceases. Pressure release along NNE sinistral wrenches results in intrusion of Great Dyke and Satellites.
(14)			ENE dextral shear faults and straightening zones form in the now brittle Limpopo Mobile Belt from continues WNW collisional movement of Kaapvaal Craton against Rhodesian Craton. These dextrally drag nappe and cross fold structures on a regional scale.
(13)	2600	Medium 19-21	Flat-lying cataclastic zones form in the now brittle Limpopo Mobile Belt representing the bases of collitional overthrusts. Steep ENE flaser gneisses and mylonites and breccia lines develop. A new subduction event begins on the N margin of the Rhodesian Craton. Intrusion of Chilimanzi age granites and overthrusting of the Mashaba Igneous Complex.
(12)			Cracking of the now brittle Kaapvaal and shearing of the Rhodesian cratons along NW dextral and NNE sinistral wrenches results from collisional compression due to WNW movement of Kaapvaal Craton against Rhodesian Craton.
(11)			Three sets of cross folds develop (F4, open, NW axes; F5, conical, steep NE axes; F6, open NW axes) from rotational, three-dimensional progressive deformation due to lateral spreading in the collisional zone.
(10)		High-Medium 19-21	Major F3 folds develop (isoclinal, NNE axes) which form flat overfolds in Botswana (D2). Result from a change in the collisional movement of the Kaapvaal Craton from NW to WNW against the Rhodesian Craton.
(9)	2700	High 21-29	Ophiolitic (Ultramafic Suite), and Beitbridge Group nappes overlap southern margin of Rhodesian Craton. Widespread intrusion of porphyritic granites in Rhodesian and Kaapvaal cratons as well as Limpopo Mobile Belt (Syntectonic Bulai Gneiss) due to continued crustal thickening from collision. Associated formation of volcanic Upper Bulawayan and Shamvaian greenstone belts in cratons.
(8)		High 29-36	Major F2 folds with NE axes parallel to the trend of the Limpopo Mobile Belt develop from recommencement of NW collisional movement of Kaapvaal Craton against Rhodesian Craton.
(7)	2900	High 36-38	Period of no tectonic stress. Remobilization of thickened granulitic basement and intrusion of charno-enderbites, Singelele gneiss, and volcano-sedimentary Lower Bulawayan Group greenstones due to drop in tectonic stress.

TABLE 1 (cont.)

Event Number	Approx. Age Ma	Metamorphic Grade and Depth of Burial, km	Event Description
(6)	3100	High 38	F1 recumbent cross folds form with ESE to SE axes due to lateral spreading in collisional zone. Attenuation of fold limbs to form horizontal shears which represent the base of nappes.
(5)	3300		Extrusion and intrusion of ophiolitic rocks (Ultramafic Suite) along blueschist suture zone (amphibolitic, peridotitic, and riebeckite gneisses). Ophiolite nappe begins to glide NW from suture zone. Anorthosite suite rocks intrude, derived from subducting oceanic crust inclinded SE below Kaapvaal Craton.
(4)	3350		Northward movement of Kaapvaal Craton and initiation of collision with the Rhodesian Craton. Northward oriented nappe tectonics in Sebakwian Group greenstones on Rhodesian Craton derived from orogenic belt to the south.
(3)			Deposition of the Beitbridge Group rocks in a marginal marine environment on the edges of the Rhodesian and Kaapvaal cratons. Kaapvaal and Rhodesian cratons possibly separated by more than 1000 km of oceanic crust.
(2)	3550		Formation of the Sebakwian Group and equivalent age greenstone belts on the Rhodesian and Kaapvaal cratons. Intrusion of mafic dykes into basement complex of the Limpopo Mobile Belt.
(1)	3800		Deposition of a sequence of greywackes as a result of the erosion of an island arc or Andean type orogenic belt. Later folding, metamorphism, and faulting of these rocks to form the basement complex of the Limpopo Mobile Belt.

destruction of nearly all intrusive and volcanic textures, thus erasing the evidence for an oceanic crust basement to this basin, many rocks belonging to the Ultramafic Suite and Nulli Formations (Beitbridge Group) appear to have originated from such basement. Furthermore, structural data indicate that much of the greenstone belt related lithologies in the Northern Marginal Zone and in the Lower Bulawayan Greenstone belts adjacent to the northern margin of the LMB probably represent oceanic crust and oceanic crust-derived volcanics and intrusives which have been thrust over the Rhodesian Craton from the central zone of the LMB. We can therefore speculate that a large oceanic gap more than 1000 km across existed between the Kaapvaal and Rhodesian cratons prior to 3350 Ma ago and was apparently in the process of closing.

The Beitbridge Group, which forms the sequence of cover rocks on the LMB, consists of metamorphosed sandstone, shale, carbonate rock, banded iron formation, and metamorphosed volcanics [Light et al., 1977; Light and Watkeys, 1977; Light, 1980]. These are characteristic of an Atlantic-type continental marginal marine environment. From their dis-



GENERALIZED NORTH WEST - SOUTH EAST CROSS SECTION OF THE LIMPOPO MOBILE BELT BASED ON A CONTINENTAL COLLISION MODEL

Fig. 2. Generalized northwest-southeast cross-section of the Limpopo Mobile Belt based on a continental collision model.

tribution they appear to have been largely deposited on the southern edge of the Rhodesian Craton [Light, 1980].

MIDDLE ARCHAEN

Stowe [1968] has described the earliest evidence of collisional-type horizontal tectonism in the Selukwe greenstone belt (Figure 1, D1), where Sebakwian Group greenstones were folded prior to 3350 Ma [Nisbet et al., 1981] into huge, flat-lying nappes. These appear to have moved northward from a rising orogenic belt in the south. Key et al. [1976] have suggested that this nappe may have extended hundreds of kilometers into Botswana and it is possibly related to the tectonism which produced prominent northerly trends in the gneisses between the Great Dyke (D1) and Mashaba (E1) [Moorbath et al., 1976]. This tectonism was almost coeval with the extrusion and intrusion (into the Central Zone of the LMB) of the Ultramafic and Anorthosite Suite rocks around 3300 Ma ago [Barton and Key, 1981]. These are considered to be of subduction origin, suggesting that the tectonism represents the effects of the initial contact of the northward moving Kaapvaal Craton with the Rhodesian Craton.

As previously stated Barton et al. [1977b] have determined that the parental magma of the Ultrabasic and Anorthosite suites (Messina Suite) had a tholeiitic composition similar to both modern oceanic and island arc tholeiites. A similar geochemical correlation was also demonstrated between these rocks and the calc-alkaline rocks of the circum-Pacific rise [Light, 1980] and the differentiation trends shown by them are consistent with a tholeiitic magma enriched in water (Figure 3) [Hess and Poldervaart, 1967]. This water is believed to be derived from contamination of oceanic crust by wet volcano-sedimentary rocks dragged down into a subduction zone [Light, 1980]. Barton and Key [1981] have argued that the large initial $^{87}Sr/^{86}Sr$ ratio of the magma giving rise to the Ultramafic and Anorthosite suites is consistent with its being emplaced during continental rifting or the formation of an aulacogen [Barton, 1979; Barton et al., 1979]. However, although the initial $^{87}Sr/^{86}Sr$ ratio of the Anorthosite and Ultramafic suites is large, the range of initial $^{87}Sr/^{86}Sr$ ratios of similarly aged cratonic gneisses is even greater [Light, 1980]. It is therefore conceivable that these rocks have been derived by contamination of an oceanic crustal plate by subterraneously eroded preexisting crust having been carried down into the Benioff zone, as evidenced by the finding of high $^{87}Sr/^{86}Sr$ ratios in the calc-alkaline volcanics of Java, Japan, and the central Andes [Ostmaston, 1977]. The presence of abundant quartz veining and inclusions in the Anorthosite Suite rocks east of Beitbridge (D4) and anomalous quartz-rich anorthosite rocks in the Selebi-Pikwe (B3) area in Botswana [Key, 1977] is consistent with this crustal contamination model.

The Ultramafic Suite consists of coarse-grained, cumulously layered pyroxene, amphibole, and dunitic granulites which form conformable layers and oval zoned serpentinite bodies preferentially in the magnetite quartzites (Swebebe Ferruginous Member, Beitbridge Group). All these rocks have high Ni and Cr contents, indicating a certain igneous origin and chemically follow a tholeiitic differentiation trend (Figure 3) [Light, 1980]. These conformable ultramafic layers were therefore intruded as sills or extruded lavas before or during the high-grade metamorphism and it is likely that the primary intrusive rocks were gabbroic or plagioclase-bearing ultramafic rocks [Light, 1980]. The Ultramafic Suite has the composition and volcano-intrusive nature of a highly metamorphosed ophiolite assemblage, as was first suggested by Van Biljon [1977]. The presence of large concentrations of metabasalts related to these rocks at Selebi-Pikwe (B3) [Barton and Key, 1981] further supports this idea. High-level ophiolites, which may form nappes, are considered a characteristic of colliding continental margins [Dewey and Burke,

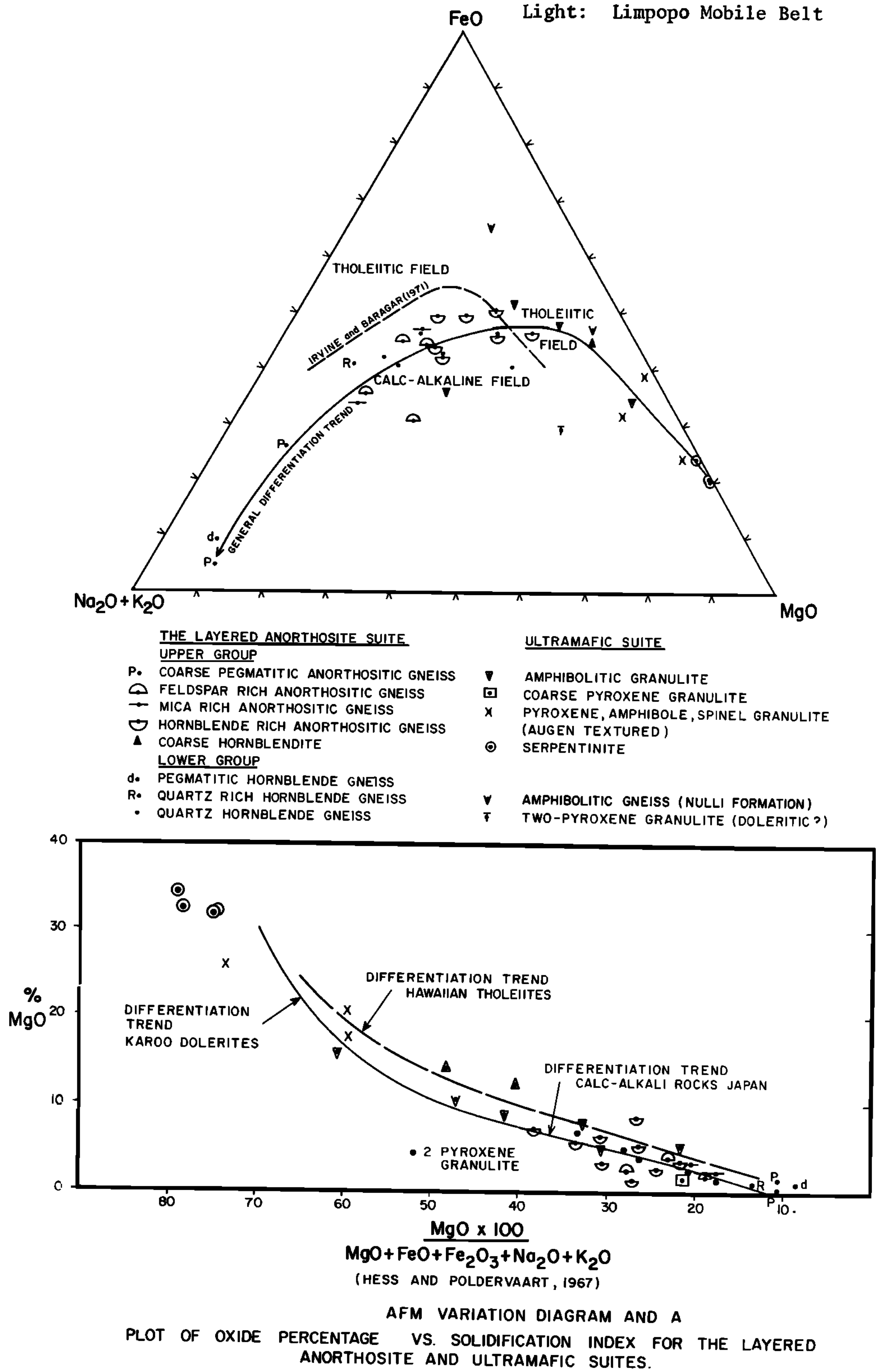


Fig. 3. AFM variation diagram and a plot of oxide percentage versus solidification index for the layered anorthosite and ultramafic suites.

1973] and such ophiolite nappes are believed to have formed at this time in the central zone of the LMB and begun to glide NW (Figure 2).

SUTURE ZONE

A major problem in relating the plate tectonic system to the origin of mobile belts is the apparent lack of a recognizable suture zone. Blueschists are preserved at a high level, but deeper down only deformed magnesio-riebeckite bearing mafic and ultramafic rocks may be present, while at the deepest level the suture may be cryptic, represented by ultramylonites, pseudotachylites, a sliver of serpentine, or a few flakes of fuchsite mica [Burke and Dewey, 1972; Dewey and Burke, 1973; Burke et al., 1976].

East of Beitbridge, there is evidence along the Nulli, Sebetwe, Bodekwa, and Lukumbwe quartzite hills (Nulli Formation, Beitbridge Group) of a suture line where strongly foliated and layered amphibolitic and peridotitic gneisses form a distinctive association with sheared or schistose to granular and massive fuchsitic quartzites. The quartzites contain laminar mica and quartz, the latter showing thin deformation lamellae. The amphibolitic and peridotitic gneisses have, respectively, segregation layers of hornblende and plagioclase and ferromagnesians with irregular streaks of serpentine after olivine. Crenulation lineations are well developed in these rocks and the olivines appear to have been stretched out during the deformation. The strongly sheared nature of the mafic and ultramfaic rocks here is quite distinct from that of the morepolygonal granulites which occur preferentially in the magnetite quartzites and suggests that the former represent a major shear line.

Rounded fragments of amphibolitic gneiss which are present in the quartzites are interpreted as volcanic bombs or the result of penecontemporaneous weathering of the interbedded mafic rocks during the deposition of the quartzites. This and other factors suggest that the amphibolitic gneiss represents water-laid tuffs or reworked volcanic ashes or lavas [Light, 1980]. The entire mafic-ultramfaic-quartzitic complex is interpreted as a deformed suture line forming the contact between the Kaapvaal and Rhodesian cratons and the layered peridotitic gneisses as metamorphosed, sheared ultrabsic slabs derived from oceanic crust. This suture has been repeated by deformation, and the zone has been provisionally traced into South Africa and Botswana and into Zimbabwe by using the distinctive association it has with the quartzite hills. At the NE end of the LMB in Zimbabwe the proposed suture is associated with the same ultramafic rock assemblage [Swift et al., 1953]. East of Beitbridge aplitic riebeckite gneisses and granulites are associated with magnetite quartzites, often where these are infolded with the Anorthosite Suite. They may represent a metamorphosed equivalent of the high level subduction zone 'blueschists' (Figure 2). Similar rocks have been found in the area west of Beitbridge (D4) by M. K. Watkeys (personal communication, 1980). Watkeys has suggested that this ancient suture may have been reactivated in Tertiary times and acted as a conduit for the intrusion of the Nuanetsi Igneous Comples (F3).

A major NE trending anorthosite belt lies on the SE side of the suture line-ophiolite complex near Beitbridge (D4) and structurally is an antiform overfolded to the SE and NW. Quartz-hornblende gneiss with calcic suite affinities (Na rich) is the major rock type, while anorthosites with calc-alkaline affinities (Ca rich) occur sporadically in thin belts to the NW. The anorthosite belt appears to have a southern limit approximately parallel to the proposed suture line and has been folded NW over the suture line-ophiolite complex in the Beitbridge area (D4). The anorthosites also appear to have been folded NW over much of the central zone in Botswana [Key, 1977], which suggests that they were derived from a root zone in the south-east, probably the Martins Drift-Shapane Hill thrust area (B4) (Figure 4). Granitic liquids and anorthositic residues [Green, 1969] may be produced at an early stage of

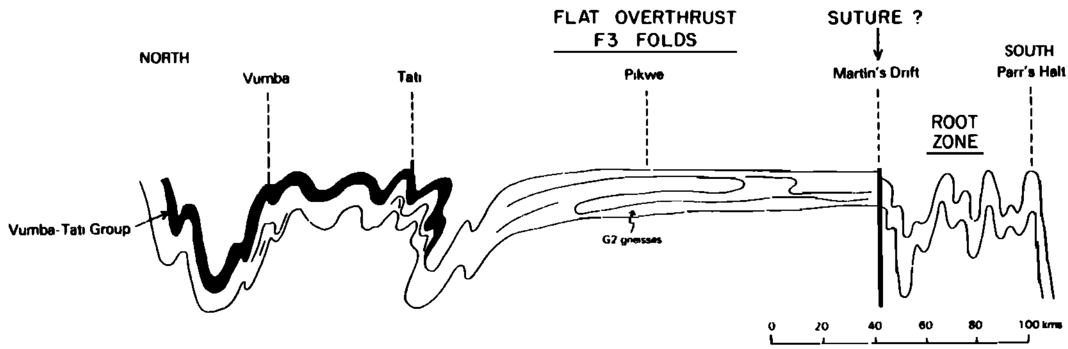


Fig. 4. Large-scale structural deformation typical of Limpopo Mobile Belt. (Modified from Key [1977].)

convergence by fractionally crystallizing calc-alkaline magma rising from the upper surface of a descending lithosphere slab. The anorthosites which are calcic at high levels and more sodic at deeper levels remain in the lower crust [Dewey and Burke, 1973]. The composition and close association of the Anorthosite Suite with the proposed suture line in the central zone of the LMB suggests that these rocks were probably formed in this way and that the subducting plate was inclined SE below the Kaapvaal Craton. This would be consistent with the NW trend of the major thrusts and nappes.

NAPPE STRUCTURES

Litherland [1973] described an early phase of recumbent folding in the northern marginal zone which formed the root zone of nappes defined by the greenstone belts of the southern part of the Rhodesian Craton. Key et al [1976] has also correlated K feldspar porphyritic gneisses in the northern marginal zone with arkosic snadstone in the basal part of the stratigraphy of greenstone belts in the S part of the Rhodesian Craton, though C. W. Stowe (personal communication, 1981) feels that this correlation may be invalid. Garnet-grunerite gneisses in the Mweza greenstone belt (D2) adjacent to the LMB on the Rhodesian Craton correlate with the gneisses of the central zone of the LMB (Swebebe Ferruginous Member, Beitbridge Group) (C. W. Stowe, personal communication, 1981). These paragneisses probably represent a klippen in this greenstone keel, and the whole structure represents the overturned nose of an upper ophiolite nappe which has been thrust north over the LMB (Figure 2) [Light, 1980].

The LMB thins from 200 km in Botswana to about 80 km in Zimbabwe (discounting the southern marginal zone), and this is associated with a change in tectonic style and dip from essentially flat nappe-like structures in the central zone in Botswana to steeply dipping folds in Zimbabwe (Figures 1 and 4). This suggests that the narrow belt in Zimbabwe represents a more deep-seated root zone for nappes, already eroded, while the shallower exposures in Botswana still have some record of the overlying thrust structures [Light, 1980]. The contact of the northern marginal zone and the Rhodesian Craton has been shown by James [1975] and Odell and Phaup [1975] to consist in places of a zone of southward dipping thrust faults and mylonites (E2) which may be equivalent to the main boundary thrust in the Himalayas. The original contact in the west has been altered by repeated deformations and was probably a major zone of northward dipping thrust faults [Key, 1977], while they dip steeply to the south in eastern Botswana [Cox et al., 1965].

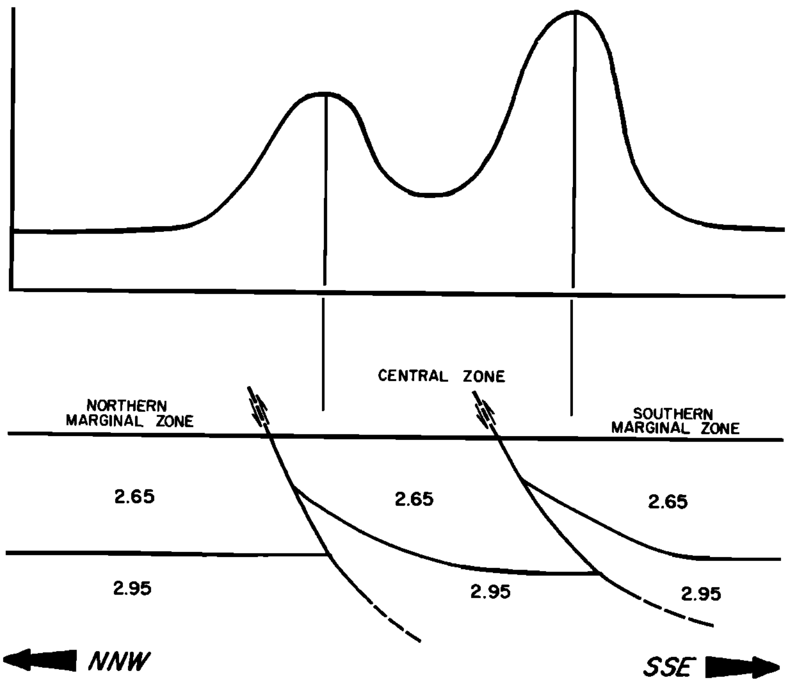


Fig. 5. Bouger gravity pattern across the Limpopo Mobile Belt [Barton and Key, 1981].

The Monachane (Botswana) and Towla (D3) paragneiss zones are considered to represent tectonic fragments (of overthrusts) of Beitbridge Group rocks resting on a granulitic and gneissic basement. Cataclastic zones are present in these paragneisses, and they have not suffered the complex deformations of the underlying basement rocks [Mason, 1973]. Elsewhere, the contact of the northern and southern marginal zones with the cratons is considered to be texturally and mineralogically gradational from greenschist to granulite, and the high-grade equivalents are believed to be reworked greenstone material [Van Reenen and Du Toit, 1977; Robertson, 1968]. This suggests that much of this high grade mafic material represents ophiolitic sills and volcanics, intercalated and overfolded to the NW and SE during the period of major collisional nappe formation in the LMB.

The Bouger gravity pattern in the eastern part of the LMB supports the contention that a portion of the central zone has been thrust over the northern marginal zone, while part of the southern marginal zone has been thrust over the central zone (Figure 5) [Barton and Key, 1981]. This pattern is also consistent with the presence of a deformed subduction zone under the central zone in which high density oceanic crustal rocks, which are closer to the surface, cause a large gravity anomaly. These high density rocks would also have been dragged up by thrusting along the mylonite zone forming the northern boundary of the LMB and thus caused the smaller gravity anomaly (Figure 2). Field evidence for this thrusting may be found in Botswana [Key, 1977].

LATER DEFORMATIONS AND INTRUSIVE EVENTS

After the beginning of nappe formation and prior to the intrusion of the charno-enderbites around 2900 Ma [Barton et al., 1979; Hickman, 1978], there was a period of isoclinal recumbent cross folding (F1) which had an ESE to SE axial trend. This cross folding is believed due to lateral spreading in the collision zone, and the axial trend suggests that direction of movement of the Kaapvaal Craton relative to the Rhodesian Craton now had a more NW trend. This fold phase produced a gneissic foliation dated at about 3100 Ma [Barton and Key, 1981] and SE plunging regional lineation [Light, 1980] interpreted as a stretching lineation parallel to the NW thrust direction similar to structures found in the Sikkim Himalayas [Subimal Sinha Roy, 1977]. Possibly as a result of the extension of the nappe structures, attenuation of the F1 cross fold limbs resulted in horizontal ductile shearing and an apparent duplication of the stratigraphic pile [Watkeys, 1976]. This fold phase is believed to correspond to the major NW-SE cross folding present in the nappe structures in the Selukwe area (D1) reported by Stowe [1968].

At the peak of this tectonism, a high-grade metamorphic event is recorded in the LMB, suggesting that these rocks were buried to depths in excess of 38 km, probably a result of thickening in the zone of collision [Light, 1980]. Similar crustal thicknesses of 60-80 km have been indicated in the northern margin of the Himalayan plateau [Cummings and Schiller, 1971] and Dewey and Burke [1973] argue that the great height of the Himalayan platform is a result of continental thickening following continuing plate convergence.

LATE ARCHAEAN

A period of no tectonic stress, which appears to represent a break in the NW convergence of the Kaapvaal Craton correlates with the 2900 Ma metamorphic event in the LMB and cratons to which no deformation can be assigned [Barton and Key, 1981; Barton and Ryan, 1977]. Garnet porphyroblasts developed in the central zone paragneisses at this time. Remobilization of the thickened granulitic basement and overlying cover rocks in the thickened crust appear to have led to the intrusion around 2900 Ma (probably as a result of a drop in tectonic stress) of charnockites in the northern marginal zone [Hickman, 1978] and also charnoenderbites and the Singelele gneiss in the Beitbridge-Messina (E4) areas [Light, 1980; Barton et al., 1979b]. The volcano-sedimentary Lower Bulaywayan Group greenstones and equivalents were formed at this time, which Hawkesworth et al. [1979] from field evidence in the Belingwe area (D2), believe relate to the 2900 Ma event.

Prior to the intrusion of the porphyritic granites there was another period of major deformation (F2) in which isoclinal, upright folds with NE trending subhorizontal axes developed parallel to the trend of the LMB. They appear to be the result of the recommencement of the NW converging movement of the Kaapvaal Craton against the Rhodesian Craton. This deformation seems to correspond to a second NE-SW cross fold phase present in the nappe structures in the Selukwe area (D1) reported by Stowe [1968]. The metamorphic grade was high in the central zone at this time, and the depth of burial is estimated from metamorphic data at 29-36 km [Light, 1980].

Anatexis of the basement gneisses, probably as a result of continued crustal thickening, occurred around 2700 Ma and led to the intrusion of the syntectonic Bulai gneiss (porphyritic granite) into the central zone of the LMB [Barton and Key, 1981]. Widespread intrusion of similar younger granites (e.g., Razi (E2) and Matok (D5) into the northern and southern marginal zones as well as into the cratons) also occurred at this time [Watkeys, 1981]. Though the syntectonic Bulai granite has an affinity with the Calc-Alkaline Suite, it has been shown to have been derived from the partial melting of basement rocks in the Beitbridge area (D4) [Light, 1980]. An identical origin has been suggested for the Razi granite (E2) in the northern marginal zone [Robertson, 1973a,b]. The general SE dip of the foliated Bulai in the Beitbridge area (D4) suggests that it is a sheet granite intruded syntectonically late in F2 along steep ductile shear zones (M. K. Watkeys, personal communication, 1982). Similarly, the foliation and lineation in the Razi granite (E2) dip and plunge 40-50 degrees SE, respectively, which Robertson [1973a,b] believes is the direction of tectonic transport and emplacement consistent with the overthrust model proposed. Associated with this granite intrusive event was the formation of the volcanic Upper Bulawayan and Shamvaian age greenstone belts on the cratons probably from the same crustal thickening process at about 2700 Ma [Wilson et al., 1978]. The metamorphic grade was high in the central zone at this time, and the depth of burial is estimated at 21-29 km from metamorphic data [Light, 1980]. It is possible that the ophiolite and Beitbridge Group nappes had overlapped the southern margin of the Rhodesian Craton by this time, since the Mweza Range Klippe (D2) lies north of the Razi Granite (E2) belt.

A major period of tight inclined cross folding (F3) with NNE trending subhorizontal axes followed after the intrusion of the porphyritic granites [Light, 1980]. Wakefield [1977] has shown that the D2 deformation in Botswana (F3 at Beitbridge) was cataclastic north of the shear lines in the Pikwe area (B3), while it remained plastic south of it. This is interpreted as a result of the differing competence of the central zone paragneisses, which folded, to the northern marginal zone basement granulites, which sheared, during the NW oriented overthrusting and nappe formation. The major F3 cross folds probably formed as a result of a change in the relative movement direction between cratons from NW to WNW. The depth of burial is estimated as between 19 and 21 km [Light, 1980; Horrocks, 1980], while the metamorphic grade retrogressed from high to medium.

Three sets of cross folds developed on the major F3 folds: F4, which were open, upright, concentric folds with moderately plunging NW axes; F5, upright conical folds with NE to ENE subvertical axes; and F6, very open folds with subhorizontal NW axes. It is believed that these cross folds formed as a result of rotational, three-dimensional progressive deformation due to lateral spreading in the zone of collision, similar to the fold sequence in the Sikkim Himalayas [Subimal Sinha Roy, 1977]. The fact that the deformation had become concentric at this time indicates that the central zone of the LMB was now more brittle, no doubt due to a large drop in temperature indicated by the retrogression in metamorphic grade.

The fact that the LMB locally becomes E-W oriented in Botswana is believed partly due to major F4 buckling but also a result of right lateral drag of the northern margin of the LMB by the westward movement of the Kaapvaal Craton.

Cracking of the now brittle Kaapvaal and shearing of the Rhodesian cratons occurred along NW dextral and NNE sinistral wrenches. This event is dated as post greenstone belt deformation (Shamvaian folding) and preceded and influenced the intrusion of the younger granites at about 2600 Ma (M. K. Watkeys, personal communication, 1980). These shear fractures are believed to result from the collision of the Kaapvaal Craton with the Rhodesian Craton and have displaced the early ENE mylonite zones. Flat-lying cataclastic zones appear to have formed now in the brittle LMB and may represent the soles of overthrusts. These are dated in the Pikwe area (B3) as post about 2600 Ma [Wakefield, 1977]. Flat mylonites and flaser gneisses are believed to have developed at the same time in the Singelele Gneiss Suite east of Beitbridge (D4) [Light, 1980], while similar flat mylonitic zones in the Pikwe (B3) area have been related by Wakefield [1977] to major recumbent F3 structures. Flat flaser gneisses in the Msane river (D3), north marginal zone [Coward et al., 1976a,b] have east-west lineations suggesting overthrust movement that way. Further support for this relative westward movement of the Kaapvaal Craton against the Rhodesian Craton as defined by Barton and Key [1981], is the fact that the N-S trending eastern boundary of the Kaapvaal Craton has been displaced about 400 km westward from the N-S boundary of the Rhodesian Craton. Steep ENE trending, dextral cataclastic zones in which mylonites and flaser gneisses developed seem to be associated with this period of dextral shearing of the LMB [Light, 1980], ENE dextral shear faults and straightening zones are a slightly later event [Wakefield, 1977]. They formed in the brittle LMB prior to the intrusion of the Great Dyke (D2) (M. K. Watkeys, personal communication, 1980). Structures in Lethakane shear zone (B3) which formed at this time indicate that though the displacement was dextral, the direction of maximum compressive stress was NW-SE during the entire development of the shear zone [Wakefield, 1977]. This is consistent with the continental collision model proposed for the LMB, and these shears appear to have dextrally dragged nappe and cross fold structures on a regional scale [Light, 1980]. Tremendous strike slip faulting in Tibet associated with continental collision had the effect of rotating pieces of continental lithosphere up to 1 million km^2 in area about vertical and inclined axes through angles of tens of degrees [Molnar and Tapponnier, 1975].

The Witwatersrand basin may represent a retroarc basin in the terminology of Dickinson and Yarborough [1976] formed during the late stages of the collision (H. R. Winter, personal communication, 1981). Nisbet et al. [1981] believe that a subduction zone may have formed around 2600 Ma on the northern margin of the Rhodesian Craton. However, the intrusion of the Chilimanzi age granites (E1) around 2600 Ma and the overthrusting of the Mashaba igneous complex (E1) is considered a thermal consequence of the thrusting of the LMB onto the Rhodesian Craton at this time [Nisbet et al., 1981].

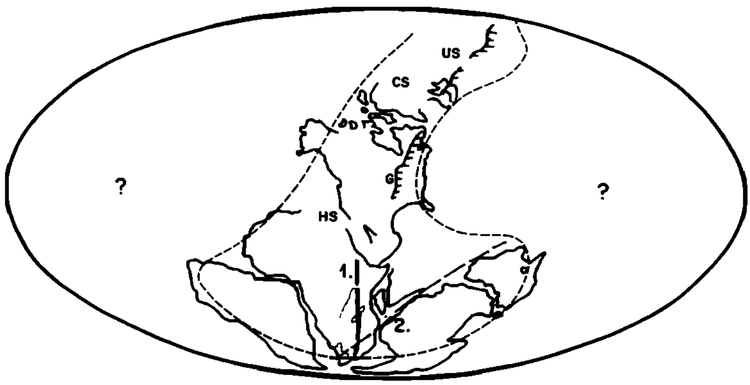
The long period of plastic and cataclastic deformation of the LMB appears to have ceased at about 2500 Ma ago (M. K. Watkeys, personal communication, 1980) corresponding to the time of intrusion of the Great Dyke (D2) into the Rhodesian Craton [Allsopp, 1965]. This is consistent with the palaeomagnetic evidence, which suggests that the Rhodesian and Kaapvaal cratons had formed a stable unit since about 2300 Ma [Piper et al., 1973; McElhinny and McWilliams, 1977]. It is believed that when the WNW collisional movement of the Kaapvaal Craton against the Rhodesian Craton ceased, pressure release along the major NNE sinistral wrenches produced conduits along which the Great Dyke and its satellites intruded. The metamorphic grade was medium at this time, and the depth of burial is estimated at 19-21 km [Light, 1980; Horrocks, 1980].

REGIONAL RELATIONSHIPS

Bouger gravity data from Pretorius [1979] suggest that the northern margin of the LMB can be traced SE from the Rhodesian Craton to the Atlantic Ocean in the vicinity of the Orange River mouth, while the southern margin appears to bend around the Kaapvaal Craton in Botswana toward Marydale in Namaqualand [Light, 1980].

The metabasaltic Marydale formation of about 3000 Ma age and the grey gneisses and old granites of the eastern Kheis domain, 2600 Ma to 2900 Ma years in age [Vajner, 1974], may represent material intruded into the SE extension of the LMB during this major collisional event. The trend of the LMB can also be extended along the bulge of Africa when plotted on a reconstruction of the Proterozoic supercontinent (Figure 6) [Piper, 1976] and up the junction of India and Antarctica and Australia suggesting that it represents a major ancient structural discontinuity which has been active since Archaean times. This trend has been previously suggested by Engel and Kelm [1972].

The NNE trend of the Great Dyke is continuous with a series of igneous centers of differing ages which lie along a great circle over a distance of 3800 km from the Orange Free State, South Africa, to Ethiopia [Cousins, 1959; Vail, 1977]. When this alignment is plotted on Piper's [1976] Proterozoic supercontinental model, the series of igneous centers represents a N trending median line of the whole continental



PROTEROZOIC SUPERCONTINENT MODEL
(PIPER, 1976)

DASHED LINES = PRECAMBRIAN CRUSTAL DISTRIBUTION LIMIT
HS = HERCYNIAN SUTURE
CS = CALEDONIAN SUTURE
US = URAL SUTURE
G = GRENVILLE SUTURE
1. = TREND OF IGNEOUS CENTRES
2. = LIMPOPO MOBILE BELT

Fig. 6. Proterozoic supercontinent model [Piper, 1976].

group and could be continued to the north along the Grenville suture in North America and into Asia. It is concluded that the series of igneous centers has intruded along a major zone of release fractures formed in the old cratonic crust following the cessation of the WNW compressional forces which had already formed the LMB. This N-S trending major discontinuity in the continental crust appears to have been an active zone of intrusion since that time.

CONCLUSION

A plate tectonic, continental collision origin for the LMB appears to be a valid hypothesis from the data presented in this paper. The Kaapvaal Craton appears to have been driven intermittently N, NW and then WNW against the Rhodesian Craton forming the NE-SW trending collision zone, the LMB. Compressional tectonic activity began before 3350 Ma, was static around 2900 Ma, was reactivated in earnest around 2700 Ma, and ceased at the time of intrusion of the Great Dyke at about 2500 Ma.

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