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The Rhodesian Archaean Craton - an Essay in Cratonic Evolution

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The Rhodesian Archaean craton – an essay in cratonic evolution

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CONTENTS

The Rhodesian Archaean craton consisting largely of granite-greenstone terrain is briefly described, and is discussed mainly in the light of points arising from recent work in the south and central parts of the country. The Sebakwian–Bulawayan–Shamvaian terminology and the evidence for more than one age of greenstone belt development are reviewed. An attempt is made to trace the evolution of the craton from the earliest recognized rocks up to the emplacement of the Great Dyke ($\pm$2500 Ma). Evidence is presented for the existence of an earlier sialic basement to the main greenstone belt cover, and to show that the main greenstone belts developed during the approximate time-span 3300 to 2900 Ma. Between approximately 2900 and 2500 Ma the main events are traced both in the craton and the Limpopo mobile belt. Proterozoic events at 2000 to 1800 Ma affecting the craton and mobile belt are briefly described and discussed.

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1. Introduction

The Rhodesian Archaean craton is a portion of the crystalline shield of southern Africa. It consists largely of a granite and gneiss terrain incorporating early Precambrian greenstone belts (figure 2). This assemblage constitutes the Rhodesian basement complex. The craton is bounded by mobile belts. To the north and northwest is the Zambezi belt; to the east is the Moçambique belt, and to the south is the Limpopo belt. All of these mobile belts are polymetamorphic with the last metamorphic events dated at between 400 and 650 Ma in the north and east (Vail & Dodson 1969). In the northwest the craton and the Zambezi belt are largely obscured by later cover, partly Proterozoic and partly Mesozoic to Recent in age.

The Limpopo mobile belt separates the Rhodesian craton from the Kaapvaal craton to the

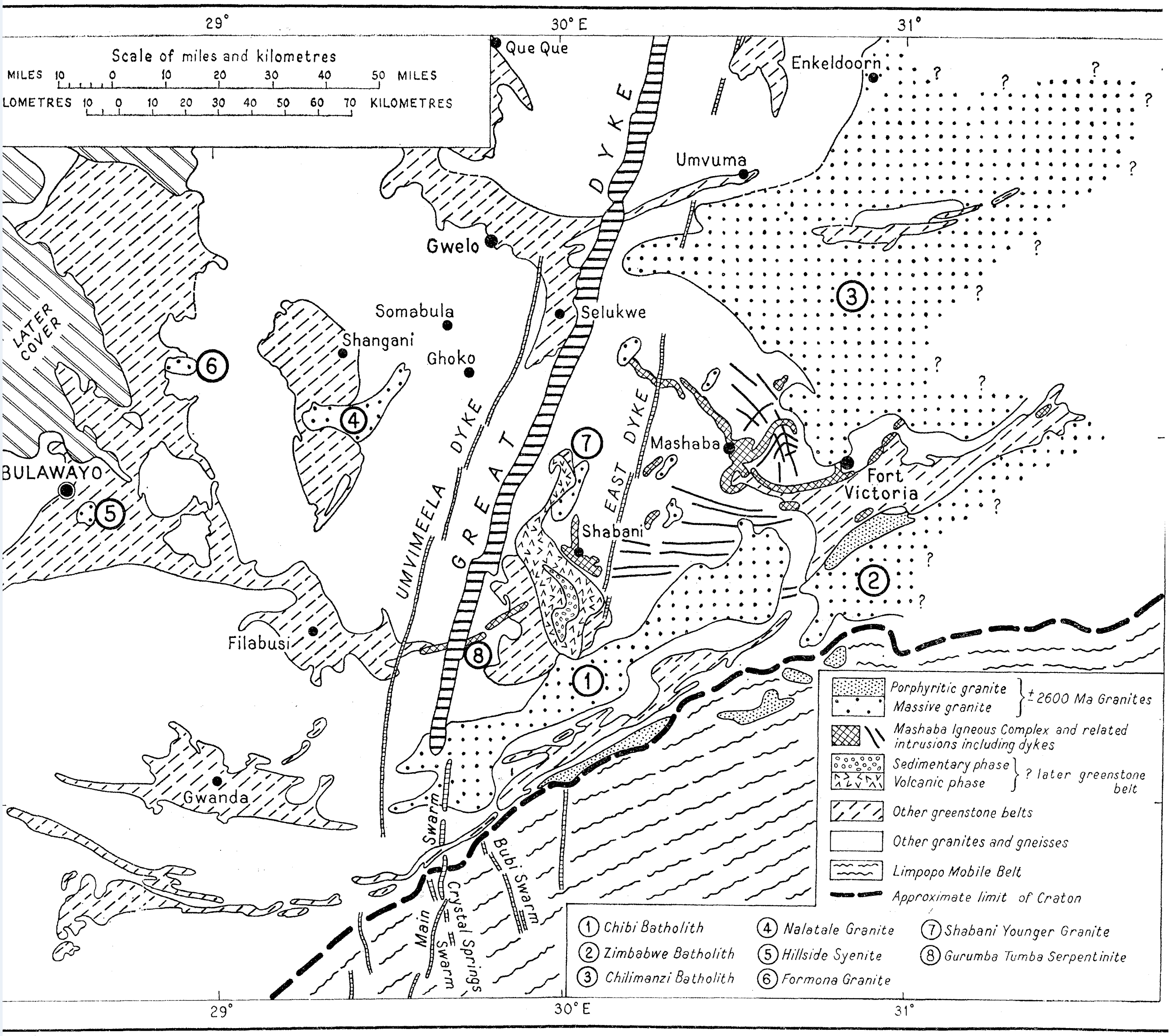

Figure 1. Main events in the Rhodesian craton during the approximate time-span 2900 to 2500 Ma.

south (Anhaeusser, Mason, Viljoen & Viljoen 1969). It can be divided broadly into a northern marginal zone, a central zone and a southern marginal zone. With the exception of the Messina formation of the central zone the rocks of the Limpopo mobile belt consist of reworked cratonic granite-greenstone terrain (Robertson 1968; Worst 1962*a*; Mason 1970). The passage southwards from the craton into the granulite facies assemblages of the northern marginal zone is gradational (Robertson 1968) and the mobile belt boundary is taken arbitrarily as the northern limit of granulite facies development (figures 1 and 2).

Cutting across the basement complex in a direction slightly east of north is the Great Dyke of Rhodesia. South of the Great Dyke proper several related satellite intrusions cut the folded high-grade metamorphic rocks of the northern marginal zone of the mobile belt. Dating of the Great Dyke and one of these satellites gives an age of approximately 2500 Ma, thus setting an upper age limit on the basement complex structures and on the deformation and metamorphism in the northern marginal zone.

The earliest attempt at a synthesis of the Rhodesian Precambrian was the classic review by Macgregor in 1951. In dealing with the basement complex he drew attention to the characteristic synclinal structure of the greenstone belts and elaborated on the idea, first suggested by Maufe (Maufe, Lightfoot & Zealley 1919), of a causal connexion between the structures in the greenstone belts and the upward rise of the intervening ovoid granitic masses. Macgregor termed these ovoids gregarious batholiths. Also, based on unconformities and similar lithologies Macgregor (1947, 1951) recognized three major systems in the greenstone belts, the Sebakwian followed by the Bulawayan and finally the Shamvaian. This tripartite subdivision is still retained by the Rhodesia Geological Survey with the modification that the lithostratigraphic term 'group' is used in place of 'system' for the major unit of subdivision.

In the past twenty years much new work has been done on the basement complex, largely in the central and southern parts of the craton, although it is only in the past decade that any serious attempt has been made to map the granites and gneisses in detail. The individual accounts of these areas are contained in various bulletins of the Rhodesia Geological Survey published since 1951.

Regional aspects of the basement complex have been dealt with in several recent publications. Viewing (1968) has reviewed the geochemistry; Bliss (1969) proposed a preliminary classification of the granites; Bliss & Stidolph (1969) reviewed the basement complex as a whole; Stowe (1971) has outlined the tectonic development, and Phaup (in press) has summarized the information available up to 1970 on the granites and gneisses and has produced a new map of the granitic rocks of the craton.

In Salisbury, in 1967, the Rhodesian branch of the Geological Society of South Africa held a symposium on the Rhodesian basement complex (Visser 1968) and in 1971 held a complementary symposium dealing largely with the granites and gneisses. The papers presented at this second symposium are to be published as Special Publication No. 3 of the Geological Society of South Africa.

This paper is not meant to be a comprehensive account of the basement complex, but is an attempt to review some of the more important points arising from the work since Macgregor's day and to summarize the main stages in the evolution of the Rhodesian craton, as the author sees them, from the earliest Archaean to the intrusion of the Great Dyke and to some of the events of the early Proterozoic. For the purposes of this paper the upper limit of the Rhodesian Archaean is taken at 2500 Ma. Discussion of the Deweras, Lomagundi and Piriwiri groups is

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beyond the scope of this paper, and the problem of the Messina formation, which is an integral part of the central zone of the Limpopo mobile belt, is ignored.

All Rb–Sr ages quoted have been calculated using the smaller decay constant (1.39).

2. The greenstone belts

(a) *The succession*

The broad succession found in the Rhodesian greenstone belts is similar to that found in greenstone belts in other parts of the world (Anhaeusser *et al.* 1969). There is an essentially volcanic phase succeeded in the larger belts by an essentially sedimentary phase. A low grade of metamorphism is a feature over the major parts of the belts.

The sedimentary phase is called the Shamvaian group (figure 2) and in several areas this contains some volcanic rocks. Andesitic to felsic types, often pyroclastic, have been recorded from areas around Sipolilo (Wiles 1965, 1968) and Hartley (Wiles 1957); basaltic and felsic types have been described from the Victoria belt (Wilson 1964). Sediments, however, markedly predominate, and in all areas several unconformable formations are usually distinguishable. Bliss & Stidolph (1969) have briefly reviewed the sedimentary associations in the greenstone belts and they see no clear-cut distinction between the sediments of the Shamvaian group and those intercalated with the underlying volcanic phase. Their work emphasizes the need for sedimentological studies to obtain a clearer understanding of the patterns of sedimentation and volcanicity during the evolution of the greenstone belt depositories.

The sedimentary phase of the greenstone belts, however, marked the cessation of major volcanism. Shallow water sediments became the norm and rapid erosion of volcanic and granitic terrains is indicated by the prominent, poorly sorted subgreywacke and arkosic suites represented. The conglomerates are characterized by the extreme polymict nature of the boulders, cobbles and pebbles. Granite pebbles are characteristic and sometimes, although no means always, predominate (Bliss & Stidolph 1969). Stowe (1971) has suggested a general progression from low-energy environment phyllitic rocks to high-energy rapidly deposited arenites derived from gneiss. This, however, is an over-simplification in view of the thick developments of phyllites in the upper parts of the sequences near Bulawayo (Amm 1940; Macgregor 1951) and Fort Victoria (Wilson 1964).

The underlying volcanic phase (figure 2) is made up largely of the Bulawayan group. The local sedimentary intercalations are usually of the banded ironstone–phyllite association of Bliss & Stidolph (1969). In the west, from Gatooma to Bulawayo, the upper part of this volcanic phase is characterized by a thick development of andesites and felsites often with thick agglomerates (Harrison 1970; Macgregor 1928; Macgregor, Ferguson & Amm 1937; Amm 1940). Cyclic volcanicity on various scales has been recognized in some of this region (Lower Gwelo – D. Edwards, personal communication). The Umniati group (Bliss 1970) west of Gatooma consisting largely of andesites and dacites with pyroclastics is in this paper considered as part of this upper sequence.

The rocks underlying this western andesite-felsite unit and making up the bulk of the volcanic phase in the larger greenstone belts away from it are mostly basalts which in some areas show pillow structures; andesites are poorly developed and felsites are largely absent. These areas include the Gwelo (Tyndale-Biscoe 1949), Shangani–Fort Rixon (Harrison 1969) and Victoria belts (Wilson 1964).

In the lowest preserved parts of the greenstone belts below this main basaltic sequence there are often the remains of an earlier unit in which ultramafic rocks, both intrusive and extrusive, are conspicuous, with mafic and sedimentary types in varying amounts. Some of the narrower belts such as Umtali–Odzi (Phaup 1937; Swift 1956), Umvuma–Felixburg (Bliss 1962), Mweza (Worst 1962*a*), Antelope and Lower Gwanda (Phaup 1932, 1933) are composed almost entirely of this assemblage which lithologically is comparable to the lower part of the Onverwacht group of the Barberton belt in the Transvaal (Anhaeusser *et al.* 1969; Viljoen & Viljoen 1969). It is recognizable also in areas such as the eastern part of the Gwanda belt (Tyndale-Biscoe 1940; R. E. P. Fripp, personal communication), the southern part of the Mwanesi belt (Worst 1962*b*) and the western end of the Victoria belt (Wilson 1968*a*).

At the western end of the Victoria belt in the Mashaba area the lowest preserved unit is a sedimentary formation which includes banded ironstones, limestone and clastic sediments. These rocks are interbedded with, and succeeded by, greenstones largely tremolite chlorite schists. Westwards, on the southern margin of the Gwelo belt, volcanic and sedimentary rocks constitute the lowest assemblage (Amm 1946; Macgregor 1937; Tyndale-Biscoe 1949). The sediments comprise banded ironstone, phyllites, micaceous quartzites and siliceous granulitic types, and in some areas predominate over the volcanics which are largely amphibolite (Macgregor 1937). Serpentinite and derived schists occur in this assemblage and in the lower parts of the overlying greenstone sequence. To the west and south a similar sequence is recognizable skirting Macgregor's Shangani batholith from east of the Lonely Mine (Macgregor 1928, 1947) along the eastern margin of the Shangani–Fort Rixon greenstone belts (Harrison 1969) to Filabusi. In these areas also the dominant types are igneous rocks, often ultramafic, with sedimentary rocks in varying amounts reaching a considerable thickness in the Shangani area (Harrison 1969). Feldspathic tuffs are recognized in the sequence east of the Lonely Mine (D. Muirhead, personal communication).

(*b*) *The Sebakwian confusion*

The simplest interpretation of this lowest assemblage and the one adopted for the purposes of this paper (figure 2) is that it forms an early but integral part of the main greenstone belt succession.

On the latest official geological map of Rhodesia, issued by the Rhodesia Geological Survey in 1971, all occurrences have been assigned to the Bulawayan group except those skirting the Shangani batholith which have been assigned largely to the Sebakwian group. The evidence for the existence of this western Sebakwian assemblage as a unit separate from the Bulawayan group is based on the recognition of an unconformity between it and the overlying, more obviously Bulawayan type volcanics northwest and southeast of Gwelo. Differences in structural elements and metamorphic grade have been described across this unconformity (Macgregor 1937; Amm 1946; Tyndale-Biscoe 1949). While there may well be an unconformity the author is not convinced that the published evidence in itself is conclusive of a break of major significance. The area requires re-investigation and it remains to be seen, as noted by Bliss & Stidolph (1969), whether these features are the products of disharmonic folding and metamorphism resulting from the intrusion of the later granites.

What is regarded by Tyndale-Biscoe (1949) and Phaup (1968) as the same unconformity is recognized in the Selukwe area between the base of the Wanderer sedimentary formation and the stratigraphically underlying eroded surface of the chromite-bearing Selukwe ultramafic complex. Pebbles of the ultramafic rocks and chromite occur in the basal conglomerate. Structurally the sequence is inverted and forms part of the imbricated lower limb of the Selukwe Nappe (Stowe 1968*a*, *b*; Cotterill 1969).

Stowe (1968*b*) describes the nappe as having slid northwards from a zone of origin now marked by the Ingezi fold belt. This is a belt of migmatites which he regards as the root zone of a greenstone belt, although the migmatites involved may largely represent something older still. It lies within the area of the Shangani batholith, parallel to and about 50 km southwest of the main Gwelo greenstone belt. Stowe has offered two interpretations of the timing of the nappe emplacement. In his earlier work (1968*a*) he, like Tyndale-Biscoe, equates the Wanderer formation with the base of the Gwelo Bulawayan group and describes the nappe tectonics as post-Bulawayan. In an alternative interpretation (1971) he assigns all of the rocks within the nappe to the Sebakwian group and suggests emplacement of the nappe before the deposition of the Gwelo Bulawayan volcanic sequence, thus invoking a major tectonic episode between this Sebakwian unit and the Gwelo Bulawayan group.

The term Sebakwian has also been used to describe supracrustal remnants in gneisses wherever it was thought that these remnants were older than the Bulawayan group volcanic rocks, such as east of Que Que (Harrison 1970), southwest of Selukwe (Stowe 1968*a*) and north and west of Mashaba (Wilson 1968*a*). Some of these remnants (see § 3) are possibly relics of even earlier greenstone belts which together with granitic rocks formed a basement to the subsequent main greenstone belt cover.

Facing this problem in the area south and west of Selukwe, Stowe (1971) and Bliss & Stidolph (1969) introduced as a provisional nomenclature the terms Sebakwian I for the supracrustal remnants of this possible earlier basement and Sebakwian II for those rocks already assigned to the Sebakwian group around the margins of the Shangani batholith, in particular those northwest and southeast of Gwelo and (Stowe 1971) to all of the rocks of the Selukwe Nappe.

3. The 'something older'

The problem of the 'something older' which may have been the floor to the main greenstone belt deposition is one fraught with difficulties since subsequent granite episodes have largely obliterated any basement-cover relationships. Indisputable evidence that part of a greenstone belt succession was laid down on an older gneissic basement comes from the Belingwe belt west of Shabani where there is clearly an unconformity (Macgregor 1951; Oldham 1970). There is the possibility, however, that this sequence is not part of the main greenstone belt development but something later still (see § 6).

Narrow remnants of 'Sebakwian' supracrustal rocks occur in gneissic and migmatitic areas in the central part of the craton (figure 2). These, although usually at a higher metamorphic grade, are similar to the lowest assemblage found in the main greenstone belts and the problem is to decide which, if any, are part of the 'something older'. Viljoen & Viljoen (1969) would regard all these supracrustals as fragments of this lowest assemblage, and the gneisses as essentially younger than the greenstone belts which they surround. Stowe (1971), however, considers several areas in the central part of the country as part of the 'something older' which he calls the central protocraton.

In the area southwest of Selukwe Stowe (1968*b*, 1971) recognizes something older than his Sebakwian assemblage. Irrespective of which interpretation of the timing of the nappe development is accepted, the folding equated by Stowe with the nappe generation in the north-westerly trending Ingezi belt refolds the northerly trending line of the Ghoko sequence of supracrustal rocks. These rocks are developed in a zone about 40 km long and about 2 km wide and form the eastern limb of a large synform the west rim of which is marked by relics in gneiss. The rocks include banded ironstones, serpentinites and tremolite schists. The Ghoko sequence Stowe (1971) has provisionally called Sebakwian I and considers it as probable basement to his Sebakwian II cover. On the basis of higher metamorphic grade and what appears to be an unconformity he recognizes relics of an even older basement to his Sebakwian I cover, and has called these oldest relics pre-Sebakwian.

Within the Selukwe Nappe itself are two sedimentary sequences, the Wanderer formation stratigraphically overlying the Selukwe ultramafic complex with marked erosional unconformity, and the Mont d'Or formation stratigraphically underlying the ultramafic complex and partly intruded by it (Stowe 1968*a*, *b*). The basal member of the Wanderer formation consists of arkoses and sericitic grits with interbedded and basal conglomerates. Clastic grains include sugary quartzite, jaspilite, chlorite schist, ultramafic rocks, chromite, pegmatite, granophyre and gneissic granodiorite. Well rounded boulders of gneissic granodiorite up to 50 cm in diameter are recorded. The rocks of the Mont d'Or formation are largely rapidly deposited grits with a chlorite-sericite or sericite matrix. Feldspar grains are now mostly altered to sericite but some of oligoclase and perthite are recognizable as well as grains of detrital chromite and a few rounded zircons.

Northwards, in the Rhodesdale batholith, the possible extension of the northerly trending Ghoko sequence occurs in the north-northwesterly quartzite and ultramafic remnants seen in the gneisses east of Que Que as described by Harrison (1970). To the north of Que Que, in the Sebakwe River, a conglomerate is exposed near the base of the preserved greenstone belt succession. This conglomerate contains boulders of pegmatite, granite and gneiss. Some of the pegmatites and granites are muscovite-bearing. Harrison (1970) also describes pebbles of quartzite similar to those quartzites occurring as remnants in the gneisses to the east.

Macgregor (1932) considered this conglomerate to mark the base of his Bulawayan System and to be resting unconformably on his Magnesian series to the east, later to be included as part of his Sebakwian System (Macgregor 1947, 1951). Harrison (1968, 1970) has shown that this is not a basal conglomerate but a locally developed intraformational bed low in the preserved lava sequence. The main ultramafic mass to the east, which he has renamed the Que Que ultramafic complex, he regards as a later intrusion. All that Harrison now assigns to the 'Sebakwian' group are the supracrustal remnants in the gneisses to the east.

Farther east still, north of Umvuma, Bliss (1962) has mapped 'pre-Bulawayan' rocks occurring as remnants in a northerly trending zone between the Umvuma and Mwanezi greenstone belts. The rocks comprise garnetiferous amphibolites, quartz-grunerite ironstone, quartz-diopside rocks and quartz-mica schists – a largely sedimentary sequence – all occurring in and intruded by a gneissic granodiorite which forms part of the Rhodesdale batholith. He concludes that all the evidence indicates that these rocks, including the gneissic granodiorite, are older than and formed a basement to the rocks of the Mwanesi and Umvuma greenstone belts.

To the south, in the country north and west of Mashaba are abundant northerly trending 'Sebakwian' supracrustal remnants in gneisses and migmatites (Wilson 1968*a*). They include quartz-magnetite ironstones, diopside quartzites, amphibolites, garnetiferous biotite schists and ultramafic rocks, some of which contain chromite. Later granites and the intrusion of the

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Mashaba igneous complex have obscured relationships between these rocks and the sedimentary formation, near Mashaba, which forms the lowest part of the preserved succession in the main greenstone belt, but again the impression is that the remnants and at least some of the gneisses are part of the 'something older'. The sedimentary formation contains some clastic sediments derived in part from a granitic source. Arkoses contain grains both of sodic plagioclase and potash feldspar, and some of the conglomerates also contain small clasts of microcline-bearing granite. Detrital chromite has also been recognized in some of the sediments (P. Cotterill, personal communication).

The field evidence, therefore, from several areas, clearly shows that granite and gneiss were exposed on surface and were being eroded at the time of deposition of sedimentary rocks which now occur low in the preserved succession of the main greenstone belts in the central part of the craton, and indicates an earlier sialic basement to at least some of this cover. It also shows that this early granitic terrain was not entirely sodic and also that it contained supracrustal rocks similar to those now preserved in the lowest assemblages of the main belts. It strongly suggests that some of this terrain with these supracrustals infolded on an early northerly trend is preserved in the central part of the craton (figure 2) as proposed by Stowe (1971).

4. The geochronology

It is convenient now to consider the period in time represented by the main greenstone belts.

(a) *Potassium–argon ages*

The lower limit of the main greenstone belts is still based on the scanty data from the Sebakwe River conglomerate north of Que Que and the nearby Piper Moss Mine (Vail & Dodson 1969; Harrison 1970). Muscovite from a granite boulder in the conglomerate gave an apparent age of 3295 ± 100 Ma; muscovite from a pegmatite from the Piper Moss Mine gave 3300 ± 99 Ma and 3440 ± 103 Ma on two determinations, averaging 3370 Ma. These determinations were carried out by Kulp at Lamont.

The setting of the Sebakwe River conglomerate has already been described (§ 3). Bliss & Stidolph (1969) describe the Piper Moss date of 3370 Ma from 'muscovite in a pegmatite vein cutting Sebakwian I inclusions in the Rhodesdale Gneiss'. From the available data in the technical files of the Rhodesia Geological Survey, this is incorrect (A. E. Phaup, personal communication). The Piper Moss Mine lies in the Que Que gneiss, a unit of the Rhodesdale batholith as described by Harrison (1970). In the mine workings this gneiss is cut by pegmatites and by mafic dykes. These mafic dykes also apparently cut the pegmatites. The mineralization is later than all these events. Within the mine, pegmatites with mica crystals up to 100 mm across are known from one area only and it was material similar to this but taken from the mine dumps which was used in the determination. At that time (1959) underground access was impossible and still is.

These data, however, indicate a granitic event around 3300 Ma which *tentatively* can be equated with the intrusion of the Que Que gneiss to give a minimum lower limit for the deposition of the main greenstone belt sequence to the west and a maximum upper limit for the earlier basement.

Wilson & Harrison (in press) recognize two later granitic events with apparent ages of about 2900 and 2600 Ma. Their K–Ar determinations were carried out by Fitch & Miller of Cambridge on samples from granites and gneisses in the south and central parts of the craton from Bulawayo eastwards to Fort Victoria. In the Bulawayo area in particular the diapiric-like Heany and Essexvale tonalites, dated at 2962 and 2848 Ma respectively, clearly intrude rocks of the volcanic phase of the greenstone belt including andesites of the upper part of the sequence. They also deform these volcanics and a thick sedimentary phase which unconformably overlies them. These sediments are assigned to the Bulawayan group on the latest official geological map of Rhodesia issued by the Rhodesia Geological Survey in 1971. Macgregor (1947, 1951) assigned them to his Shamvaian system. From his and Amm's descriptions (Amm 1940) these sediments, in particular the lower sequence with its thick arkoses and conglomerate beds containing granite pebbles, are typical of the sedimentary phase marking the end of major volcanism in the greenstone belts. They have been regarded as such in this paper (figure 2).

These data therefore give a maximum upper limit of about 2900 Ma to the deformation of the sedimentary phase in the Bulawayo greenstone belt. Wilson & Harrison (in press) recognize the 2900 and 2600 Ma granite events in the sampled area between Bulawayo and Fort Victoria and northwards to south of Selukwe. In the Rhodesdale batholith east of Que Que Harrison (1970) records several other K–Ar results which suggest the presence of both these events in this region also. The 2600 Ma granite event is something distinct from the 2900 Ma (see § 7).

The K–Ar results from the granites around Bulawayo now mean that the Bulawayan group of this area, in particular the primitive life forms represented by the well-preserved algal stromatolites of the Huntsman limestone quarry about 60 km north-northeast of Bulawayo (Macgregor 1941), must be at least 3000 Ma old.

(*b*) *Model lead ages*

Vail & Dodson (1969), in their review of Rhodesian geochronology, list 60 lead isotope ages from the Rhodesian craton, done by various laboratories at various times. Of these 37 give model ages greater than 2600 Ma of which eight give ages in the range 3000 to 3200 Ma. All but one of these eight lie in a narrow strip, less than 100 km wide, straddling the Great Dyke south of Que Que.

The recent work of Robertson (1969) is much more significant in that he presents results of new isotopic analysis of about 80 Rhodesian galena samples. These were collected from mines occurring in most of the main greenstone belts and adjacent granites and gneisses. The areas covered include Gatooma, Que Que, Bulawayo, Filabusi, Gwanda, Belingwe, Mashaba and Umtali. Robertson recognizes two groups of leads, the Que Que type and Bulawayo type, which have had different isotopic histories. His simplest interpretation of the pattern is that both types were emplaced by a major mineralizing event at about 2750 to 2900 Ma. The figure 2900 Ma occurs frequently in his descriptions of each area. The Que Que leads however show evidence of an earlier event at 3340 to 3500 Ma. These two events equate well with the 2900 and 3300 Ma minimum K–Ar ages dating the granitic episodes.

Robertson's two types of lead have also different geological distributions. His Que Que leads occur in gneissic and migmatite areas largely in the central part of the craton and in rocks of the lowest assemblage of the main greenstone belts, that is, broadly speaking, in areas where relics of the earlier basement are indicated from field evidence, and in the lowest cover sequences. His oldest model lead ages of 3200 Ma are from the Texas and Cambrian mines which were both workings on quartz veins cutting migmatites and gneisses in the older basement a few kilometres west of Mashaba. His Bulawayan type leads occur higher in the main volcanic phase of the greenstone belts in the more typical 'Bulawayan' rocks, and in granite stocks close to or cutting these rocks, i.e. spatially farther from the older basement.

The geochronological evidence therefore also indicates the existence of the 'something older' and allows the placing of the main greenstone belt deposition and deformation between the 3300 and 2900 Ma granitic events. Both these must be regarded as minimum ages, the 3300 Ma being a maximum upper limit for the earlier basement.

5. Discussion – evolution of the craton

(a) *Time: ? to* 3300 Ma

Little can be said about the evolution of the earlier pre-3300 Ma basement without further detailed mapping and more application of sophisticated dating techniques. The apparent northerly trend seen in the supracrustal relics and the gneisses in the central part of the craton approximates to the trend of the later cover sequence from Bulawayo to Sipolilo. The preserved supracrustals are meta-sedimentary and meta-igneous types largely similar to those of the lowest preserved assemblages of the later main greenstone belts, suggesting they are remnants of similar earlier structures eroded to a much deeper level. Ferruginous, mafic and ultramafic types predominate – the 'resister' rocks in any granitization process. Whether clastic sediments were a prominent part of this early basement is not known. None is preserved but some may be represented in the gneisses. The granitic phenoclasts and feldspar grains found in the lowest sediments of the main greenstone belt cover indicate that the granites of this early phase were not entirely sodic but included microcline-bearing types. Stowe's detailed work (1968*b*) south and west of Selukwe shows that the story is not a simple one but little is to be gained from over-speculation at this stage.

(b) *Time:* 3300 *to* 2900 Ma

This approximate time-span covers the initiation and infilling of the main greenstone belt depositories, their deformation and granite intrusion, and the development of the main greenstone belt pattern, a pattern subsequently modified, particularly in the east and south of the country, by the 2600 Ma granites. Although complex deformation resulted from the 2900 Ma granites the synclinal structure and early deformation still seems best regarded as gravity induced, as pointed out long ago by Maufe & Macgregor and amplified greatly in the Barberton model (Anhaeusser *et al.* 1969). Nappe tectonics as described by Stowe (1968*a, b*) in the Selukwe area are an unusual feature not only of Rhodesian belts but of the world's greenstone belts in general. Sutton (1971) has suggested that this may not be a unique occurrence. Pretorius (in Stowe 1968*b*, discussion) has called for a re-examination of greenstone belts with this point in mind and has drawn attention to the great predominance of synclines over anticlines in the Barberton belt of the Transvaal suggesting that some kind of nappe tectonics might be a possible explanation.

(i) *The greenstone belt depositories*

Stowe (1971) has summarized his ideas on the development of the Rhodesian craton. His primary phase is the formation of the early basement and he takes this to constitute his central protocraton, consisting of the Rhodesdale, Shangani and Chilimanzi nuclei. These correspond approximately to the Shangani and Rhodesdale batholiths and some of the country to the east. His next phase 'was marked by the development of small, Bulawayan eugeosynclines on either side of, and across the central protocraton'. He envisages 'the growth of the protocratonic centre by eugeosynclines and orogenesis around its margins'. Implicit in his ideas is the concept of crustal growth largely by lateral accretion.

Others among recent authors (Anhaeusser *et al.* 1969; Windley & Bridgwater 1971) have tended to think in terms of mantle-tapping fractures in an early unstable, and probably thinner, sialic crust as the factors controlling greenstone belt depositories. Windley & Bridgwater suggest that the outcrops now seen give a reasonable impression of the size of the narrow elongate basins. To the writer it seems that not only can the distribution of Rhodesian greenstone belts be explained by such fractures, whatever might have caused them, but it indicates a definite fracture pattern controlling the depositories. Three trends are evident in the present configuration of the main greenstone belts, north-northeast, northwest and east-northeast (figure 2). One of these directions approximates to the tend of the supracrustal rocks of the earlier basement, and the three thereafter form an 'eternal triangle' in Rhodesian geology in that one or other of these directions is evident throughout all of the subsequent geological history of the country.

The north-northeast trend is seen in the largest of the greenstone belts from Bulawayo to Sipolilo. Although partly masked by a cover of Proterozoic and younger rocks, this belt has a length of over 500 km and widths of up to 50 km near the Lonely Mine and perhaps 80 km southwest of Gatooma. The same trend is seen again in the smaller Mwanesi belt south of Salisbury. The northwest trend is evident in the Gwelo belt and from Filabusi to Bulawayo. The east-northeast trend is dominant in the Umvuma–Felixburg, Umtali–Odzi, Victoria and Mweza belts over a distance of about 450 km, and is seen again to the south in the Kaapvaal craton in the Murchison and Barberton belts. The northerly trending Shangani–Fort Rixon belt is best regarded as part of the sequence to the west, separated from it by later granitic intrusion. The picture is more complex in the northeast of the country but the Salisbury belt has suffered later deformation by the Chinamora batholith, whereas the belts in the extreme northeast have been modified by the later Zambezi episodes of metamorphism and tectonism.

The writer therefore prefers to think of the early basement as a much larger crustal fragment than Stowe's central protocraton with the greenstone belt depositories developing on it, not around it, on sites controlled by this suggested fracture pattern. What was later to become the Limpopo mobile belt is also thought to have been the site of such depositories at this stage, probably with a dominant east-northeast trend. Apart from the Messina formation of the central zone, the rocks of the Limpopo mobile belt consist of reworked granite–greenstone terrain (Worst 1962*a*; Robertson 1968; Mason 1970).

The infilling of the depositories began with sedimentary and volcanic rocks of the early ultramafic assemblage. The granite debris in these early sediments, although limited, indicates that there was not a complete volcanic-rock cover between the depositories at this stage. With further development of the volcanic phase it is possible that a thin such cover did exist later. Intercalated sediments occur at any horizon in the sequence of the volcanic phase. Some are chemically induced but others indicate limited erosion and uplift of the volcanic pile. No granite debris is recorded from these sediments. Although the broad chain of events in the greenstone belts is similar, there is no reason to suppose that everywhere their evolution was strictly contemporaneous. Downsagging and early deformation would be dependent on the volume and local thickness of eruptives. Unconformities are to be expected and do not necessarily indicate regional tectonic events. Subsequent general uplift is indicated, however, towards the close of the volcanic phase resulting in the deposition of the volcanic and granitic debris of the Shamvaian group sediments. These sediments were probably deposited in isolated basins, although again it does not follow that everywhere this uplift started at the same time. But, apart from the problem of the timing of the Selukwe nappe, there is so far nothing to indicate that depositories on any one trend evolved to any great extent before those on any other trend or, that one direction of the postulated fractures developed before any other.

(ii) *The* 2900 Ma *granites*

Accepting the synclinal structure of the greenstone belts as largely a gravity response to extruded heavier mantle material sinking into a less dense sialic substratum, then the interface between basement and cover becomes important in the control of any granitic magmas generated from the partial melting of this downsagged sialic crust (Windley & Bridgwater 1971). Water from the early sediments could influence the temperature of development and the rise of such magmas but mobility sufficient to produce at least an intrusive gneissic migmatitic zone on this interface could be expected. Farther from the interface the conditions would be more favourable for the development of hotter, drier magmas which could rise and further deform the cover. The diapiric-like tonalites of the Heany and Essexvale masses fall into this latter category. The arcuate intrusive margins of the large ovoids such as the Shangani and Rhodesdale batholiths become explicable as remobilized and modified basement-cover interface zones.

The minimum age of 2900 Ma has been taken conveniently as the date for the granitic event marking the upper limit of the main greenstone belt development. Even accepting this as a true age, granite magmas must have been forming over a considerable time. The felsitic lavas and pyroclasts of some of the Shamvaian group sequences can be regarded as surface expressions of granitic plutonism. The thick development of andesites and felsites in the upper sequence of the main volcanic phase in the west could indicate sialic contamination of basaltic magma together with some even earlier crustal melting. This major andesite belt is confined to the north-western part of the craton and it could be argued that it indicates the proximity of a continental margin. It may, however, merely be a fortuitous erosion feature due in part to later uplift by the 2600 Ma granites in the eastern half of the craton thus exposing only the lowest assemblages in most of these belts. Another possibility is that this western belt represents one in which greater volumes of basic magma were erupted with consequent greater and faster downsagging, and more crustal contamination, if andesitic magmas are indeed generated in this way.

The distinct granites so far recognized in this 2900 Ma group, as opposed to the approximate 2900 Ma overprinting in the earlier gneisses, are largely tonalites but include some porphyritic granodiorites and adamellites (Wilson & Harrison, in press). Some of these tonalites such as Somabula (Stowe 1968*a*) and Mashaba (Wilson 1968*a*) masses show largely gradational margins with the surrounding gneisses. Others such as the Heany and Essexvale bodies are clearly intrusive, as are the more potash-rich types of this event. More field data and age determinations, however, are required before conclusions can be reached as to any possible pattern to the sequence of intrusions of the various granitic types. Whether all of the granites of this 2900 Ma group are to be explained by partial melting of downsagged crust or whether addition to and progressive thickening of the sialic crust is also to be invoked at this stage is not clear. What is apparent, however, is that after this 2900 Ma event the craton, at least in the south, became stabilized and sufficiently rigid to allow brittle fracturing on a major scale.

6. The Mashaba igneous complex

No reliable age determinations are available, but the Mashaba igneous complex represents a period of mantle activity with intrusion and brittle fracturing in the crust between the 2900 and 2600 Ma granite events of Wilson & Harrison (in press). Its nearest time equivalent in the Kaapvaal craton is perhaps the Usushwana complex of Swaziland dated at 2874 ± 30 Ma (Davies, Allsopp, Erlank & Manton 1970).

The Mashaba complex occurs at the western end of the Victoria greenstone belt (figure 1). In its present plane of erosion it is largely a layered, predominantly ultramafic intrusion (Wilson 1968*a*, *b*). The sill-like portion contains macro-units of olivine and orthopyroxene cumulates similar to the lower portions of the Great Dyke; one unit shows well-developed, fine-scale rhythmic layering of chromite and serpentinized olivine. There is a deficiency of gabbro. Ultramafic rocks are preserved for a distance of over 50 km in contact with roof rocks (gneisses), largely in an arm of the complex extending northwest from Mashaba as a northeasterly dipping sheet. Part of the complex is an ultramafic dyke which, together with mafic dykes, forms a ring pattern cutting, and roughly centred on, the earlier Mushandike granite (Wilson 1968*a*). A modified radial pattern of mafic dykes is also present, part of it extending westwards towards Shabani and southwards (I. D. M. Robertson, personal communication) towards the later Chibi batholith. The whole complex is best regarded as an 'open system' and the remains of a gigantic magma chamber under a major volcanic centre.

The writer would equate several other ultramafic, or largely ultramafic, intrusions in the southern part of the craton with this Mashaba event. These include (figure 1) the several sill-like bodies cutting the gneisses between Mashaba and Shabani (Wilson 1968*a*, *b*) including the large differentiated body of the Shabani intrusion (Keep 1929; Laubscher 1968). Farther west still is the Gurumba Tumba serpentinite which clearly cuts at a high angle the folded rocks of the lower part of the Belingwe greenstone belt on the west, and extends to the Great Dyke and possibly beyond (Worst 1956; Oldham 1970). Other likely correlatives are the small serpentinite intrusions in the Victoria Shamvaian group on the south side of the greenstone belt (Wilson 1968*a*, *b*) and possibly the several serpentinite bodies at the east end of the Victoria belt (Wilson 1964) and the serpentinite mass south of Umtali (Phaup 1937). Several other ultramafic intrusions have been described as 'post-Bulawayan' in recent years (Bliss & Stidolph 1969). Later work may show that some of these are also assignable to the Mashaba complex event but for the moment they are best regarded as part of the main greenstone belt evolution.

The possibility of the deposition of a greenstone belt succession, with its early phase contemporaneous with the Mashaba complex event, arises from the work of Oldham (1968, 1970) in the Belingwe area. In the Belingwe greenstone belt, occupying a north-trending central syncline, is an upper sequence consisting of a volcanic unit, with intercalated sediments, capped unconformably by a sedimentary unit. Oldham estimates a total thickness of about 1.5 km in the volcanic unit in the north, increasing to about 6 km in the south. The base of this lower unit is a sedimentary formation with conglomerates, in part basal and in part intraformational. These sediments rest unconformably on the underlying rocks of the greenstone belt and overstep to rest with marked unconformity on the gneisses on the east side of the belt. West of Shabani this unconformity can be traced for over 20 km in contact with the gneisses with little variation in thickness of the sedimentary formation. Some of the conglomerates contain granitic pebbles. This unconformity is also described by Macgregor (1951) from the earlier work of Keep (1929).

In the few exposures of the contact in this sector there is no doubt of an unconformity and a preserved relationship between basement and cover. The problem is what is the cover and what is the basement.

It is possible that this is an upper sequence of the main greenstone belt development fortuitously preserved on the earlier basement. It is also possible that it represents a later greenstone-belt type of deposition on a 2900 Ma basement. The rocks are intruded north of Shabani by a younger granite; to the south the whole belt is truncated by the pre-Great Dyke Chibi Batholith. Both these granites are assigned to the 2600 Ma granite event by Wilson & Harrison (in press).

Oldham's interpretation of the area favours the second possibility. He argues (1970) for the intrusion of the Shabani ultramafic sill during the early stages of deposition of the volcanic unit. The deposition of the later sequence as a whole he records (1968) as post-dating the granite intrusion and folding on an east-northeasterly trend of the rocks below the unconformity. This folding he equates (1970) with the major east-northeasterly plunging synform mapped by Worst (1956) in the south part of the Belingwe belt. This interpretation from Worst's map is not conclusive since, from the map data alone, the structure at the south end of the Belingwe belt is explicable also as an interference pattern with the east-northeasterly trend as a late event. Farther east, however, in the Victoria belt Wilson (1968*a*, *b*) times the intrusion of the Mashaba complex as after the main phase of folding which imparted the east-northeasterly trend to the Victoria greenstone belt including the Victoria Shamvaian group, although this deformation was probably further accentuated by the subsequent emplacement of the 2600 Ma granites to the south (see § 7).

It is apparent, therefore, that the answer will come only from detailed structural work in this southern part of the craton and in the adjacent northern marginal zone of the Limpopo mobile belt. But for the moment there is the intriguing possibility of a greenstone belt cover sequence later than the development of the main greenstone belts and equating roughly in time with the plutonism of the Mashaba igneous complex and related intrusions.

7. The 2600 Ma granites

The recent K–Ar dating programme of Wilson & Harrison does not in itself clearly establish this granitic event at 2600 Ma but their results taken in conjunction with field data and other age determinations show evidence of major granitic intrusion into the craton about this time and pre-dating the Great Dyke (Wilson & Harrison, in press).

The granites assigned to this event are often medium rather than coarse grained, generally massive rocks with some porphyritic varieties. They are intrusive into the older rocks and in the mapped areas clearly truncate earlier structures. As understood at the moment they are largely adamellitic and some are associated with Be–Li pegmatite mineralization. Most of the analysed samples from the south half of the craton have $K_2O:Na_2O$ ratios between 1.0 and 1.5, but a few in the extreme south fall into the granite *sensu stricto* category of Harpum (1963) with this ratio greater than 1.5 (Bliss & Stidolph 1969). Whether the granodioritic Bulawayo fine-grained granite is part of this event is not clear. West of Bulawayo this cuts the earlier porphyritic granite dated by K–Ar at 2899 Ma (Amm 1940; Wilson & Harrison, in press).

The granites are developed as large masses in the southern part of the craton (figure 1). They are best known, from recent mapping, in the intrusions of the Chilimanzi, Chibi and Zimbabwe batholiths (Wilson 1964, 1968*a*; Worst 1956, 1962*a*; Robertson, in press) and appear to occupy much of the eastern part of the country (Phaup, in press), although mapping in this area is far from complete. North of Salisbury the major part of the Chinamora batholith is probably related to this 2600 Ma event.

Westwards the granites are smaller. The Nalatale granite (Harrison 1969; Stowe 1968*a*) almost severs the Fort Rixon–Shangani greenstone belt. The Formona granite (Macgregor *et al.* 1937) farther east is a coarse-grained intrusion; the Hillside syenite (Amm 1940) is a stock cutting the Shamvaian group rocks of Bulawayo. Several stocks also occur in the country between Shabani and Mashaba (Catherall, in press; Wilson 1968*a*).

On the regional scale the granites show an east-northeasterly trend (figure 1) which is very marked in the Chibi batholith but evident also from the northwest tip of the Chilimanzi batholith to the Nalatale granite and beyond. On a larger scale trends are also seen in and between the smaller granite masses between Mashaba and Shabani (Catherall, in press; Wilson 1968*a*, *b*).

In some areas the granites show gneissic margins (Bliss & Stidolph 1969), but in the author's experience in the south and central part of the craton most of the granites are even-grained and unfoliated right up to their sharp contacts with the older rocks. They display some thermal metamorphism with the local development of andalusite and cordierite in rocks of suitable composition (Wilson 1968*a*). Their final emplacement at the levels now exposed appears to have been passive, with stoping and fragmentation of country rock forming xenoliths.

The regional scale, however, presents a different story. North of Shabani Oldham (1970) suggests the emplacement of the Shabani younger granite caused updoming and late buckling of the synclinal structure of the Belingwe greenstone belt. Wilson (1968*a*, *b*) has attributed the deformation of the Mashaba igneous complex and surrounding area largely to the intrusion of these granites. Here the broad picture is one of the southwesterly slide on a grand scale of roof rocks off the rising dome of the Chilimanzi batholith to the northeast, with further buckling and final passive granite injection into the deformed structures. The major lubricants for the southwesterly slide were the ultramafic rocks of the Mashaba complex. The several stages of major deformation suggest several heaves of intrusion to the Chilimanzi batholith.

The Chibi batholith to the south is also probably, at least in part, responsible for the arcuation of the southwest margin of the Victoria greenstone belt.

The Zimbabwe batholith south of the Victoria belt probably tightened the earlier folds of the main greenstone belt deformation. Over the whole length of the southern part of this belt axial planes are overturned to dip steeply southwards towards this granite (Wilson 1964). The same feature is seen in the Mweza greenstone belt to the southwest (Worst 1962*a*). Here, however, as in the southern part of the Victoria belt, there is the added problem of the, as yet undetermined, effects on the craton of the nearby deformation in the northern marginal zone of the Limpopo mobile belt. The boundary of the mobile belt on figures 1 and 2 and on the latest official geological map of Rhodesia published in 1971 is the granulte facies limit.

Robertson (in press) suggests that the Chibi and Zimbabwe batholiths and perhaps even the Chilimanzi, are late stage derivatives of the granulite metamorphism of the mobile belt. A further, rather distinctive, porphyritic granite (figure 1) occurs cutting the granulite facies rocks in the northern marginal zone south of the Mweza greenstone belt (Worst 1962*a*). A similar porphyritic granite is described cutting the lower facies rocks of the craton on the south side of the Victoria belt (Wilson 1964). Robertson relates these intrusions also to the mobile belt suggesting that they are later than the Chibi and Zimbabwe granites and derived from the dying stages of the metamorphism.

In this respect van Breeman's (1969) study of the geochronology of the Limpopo mobile belt is significant. For the Bulai and Singelele granites near Messina in the central zone he obtained a Rb–Sr isochron date of 2690 ± 60 Ma. Bahnemann (in Morrison & Wilson 1971) considers the Singelele granite in particular to be remobilized and melted 'basement gneiss' and thinks this remelting took place during the peak of the metamorphism and deformation. He would regard this 2690 Ma age as dating the emplacement of the Bulai and Singelele granites, and the deformation and the high-grade metamorphism in the central zone (in van Breeman 1969).

8. The Great Dyke

The next major event in the craton was the emplacement of the Great Dyke and its associated satellites. Davies *et al.* (1970) have dated this intrusion at 2532 ± 89 Ma which is in good agreement with Allsopp's earlier minimum age of 2530 ± 30 Ma (Allsopp 1965). In both instances the results are from the Rb–Sr isochron method.

The Great Dyke is well documented (Worst 1958, 1960; Jackson 1970; Hughes 1970; Podmore 1970). Briefly it is an elongated mass of mafic and ultramafic rocks extending for nearly 500 km across the craton in a direction slightly east of north. Its maximum width is about 11 km. It is not a true dyke but the remains of four lopolithic intrusions arranged in a straight line and downfaulted into a graben-like structure (Worst 1958, 1960). The rocks are igneous cumulates with ultramafic types predominant in the present plane of erosion. All four complexes are similar in structure and component rock types and show marked layered sequences indicative of differentiation and crystallization in a stable environment.

To the east and west the Great Dyke is flanked by true dykes of quartz gabbro termed the east and Umvimeela dykes respectively. South of the Great Dyke are several satellite intrusions which are also true dykes but are all largely olivine-bearing melanorites (Robertson & van Breeman 1970). The largest of these, the main swarm, extends from the craton into and across the folded high grade metamorphic rocks of the northern margin zone of the Limpopo mobile belt. Dykes of this swarm have been dated by Rb–Sr isochron at 2600 ± 120 Ma (van Breeman 1968). Two narrower, north-northwesterly trending sets of dykes, the Crystal Springs and Bubi swarms respectively, are also assigned to the Great Dyke intrusive event (Robertson & van Breeman 1970). These occur entirely within the northern marginal zone and one of the smaller dykes of the Bubi swarm shows a chilled margin against the metamorphic rocks.

East of Fort Victoria the Popoteke and associated faults (Wilson 1964) cut the Victoria greenstone belt and extend northwards through the Umvuma greenstone belt and beyond (Bliss 1962). In places these faults are filled with quartz gabbro similar to that of the east dyke. They also show a marked sinistral displacement, part of which at least appears to pre-date the gabbro intrusion. In the Chinamora batholith northeast of Salisbury similar faults, some of which also contain a gabbro, are known (P. Snowden, personal communication). No petrological study has yet been made of this gabbro but the faults are likely to be a continuation of the Popoteke fault pattern.

No such sinistral movement can be recognized in the Great Dyke itself, although this would be difficult to detect, nor in the main swarm satellite at the south end which possibly represents a feeder to a Great Dyke type of layered complex (Robertson & van Breeman 1970). Nor is any such displacement reflected in the east and Umvimeela dykes. Their fractures possibly represent marginal hinge lines formed during the downsagging and downfaulting of the Great Dyke itself (Worst 1960).

Bliss & Stidolph (1969) describe other possibly related fractures, and they and Robertson (Robertson & van Breeman 1970) discuss the various hypotheses put forward to explain the Great Dyke fracture pattern. All three conclude that the Great Dyke occupies a dilation fracture. Although the fracture pattern cuts across the basement complex structures it is not a new trend but one seen in the major greenstone belts to the west between Sipolilo and Bulawayo, and is also apparent even in the earlier basement relics in the central part of the craton.

9. Discussion – evolution of the craton

Time: 2900 *to* 2500 Ma

During this approximate time-span it is possible to trace major events both in the craton and in the Limpopo mobile belt. The Mashaba igneous complex separates in time, at least in the southern part of the craton, the 2900 Ma granites from the 2600 Ma granites. In the mobile belt remelting and the peak of metamorphism and deformation is indicated at about 2700 Ma in the Messina area of the central zone. At about 2500 Ma the Great Dyke and its satellites were intruded into the craton and northern marginal zone of the mobile belt (figure 1).

Robertson (in press) has suggested that the intrusion of the Great Dyke magma, the intrusion of the 2600 Ma granites and the development of the Limpopo mobile belt may all be different manifestations of the same mantle process. To these events can now be added the intrusion of the earlier Mashaba igneous complex suite.

In a general discussion on crustal evolution, Sutton (1971) describes the development of the earliest, long-lived, rigid blocks, some of which have survived to the present, as somewhere between 2700 and 2400 Ma. In these blocks he notes several features of which one is the development of major ultramafic and mafic intrusions, including dyke swarms. These intrusions occupy regular systems of fractures indicating formation during regional stress. He suggests the possibility that they represent magma intruded into fracture systems developed in large segments of rigid crust when movements were taking place in adjoining mobile belts. Pretorius (in Sutton 1971) makes the likely suggestion that the Great Dyke originated as the Rhodesian craton failed during such deformation. The Mashaba igneous complex suite of intrusions probably had a similar but earlier origin during the early phases of development of the Limpopo mobile belt.

The picture that emerges is of a large crustal fragment of granite–greenstone terrain which in Rhodesia had already been subjected to the 2900 Ma granite event. Heat, generated by some mantle process, penetrated this crustal fragment along a linear east-northeasterly trending zone between what are now the Rhodesian and Kaapvaal cratons. In this zone the cratonic granite–greenstone terrain, in which an earlier east-northeasterly trend may have already been dominant, was reworked to produce the complex deformation and high grade metamorphism of the Limpopo mobile belt.

Early in the development of this mobile belt fracturing occurred in the still brittle cratonic crust on the northern foreland, and mantle activity, in response to the heat flow, resulted in the emplacement of the Mashaba igneous complex and related intrusions. Surface manifestations on the craton at this time were the *possible* development of greenstone belt depositories as at Belingwe.

In the main and later stages of the mobile belt activity granites were produced. These were emplaced in the southern part of the craton where some are perhaps derived directly from partial melting in the mobile belt. Similar granites, however, were also intruded well into the craton and apparently throughout the craton at about the same time, and indicate the extension of heat at this stage into the lower parts of the cratonic crust to produce widespread partial melting. The trends of the resultant granites suggest a linear distribution of these cratonic thermal highs sympathetic to the thermal high of the mobile belt.

After the emplacement of these granites stability of the cratonic and mobile belt crust was restored sufficiently to allow the development of the Great Dyke fracture pattern. Mantle response to the waning heat flow was, however, still enough to produce the Great Dyke magma.

10. The Mashonaland dolerites and the time-span 2000 to 1800 Ma

The next dated igneous phase in the craton is the intrusion of the Mashonaland dolerites (figure 2). In the northeastern half of the craton these occur as extensive undulating sheets in the granites. They extend west and southwest to cut the Great Dyke and beyond. Earlier K–Ar dates give a minimum age of 1600 Ma (McElhinny & Opdyke 1964), but recent total rock and mineral Rb–Sr isochron work on the Enkeldoorn sill and from five other sills has given an age of 1883 ± 21 Ma (Compston & McElhinny, in preparation; M. W. McElhinny, personal communication).

The Mashonaland dolerites have a well-established palaeomagnetic pole position (McElhinny & Opdyke 1964). Preliminary work on the Sebanga Poort dyke, a prominent north-north-westerly trending intrusive east of Shabani (Stowe 1968*a*; Wilson 1968*a*) has suggested a near Mashonaland pole (D. L. Jones, personal communication). This dyke is one of several described by Stowe in the country west of Selukwe. In this paper the writer has tentatively equated all these dykes and others to the west as part of the Mashonaland dolerite suite (figure 2).

The Mashonaland dolerites equate roughly in time with the event which produced the ages around 2000 Ma recorded by van Breeman (1968) from the Limpopo mobile belt. Mason (1970) describes this event as a further period of granulite metamorphism virtually confined to the central zone. The northern marginal zone is largely unaffected except for a reheating which also extended into the southern half of the craton and explains the K–Ar ages around 1800 Ma recorded from the Chibi, Zimbabwe and other granites in this southern area (Wilson & Harrison, in press). Palaeomagnetic results on the Great Dyke satellites (D. L. Jones, personal communication) can also be explained by such a reheating. The palaeomagnetic pole position of the Great Dyke is well established (McElhinny, Briden, Jones & Brock 1968) from sites along its length. The east and Umvimeela dykes also give the same pole positions along their lengths in the craton. In the northern marginal zone of the mobile belt, however, on the extensions of the east and Umvimeela dykes, the poles tend to occupy positions nearer that of the Mashonaland dolerites. The main swarm satellite dykes also give a normal Great Dyke pole position in the craton but again farther south, where these intrusions cut the metamorphic rocks of the mobile belt, there is the same tendency for their palaeomagnetic poles to plot nearer the Mashonaland dolerite position.

Granite rocks well into the craton have suffered a thermal overprint from the Mashonaland dolerites (Wilson & Harrison, in press) which makes necessary a reappraisal of earlier data (Vail & Dodson 1969; Bliss & Stidolph 1969; Bliss 1969) giving apparent ages younger than that of the Great Dyke for some of the granitic rocks of the craton.

11. Summary and conclusions

The author's main conclusions on the evolution of the Rhodesian Archaean craton are summarized in table 1. It must be emphasized that the ages 3300 and 2900 Ma are minimum limits only, based on apparent ages from K–Ar determinations. The value 3300 Ma in particular is from scanty data from the Que Que area in the central part of the craton.

There is evidence that relics of a basement (> 3300 Ma) to the main greenstone belt cover sequence are preserved in the central parts of the Rhodesian craton. Little is known of its early history. But these relics, together with evidence from the clastic sediments now occurring low in the preserved sequence of the main greenstone belts, indicate that by the time of deposition of these sediments, this early basement already existed as a granite–gneiss terrain incorporating there mains of what appear to have been earlier greenstone belts, and that it was undergoing erosion. These cover sediments also indicate that the granitic rocks of this early basement were not entirely sodic plagioclase varieties but included microcline-bearing types.

It is suggested that the depositories of the main greenstone belts evolved on this early basement on sites controlled by mantle-tapping fractures, and that the pattern of these fractures and the broad trend of the depositories are now reflected in the prominent north-northeast, east-northeast and northwest alinements seen in the present configuration of the main belts. It is further suggested that these depositories extended south of the present cratonic region, into and beyond the area that later became the Limpopo mobile belt.

The succession in the main greenstone belts is largely a progressive evolution involving an essentially volcanic phase followed by an essentially sedimentary phase which marks the end of major volcanism. The volcanic phase shows an initial assemblage in which ultramafic rocks are conspicuous. This is followed by a mafic to felsic assemblage. There is no reason to suppose that the evolution of the depositories was everywhere strictly contemporaneous, but the main greenstone belts are broadly contemporaneous in the sense that they were developed during the approximate time-span 3300 to 2900 Ma.

The structure and deformation of the main belts seem best regarded as largely gravity induced involving progressive downsagging of the volcanic and volcanic-sedimentary piles and the upwelling of granitic masses as described and discussed by various authors (Macgregor 1951; Anhaeusser *et al.* 1969). The downsagging initiated partial melting of the underlying sialic basement and produced granitic magmas which modified, and largely obliterated, any original basement–cover relationships. The granitic activity culminated in major intrusion at about 2900 Ma. The granitic types involved were tonalites with some granodiorites and adamellites. Whether addition to and progressive thickening of the sialic crust also occurred at this time is not clear, but after the intrusion of these 2900 Ma granites, the craton, at least in the south, became stable and sufficiently rigid to allow brittle fracture on a regional scale.

During the approximate time-span 2900 to 2500 Ma events can be traced in the craton and in the Limpopo mobile belt. In the craton events began and ended with the mafic-ultramafic plutonism of the Mashaba igneous complex and Great Dyke respectively. Separating these in

Table 1. Summary of the main events in the evolution of the Rhodesian Archaean craton

approximate age in Ma	cratonic area: main events	cratonic area: greenstone belt units		cratonic area: present group nomenclature of greenstone belts in the various areas			Limpopo mobile belt area: main events
2000 to 1800	Mashonaland dolerites and reheating in south margin of craton	—		—			reheating in northern marginal zone
2500	Great Dyke	—		—			Great Dyke satellites
2700 to 2600	granites ? greenstone belt(s) Mashaba igneous complex	sedimentary phase volcanic phase		(Belingwe area) Shamvaian Bulawayan			activation of the mobile belt; granulite metamorphism
2900 (minimum age)	granites			(most areas)	(Shangani batholith area)	(southwest of Selukwe)	
> 2900	main greenstone belts	sedimentary phase volcanic phase	 mafic and felsic ultramafic ± seds†	Shamvaian Bulawayan Bulawayan	Shamvaian Bulawayan Sebakwian	Nappe tectonics Bulawayan Sebakwian II OR Bulawayan Nappe tectonics Sebakwian II	main greenstone belts
3300 (minimum age)	granite(s)			(Que Que, Mashaba, Shangani batholith areas)		(southwest of Selukwe)	?
> 3300	? earliest greenstone belts	supracrustal remnants similar to lowest part of main greenstone belts		Sebakwian		Sebakwian I pre-Sebakwian	?
?	?	?					

† Seds = sediments. —— = unconformity.

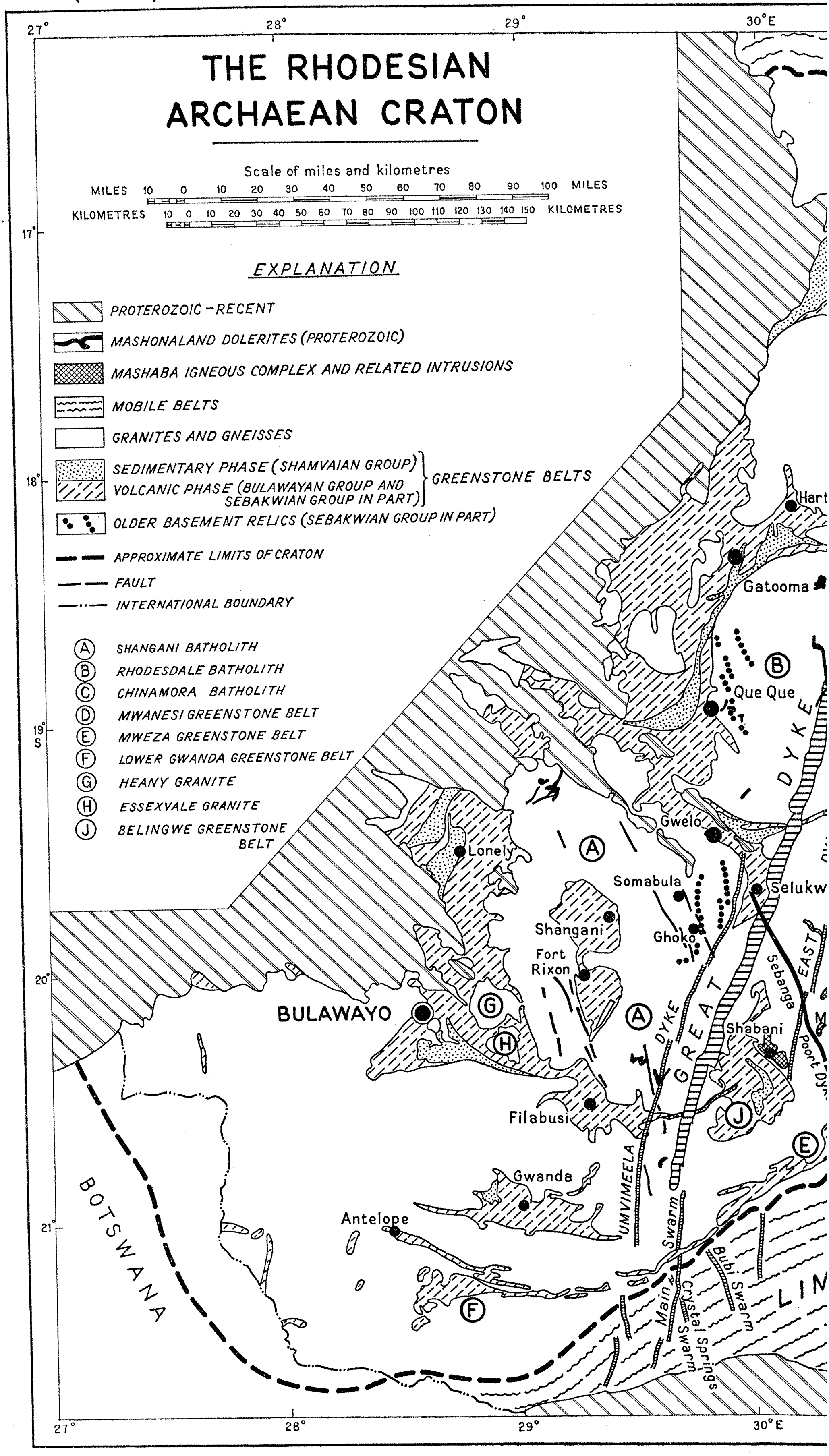
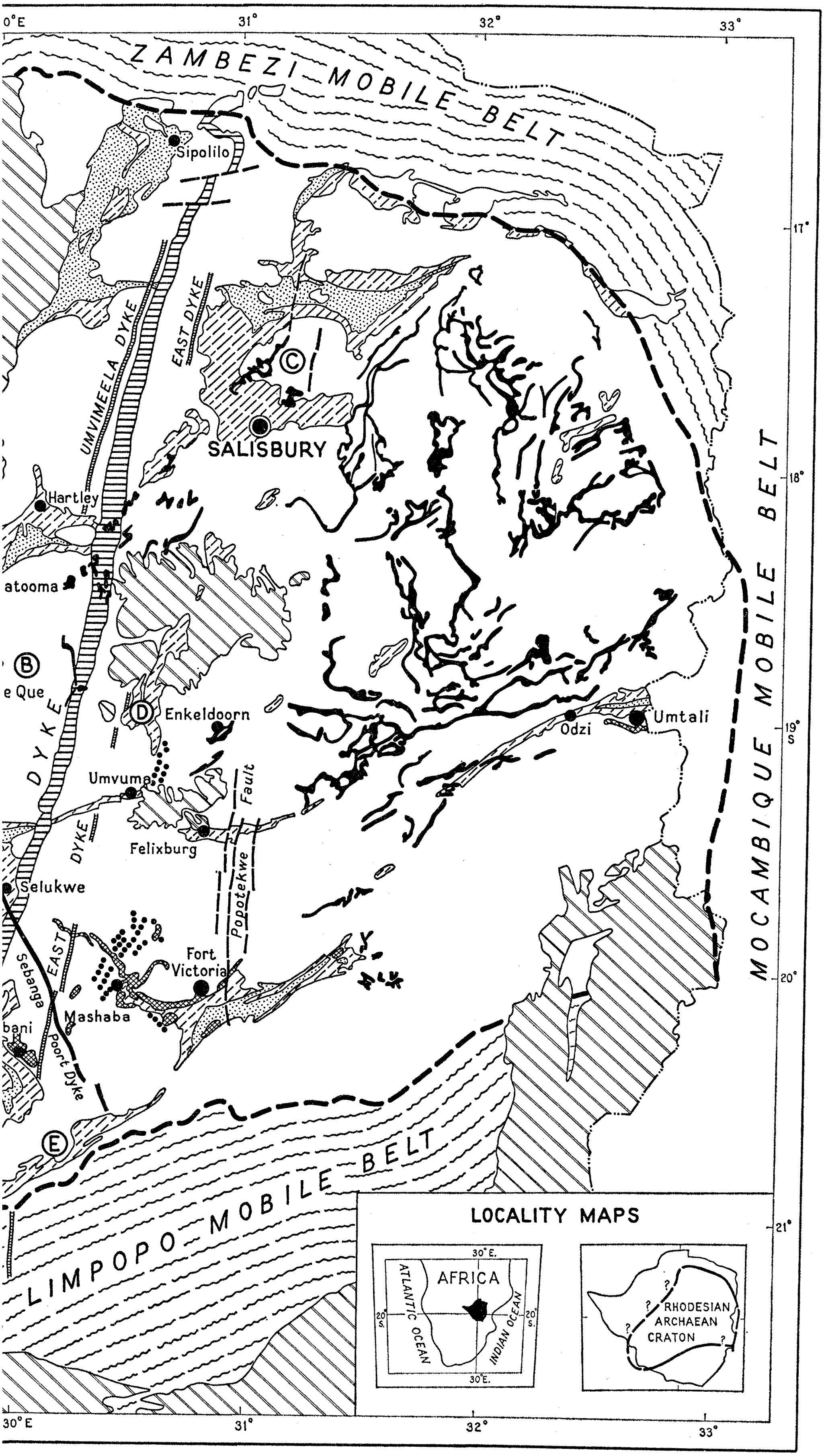


FIGURE 2. The Rhodesian



.hodesian Archaean craton.

(*Facing p.* 408)

time was the major granitic plutonism, largely adamellitic, of the '2600 Ma granites' possibly between 2700 and 2600 Ma.

In the Limpopo mobile belt the peak of metamorphism and deformation is indicated at about 2700 Ma from the Messina area of the central zone. By about 2500 Ma, at least the northern marginal zone was sufficiently rigid to allow fracture and the intrusion of the Great Dyke satellites.

It is suggested that the three plutonic episodes of the craton and the activation of the Limpopo mobile belt during this time-span are all manifestations of a changing heat flow patternt It is possible that the intrusion of the Mashaba igneous complex suite was contemporaneous with the early stages of further greenstone belt deposition, such as in the Belingwe area, but this requires further study.

At approximately 1900 to 1850 Ma the extensive suite of dolerites, comprising the Mashonaland dolerites, was intruded into the craton. Their intrusion equates roughly in time with a reheating which affected the northern marginal zone of the Limpopo mobile belt and at leas. the southern part of the craton, resulting in mineral ages in the 2000 to 1800 Ma range. A re-appraisal is therefore necessary of earlier data giving apparent ages younger than that of the Great Dyke (± 2500 Ma) for some of the granitic rocks of the craton.

It is apparent that the subdivision of the Rhodesian basement complex into a threefold sequence of Sebakwian, Bulawayan and Shamvaian systems each separated by granitic intrusion and orogenic movements, which imparted specific regional fold trends (Swift 1961) is no longer a realistic approach.

The present lithostratigraphic nomenclature retains these names with the substitution of the term 'group' for 'system' as the main unit of sub-division. Broadly, the Shamvaian group is the sedimentary phase of the greenstone belts and the Bulawayan group is the volcanic phase. The Sebakwian group is largely a sack term for supracrustal rocks regarded as older than the Bulawayan group.

This present terminology is still unsatisfactory and is confusing. Not enough attention has been paid to the detailed stratigraphy of the main belts and none of these groups is sufficiently defined in terms of stratotypes. In particular the Sebakwian group needs redefinition if any sensible use of the term is to emerge.

There is need and ample scope for further work on the rocks of the Rhodesian Archaean craton and of the Limpopo mobile belt. The areas of the 'something older' constituting the early basement need detailed investigation. The 'Sebakwian' assemblages of the Shangani batholith area, their stratigraphy, deformation and their relation to the Selukwe Nappe tectonics all need further work. Comprehensive chemical data on the granites and on the volcanic phase of the main greenstone belts, in particular on the lowest (ultramafic) assemblage, is still lacking. The structural elements from the craton into the Limpopo mobile belt are virtually unknown. The present geochronological knowledge is very inadequate.

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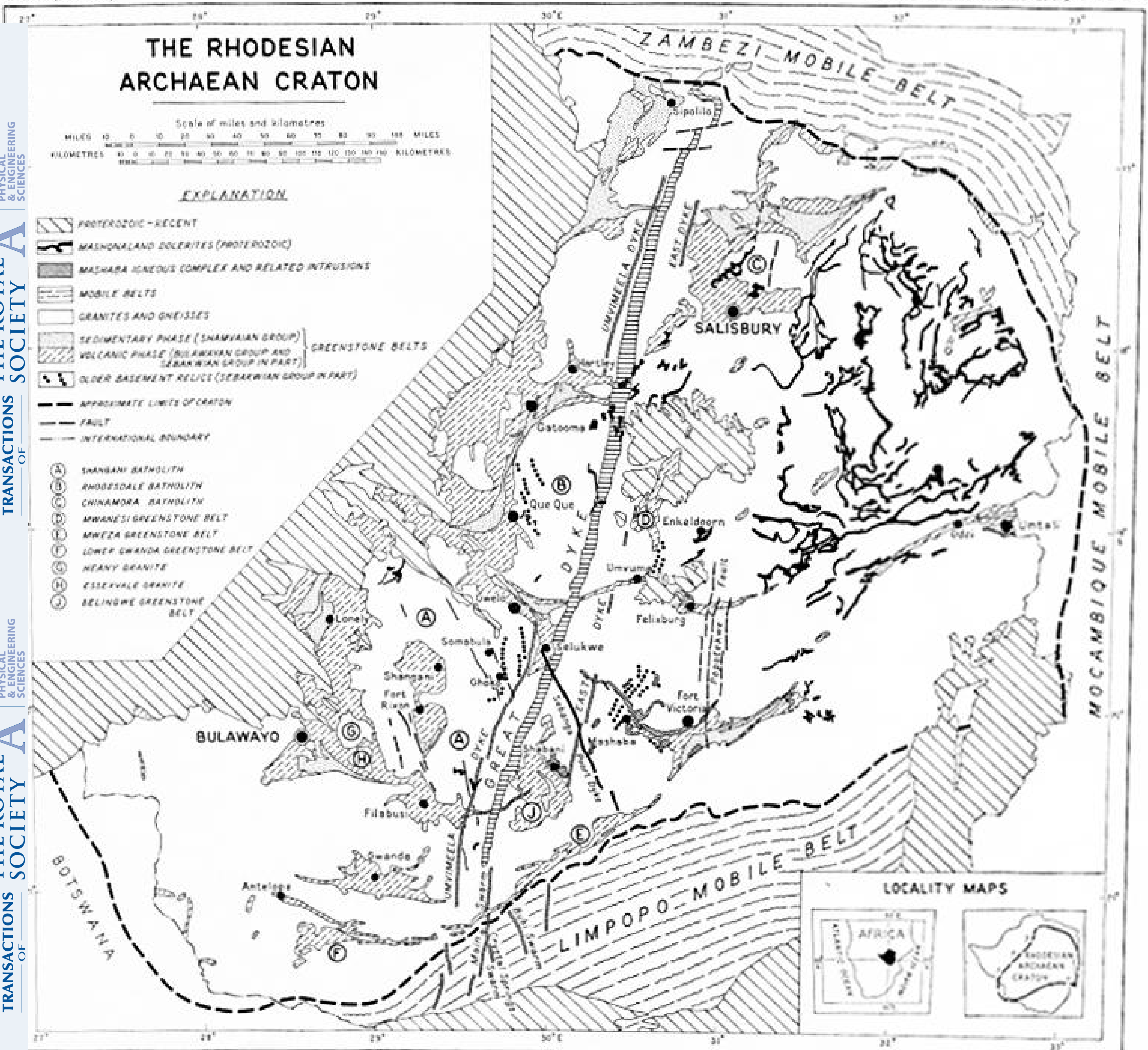

FIGURE 2. The Rhodesian Archaean craton.

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