ECONOMIC GEOLOGY RESEARCH UNIT

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ABSTRACT

Early Proterozoic red beds occur at the western, northwestern and at and near the northern margin of the Kaapvaal craton. Although these beds are virtually continuous they occur in distinct settings

. as intercalations in the Rooiberg Felsites and closely associated with elements of the Bushveld Complex (red beds of the Bushveld region (Bushveld red beds))

. in association with the Thabazimbi-Murchison, and Zoetfontein lineaments in the Waterberg region and in adjacent southeastern Botswana (Waterberg red beds); and in the Palapye region of Botswana in association with the Sunnyside shear and related faults (Palapye red beds)

. in the Northern Cape and southernmost Botswana (Matsap red beds)

. in the Soutpansberg (Soutpansberg red beds)

The commonest sedimentary environment is a braided fluvial one. Others within this suite include alluvial fan, deltaic and aeolian. Marine strata cap the continental sediments in the Northern Cape and at Palapye, and a marine influence may also be recorded at the top of the Waterberg red beds.

The limited available evidence suggests that these beds were reddened diagenetically by circulating groundwater or near-surface water. The Gamagara Formation is unusual in that detrital haematitic material, and possibly lateritic profiles also contribute to its red colouration.

The appearance of red beds as intercalations in the upper part of the Rooiberg Felsites represents a significant marker reflecting the development of increased oxygen in the atmosphere at that time. Other Bushveld red beds are intruded by the Bushveld Granite (Middelburg Basin), closely related to the unroofing of the Bushveld Granite (Nylstroom Basin), or in proximity to the Rustenburg Layered Suite (Otse Basin); these occurrences are preserved in small, but deep basins; sedimentation is controlled by surface, near surface, and upper crustal adjustments related to the Rooiberg Felsites and Bushveld magmatism.

Varying views on the significance, even presence of unconformities and consequent differing approaches to correlation issued have somewhat bedevilled understanding of the relationships within certain of the Bushveld red beds and also those attributed to the overlying Waterberg red beds. However, from a specific time, namely from the accumulation of the Alma Formation in the Waterberg, sedimentation is related to strike-slip movement on ENE-trending lineaments. Initially sediment was shed northwards from the Thabazimbi-Murchison fault zone, but it appears that major side-stepping soon took place with the left-lateral movement being carried on by the Zoetfontein fault to the north. The bulk of the Waterberg red beds were shed from this northern fault zone; the provenance is not an uplifted Limpopo Metamorphic Belt.

The Blouberg and Koedoesrand exposures may represent complex, and early, events on the Zoetfontein fault zone in which evidence of transtension and of transpression is preserved in close proximity along this transcurrent fault zone.

It is considered that the Palapye red beds may have a similar relationship to the northwest Sunnyside shear and related faults to that between the Waterberg and the ENE-trending lineaments.

The oldest haematitic beds in the northern Cape formed as slumped material in sinkholes in the Postmasburg-Sishen area in a period of erosion prior to the deposition of the Matsap red beds. The Matsap red beds are associated with significant basaltic volcanics and overlain conformably by marine sediments. Rifting at the western margin of the craton as now exposed is envisaged as responsible for the sedimentation. Subsequently, these strata were folded and thrust eastwards in the Kheis Belt; individual thrust sheets were translated more than 50 km in this event. Poor outcrop in such terrain leads to uncertainty in the overall succession and is only of the east-verging extremity of an overall belt: as a result, it remains uncertain whether the belt is ensialic or developed within a plate tectonic cycle.

In comparison, the Soutpansberg red beds contain little conglomeratic material, and significant amounts of argillaceous sediment higher up in the sequence. Up to half of the sequence is composed of tholeiitic basalt. The structure is complex, but is an extensional eastward-opening rift of aulacogen-type.

There are very widespread manifestations of $\sim$ 2000 m.y. magmatism in the Kaapvaal Craton. The most major of these are the lobes of the Bushveld Complex, and the Molopo Farms Complex near the northern margin of the craton. The elongated Groenwater gravity anomaly in the Northern Cape may reflect a similar basic/ultrabasic body at depth; if it does, then the overall Eburnian magmatism is concentrated eccentrically in a continental arc inside the setting of the post-Bushveld red bed sedimentation. This appears to be an instance in which major plume and thermal uplift has played a role in initiating subsequent crustal movements.

It is as yet uncertain how this magmatism relates to the Kaapvaal craton: of whether it represents a plate-margin phenomenon, or a trigger to ENE-trending movements related to a succeeding oblique collisional event. Strike-slip faulting dominated these later events in the Waterberg-southeast Botswana region and Palapye region, rifting in the northern Cape.

The best defined ages are the oldest, those related to the Bushveld Complex. Age data of the remaining red beds, particularly upper ages, is poor. Although a greater span in age remains possible, the available evidence suggests that all of them accumulated in the period 2100-1750 m.y. The Waterberg red bed sedimentation followed rapidly after the emplacement of the Bushveld Granite. Unroofing of the granite represents dip-slip movement in the strike-slip belt. The Palapye and Matsap red beds may well be broadly coeval with the Waterberg; specific correlations are, however, not warranted. Less clear is the age of the Kheis orogeny and that of the Soutpansberg rift. The Kheis orogeny can only be bracketed between 1800 and 1300 m.y. The Soutpansberg lies on the Zoetfontein lineament, but the extensional rift at this time reflects different, perhaps later, conditions on the same fundamental fault.

Few palaeomagnetic measurements have been made in these red beds - of these only one since the 60's. Available measurements, from Bushveld, Waterberg and Soutpansberg red beds have, however, contributed to the definition of an apparent polar wander course for Africa for the period 2050 - 1850 m.y. crossing northern Africa west-southwestwards.

Two of the palaeomagnetic measurements infer an age of $\sim$ 1100 m.y.; evidence of this age or a re-setting event in the Kaapvaal craton, comparable in age to tectonothermal events in the Namaqua and Natal belts to the southwest and south, is not well understood.

Most red beds are thought to have formed in low palaeolatitudes, either in dry or moist humid climates; many contain evaporites, or pseudomorphs after evaporitic material. Data on the palaeoclimate in which these red beds accumulated is very restricted. Evaporites have not (yet) been identified; the presence of evaporites is integral to the association of copper deposit with red beds.

Apart from the very major iron and manganese deposits immediately below, and at the base of the Matsap red beds in the Sishen-Postmasburg area, little else of economic value is known from these red beds. The issue, however, is whether or not the suite has been adequately researched.

A major type of copper deposit is associated with red beds worldwide. Such deposits tend to occur either at a first marine transgression over red beds or in non-red beds intercalated within red-bed sequences. Subsequent studies should focus on the marine strata lying above the red beds in the Northern Cape and Palapye, and the inferred marine influence at the top of the Waterberg succession. Further studies relating metamorphic grade and copper mineralisation in the Soutpansberg volcanics are also warranted.

At the present time, there is insufficient evidence to support the existence of unconformity-related uranium deposits, with PGE and gold deposits, in these rocks. Recent work now suggests that the Roxby Downs Cu -U - Au deposit in South Australia, associated with haematite, is hydrothermal and not an unusual type of sediment-hosted mineralisation as previously thought. As a result, it is no longer appropriate to consider this type of deposit in association with red beds.

EARLY PROTEROZOIC RED BEDS ON THE KAAPVAAL CRATON

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INTRODUCTION

A sequence of early Proterozoic rocks in which detrital red beds are dominant occurs in an arcuate sweep around the western, north-western, and at and near to the northern margin of the craton. Although these red beds are virtually continuous, a number of distinct settings may be distinguished:

1. The Soutpansberg rift, which extends for 200 km in the northern Transvaal, opening out eastwards and disappearing under younger cover

2. In association with post-Bushveld movement on major faults or shear zones
   (a) south of the Zoetfontein Fault, extending ENE from Blouberg into southeast Botswana, and south to the Bushveld Complex
   (b) on a continuation of the northwest Sunnyside Shear in the Palapye region of southeast Botswana.

3. Contiguous to the southern development of 2(a) in small basins in the Bushveld Complex: at Otse in Botswana, Nylstroom and Middelburg in the Transvaal.
   The oldest red beds in this occurrence are intercalations in the upper part of the Rooiberg Felsites underlying the main red-bed sequence.

4. In the Kheis Belt of thrusting and folding extending from near the Orange River in the northern Cape for 400 km into southernmost Botswana.
   The oldest haematitic beds in this region are in sinkholes reflecting erosion prior to the deposition of the main red-bed sequence.

Informally these sequences are referred to here as the Soutpansberg, the Waterberg and the Palapye, the Bushveld, and the Matsap red beds (Fig. 1).

The main part of this study reviews available information on the individual setting noted above. The format for this is not rigid, but in essence the information presented, where available, falls under the following headings:

. Distribution

. Thickness, stratigraphic relations

. Diagenesis

. Sedimentary environment (red beds)

. Determined palaeocurrent directions

. Palaeomagnetism

. Nature of the atmosphere

. Non red-bed sediments in the sequence

. Associated volcanics

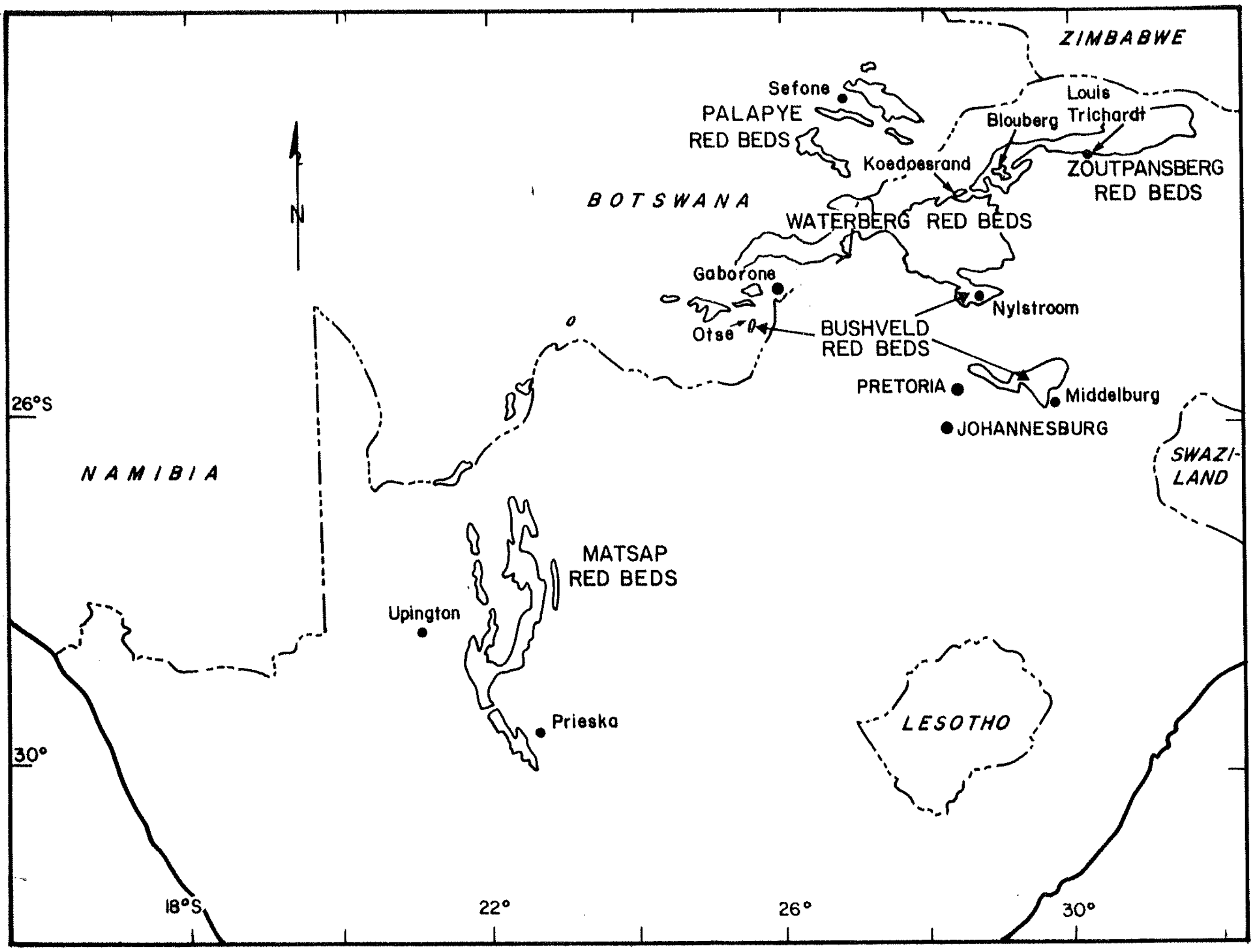

Fig. 1. Distribution of Early Proterozoic red beds

- Age/age relations/correlation
- Tectonic setting
- Relationship to plutonic bodies or tectonomagnetic events
- Structure
- Metamorphism
- Economic deposits

Using South African terminology, these red beds accumulated in the Vaalian and Mokolian Erathems (Johnson & Others, 1989); more precisely, in the later part of the Vaalian, and the Mokolian, probably restricted to the earlier part of this Erathem. Other classifications may perhaps be more relevant in this instance: all these red beds occurred in Proterozoic I of the Subcommission on Precambrian Stratigraphy (Plumb & Jones, 1986), or the equivalent time period, the Early Proterozoic, as defined by the USGS. This period covers a total time span of 2500 - 1600 m.y.; it is to this period that the title refers.

In specific studies within these red beds, such as those from the Waterberg Basin and the Soutpansberg rift, and in general comments, the importance of the Limpopo Metamorphic Belt as the provenance of these sediments has tended to be stressed (Button, 1973; Barker, 1979). Whilst much of this sediment has undoubtedly moved with a southerly component the evidence that this belt, per se, represents the major provenance area is, it is believed, not demonstrable.

It will be suggested that two other, probably inter-related factors are of greater significance in this regard:

. limited pre, syn, and immediately post-magmatic upper crustal subsidence and adjustments related to the widespread Bushveld magmatism and to other high-level intrusives emplaced some 2050 m.y. ago

. and, more significantly, distinctive crustal movements, related not only to the intracratonic Soutpansberg rift, and the rift at the present western margin of the Kaapvaal Craton in the northern Cape, but also to major strike-slip faulting extending WSW from the Waterberg region into southeastern Botswana.

CHARACTERISTICS AND SIGNIFICANCE OF RED BEDS

Before considering this sequence, it will be as well to take note of the characteristics of these distinctive rocks.

The reddening of continental clastic sediments to form 'red beds' is known from a range of ages and of tectonic settings.

Over the years a variety of mechanisms were proposed to account for the reddening of these sediments. It still remains possible for detrital, haematitic particles or lateritic soils to contribute in specific circumstances, but modern studies (see for example Walker & Others, 1978) have shown that the processes involved are diagenetic, that is, they form part of changes which the relevant sediment has undergone after inital deposition.

These processes include the progressive destruction of labile (unstable) mineral and rock fragments, by intrastratal solution; and the formation of new minerals - a process referred to as authigenesis, involving the formation in place of new minerals by replacement, recrystallisation, or secondary enlargement (of quartz grains). The nature of the intrastratal solution clearly is the major influence on subsequent authigenic minerals. Intrastratal alterations are most rapid during the early post-depositional stages, because this is the period when unstable detrital grains are most abundant.

Diagenesis is a most complex process, but nearly all the important reactions that take place at that time do so in an aqueous setting: either in waters which are near to the depositional interface, or in groundwaters which circulate in deeply buried sediments long after deposition. In a generalised way, the surface water/groundwater distinction is that between early and late diagenesis. But diagenesis may be represented by many consecutive changes.

The pigmentation itself may be derived from different sources: most notably from the 'ageing' of detrital iron hydroxides (enriched in fluviatile sediments) and from the intrastratal breakdown of detrital ferromagnesian silicates. Another source is represented by the alteration of magnetite in situ.

In the geological record continental red beds are known to occur in a range of tectonic settings such as failed continental rifts, or aulacogens; strike-slip basins associated with wrench or transform faulting; and in a variety of settings (as in the Caledonides and in the Alpine molasse) related to the later stages of major orogenics.

Tectonism may affect the rate of groundwater circulation, also interstitial temperatures and pressures; uplift of source areas will tend to increase the circulation rate and decrease temperature and pressure in depositional basins; burial on the other hand tends to decrease circulation, increase temperature and pressure.

There would appear to be an a priori argument for seeking to establish whether there are appropriate tectonic factors relevant to the formation of the early Proterozoic red beds on the Kaapvaal Craton. An extension of this is to seek for unifying factors of this nature.

Red beds occur in rocks of all ages back to the point in time at which sufficiently oxidised conditions first developed in the atmosphere. This occurred somewhat earlier than 2000 m.y., and red beds became abundant from 2000 - 1800 m.y. ago. The succession under consideration approximates to this time period.

Red beds are continental sediments, in which the most common environment of deposition is fluvial, but which do, in fact, occur through a range of alluvial, desert and delta plain associations (Turner, 1980). The following general comments are taken from Turner:

Alluvial red beds comprise a range of depositional facies ranging from marginal alluvial fans and coarse-grained braided rivers to alluvial plains with low sinuousity or meandering rivers. Alluvial fans and coarse-grained braided rivers typically develop in tectonically active sedimentary basins and are a particular feature of strike-slip basins developed on transform faults. They also occur as a marginal facies within failed continental rifts and in a variety of late-orogenic basins.

Desert red beds include a range of facies deposited in a hot, arid climate. They include the deposits of lakes, inland sabkhas, aeolian dunes and a variety of desert alluvial sediments. Desert alluvial sediments differ from the general alluvial category in being ephemeral, consisting of poorly sorted detritus. Aeolian sands are one of the most distinctive facies within the red beds although they are still poorly understood and may be more difficult to recognise in the stratigraphic record than is generally appreciated.

Delta plain red beds are deposited in deltaic complexes at continental margins and in a variety of tectonic settings. As the deposits of fluvial distributary channels and flood-plains they bear many characteristics in common with alluvial red beds, but are distinguished by the association of these sediments with deltaic sediments. The associated mudstones are colour mottled.

Red bed sequences are not uniformly red: many include variegated or drab colours. Table 1 lists depositional and redness characteristics in the differing environments.

TABLE 1 : THE CLASSIFICATION OF CONTINENTAL RED BEDS INTO MAJOR ASSOCIATIONS RELATED TO DEPOSITIONAL ENVIRONMENT AND COLOUR CHARACTERISTICS AND DIAGENETIC FEATURES

	Depositional Characteristics	Redness Characteristics
ALLUVIAL RED BEDS		
Alluvial fan-braided rivers	Pebbly alluvium Mostly channeldeposits with interbedded debris-flow and stream flood deposits	Uniformly red. Finer-grained horizons more intensely red
Alluvial plains with high or low-sinuosity streams	Sandy and muddy alluvium in FU cycles. Sand units are channel and bank deposits. Mudstones are flood plain deposits.	Variegated. Sandy units red or drab. Mudstone units red or variegated
DESERT RED BEDS	Cross stratified well-sorted sands with steep foreset in-clinations. Interbedded with inland sabkha and desert lake (gypsum and anhydrite) and poorly-sorted wadi deposits	Uniformly red
DELTA PLAIN RED BEDS		
Delta plains of river-dominated deltas	CU and FU cycles. Mainly flood-plain, well-drained swamp and lacustrine delta deposits	Variegated. Red beds usually confined to mudstones. A mottled appearance is common

FU fining upward
CU coarsening upward

Simplified after Turner (1980)

A common feature of many red beds is the occurrence of grey, green or white mottled zones: these are a secondary reduction feature.

The climatic significance of red beds has been the subject of debate. Early studies focussed on their association with hot, dry climates and the obvious development of red sediments in contemporary desert conditions.

More recently, it has been recognised that red beds may ocur in dry and most tropical climates. Walker (1974) distinguishes:

(a) a desert-evaporite red bed association in which red beds are associated with aeolian sands, desert fluvial sediments and evaporites formed in playa lakes and inland sabkhas

(b) a moist climate red bed association in which red beds are interbedded and interfingered (in rocks of appropriate ages) with coal-bearing strata.

Both the above associations indicate formation in low palaeolatitudes. If red beds do only form in such latitudes, it is important for the many sequences that lack direct climatic evidence.

Fossil magnetism, or paleomagnetism, involves studying the direction of magnetisation that iron-bearing rocks take on in the ambient geomagnetic field at the time of their formation.

Measurements of natural remanent magnetisation (NRM) allow for the definition of fossil pole positions: such studies are commonly linked to age measurements, allow for the definition of polar wander curves which in turn have relevance to an understanding of previous continental movements and reconstructions of individual continents. The measurements, and the interpretations of the applications, are complex and not discussed here.

The main types of NRM are: thermoremanent magnetisation (TRM), acquired by the cooling of magnetic minerals (in igneous rocks) through respective Curie points for those minerals; chemical remanent magnetisation (CRM), involving the formation of magnetic particles by low-temperature chemical reactions; and detrital remanent magnetisation (DRM), in which particles are aligned either during deposition, or on rotation in unconsolidated sediments.

All sediments carry iron oxides, but it is the red beds that carry significant amounts, and whose palaeomagnetism has been studied, in Precambrian and Phanerozoic rocks. Although red beds usually do contain detrital iron oxides, it is now accepted that CRM is the mechanism through which the majority of red beds were magnetised, and that haematite is the principal magnetisation carrier involving both pigment and specularite (Turner, 1980). The magnetic properties of haematite have been studied by many workers, but a comprehensive understanding of the mineral is still lacking.

Note also that there have been relatively few recent studies which have taken into account the twin complexities of diagenetic processes and magnetic properties. These two form a vital part of red bed studies.

Many red beds show classical palaeomagnetic results. They have two groups of directions, antipodally opposed, with positive and negative polarities apparently acquired during normal and reversed geomagnetic fields. The normal and reversed zones may be laterally persistent and it may be possible to correlate them in a way which does not conflict with the lithostratigraphy. The directions in red beds of this type are consistent with those from associated igneous rocks and field tests of stability usually indicate that the magnetization formed prior to folding. Red beds of this type are referred to as Type A red beds (Turner, 1979). They occur throughout the stratigraphic record and are generally believed to have been magnetized during or shortly after deposition.

A large number of red beds show more complicated palaeomagnetic results. They cannot generally be resolved into two groups of opposite polarity and usually reveal intermediate directions; more complex distributions of directions may be seen. Individual specimens can be shown to have composite magnetizations, either by specimen splitting or by chemical or thermal demagnetization. Red beds with magnetizations of this type are referred to as Type B red beds. It may be possible to isolate individual components by thermal or chemical demagnetization and, in some cases, these may be palaeomagnetically significant. Type B red beds, like Type A red beds, tend to be distributed throughout the stratigraphic record. Type B magnetization seems likely to have been acquired over a longer period, thus representing a more advanced stage of diagenesis.

A number of red beds are characterized by a magnetization which consists of a single, well-grouped direction and which is stable to thermal and chemical demagnetization. Field tests often indicate that the magnetization formed after folding and the direction is often widely divergent from that of associate or contemporaneous igneous rocks. These red beds are referred to as Type C; the direction is often close to the present Earth's field, a feature taken to indicate that the NRM was acquired in relatively recent times.

THE WATERBERG SOUTH OF THE ZOETFONTEIN FAULT IN THE NORTHWEST TRANSVAAL AND SOUTHEASTERN BOTSWANA, AND IN THE BUSHVELD REGION

A major area of Early Proterozoic red beds lies south of the Zoetfontein fault from a point to the west of Blouberg, and extending into southeastern Botswana.

In South Africa this region encompasses the Waterberg. Thabazimbi lies at the western limit, a point west of Potgietersrus the eastern limit, an extent of some 150 kms. Further south there are other areas lying within the central region of the Bushveld Complex: the Waterberg continues into the distinctive tectonic setting of the Nylstroom area, and the Middelburg Basin is a separate area lying near the south-central part of the complex.

The Waterberg Group crosses into Botswana between Buffelsdrift and Olifantsdrift, with outcrops occurring in a belt through Mochudi to Molepolole. The Zoetfontein fault has been traced under cover from aeromagnetic surveys (Hutchins & Reeves, 1980) to the base of the north-south-trending Kalahari line at Longitude 22°E. This fault is generally regarded as the northern limit of the Kaapvaal craton; there is no evidence other than in the Palapye region of any occurrences of these red beds to the north of the fault (Fig. 2; Meixner & Peart, 1984). In Botswana this occurrence has been referred to as the Notwani Sector.

To the south in Botswana there are further outcrops. The small basin at Otse, as was the case at Nylstroom and at Middelburg, represents a distinct tectonic setting adjacent to the Bushveld Complex. More 'typical' Waterberg can be traced as a relatively thin platform from the Kanye district through Dikgomo di Koe as far west as Khakhea (23°30'E). Much of the Waterberg lies under the Kalahari cover.

From Khakhea SSE limited surface information is augmented by borehole information from the Molopo Farms Complex Project (Gould & Others, 1987). In this area the Waterberg may be more extensive than shown from the limited drilling, and much of the western part of the Bushveld-type and -age Complex may lie between (older) Transvaal and younger Waterberg strata.

The Molopo Farms Complex is of special significance in that it occurs at the intersection of the north-south-trending Kheis belt of Olifantshoek strata, which can be traced as far north as Jwaneng, and the 'Waterberg' to the ENE. Gould & Others (1987) infer from this area that the rocks of these two trends are equivalent, thus that the much earlier correlation of the Matsap (now represented in particular by the Volop Group) and the Waterberg is correct.

A succession of mostly fluvial sediments has been established in these rocks in South Africa (de Vries, 1969; Jansen, 1982; Callaghan, 1987). The main area is regarded as composite, with 'Early' Waterberg in the Nylstroom area overlain by more extensive 'Middle' and 'Late' Waterberg to the north. 'Middle' Waterberg extends into southeastern Botswana. Other areas of 'Early' Waterberg sedimentation occur immediately across the Botswana border at Otse, and in the Cullinan-Middelburg region east of Pretoria. All the instances of 'Early' Waterberg are small, but contain thick sedimentation on or adjacent to the Bushveld Complex.

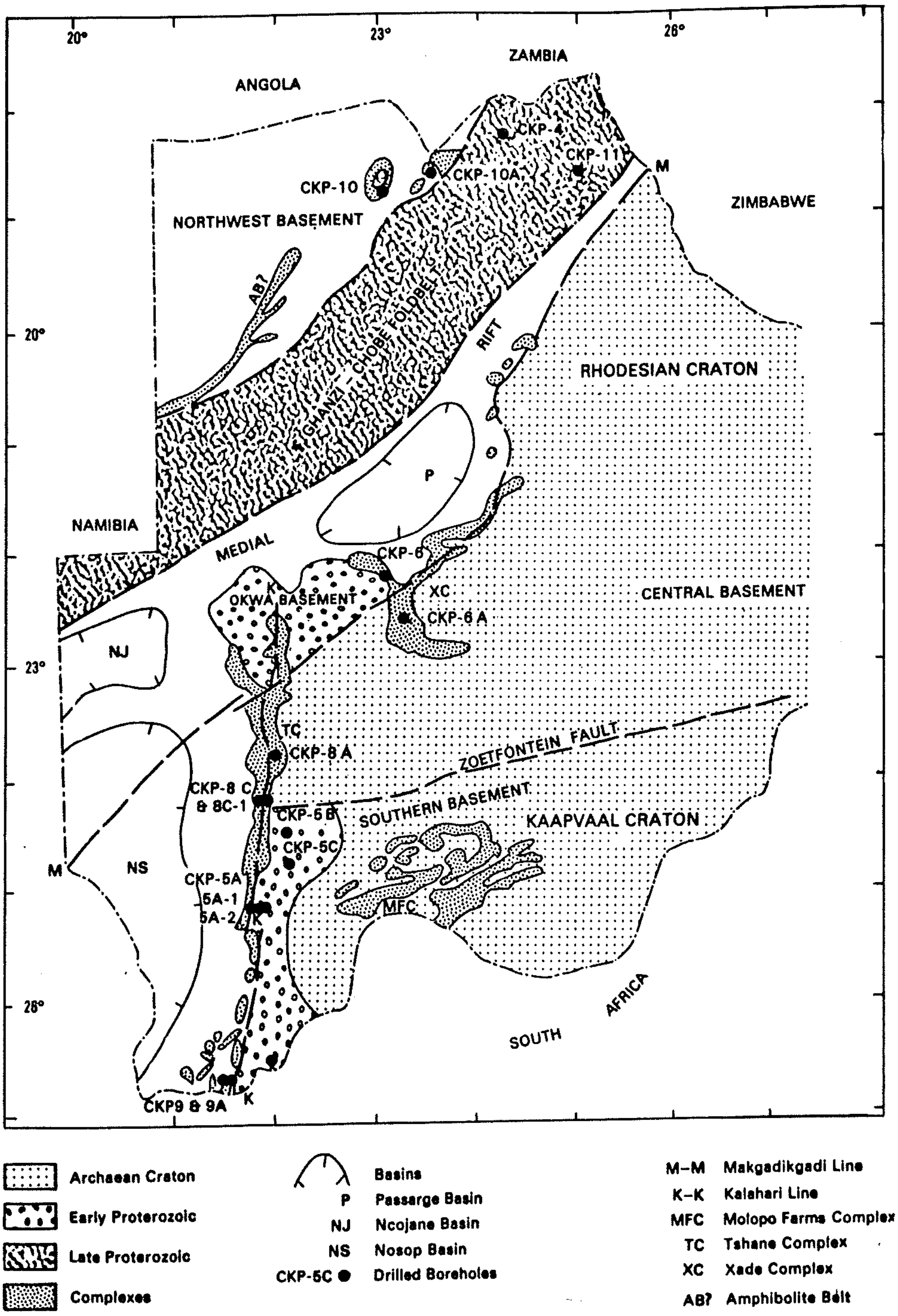

Fig. 2. The Kalahari drilling project: basement features and cored boreholes (Meixner & Peart, 1984)

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The stratigraphic subdivision established by de Vries (1969) forms the basis of the lithostratigraphic subdivision adopted by the Survey and by SACS (1980). Thirteen formations were recognised in this, but the Blouberg Formation, with its distinctive lithology and tectonised outcrops is now widely considered to be earlier than the Waterberg as developed elsewhere. A similar comment may apply to the Koedoesrand Formation. The Waterberg rests unconformably on gneisses and granites of the Kaapvaal craton, the Transvaal sequence including the Rooiberg felsites, and the mafic and granitic phases of the Bushveld Complex. That part of the succession which extends into Botswana also rests on the Ventersdorp, and the sub-outcrop of the Molopo Farms Complex.

Earlier red beds than these classified as Waterberg occur in the Transvaal: the pre-Bushveld granite Loskop Formation underlying the Wilge-rivier Formation in the Middelburg Basin and its suggested correlatives the Glentig Formation below the Swaershoekberge Formation in the south of the main basin and possibly the isolated occurrences at Rust der Winter; and also the even older red beds known to occur in the upper part of the Rooiberg felsites. The closely related distribution of these, and of the 'Early' Waterberg has long drawn comment (see for example van Biljon, 1976). With the exception of the Blouberg and Koedoesrand Formation, all the red bed sequences in the Transvaal are shown in Table 2. This table shows the units in relation to existing formal nomenclature, and relationships between individual units.

		NORTH, NORTHEAST, CENTRAL WATERBERG BASIN	SOUTH, SOUTHWEST AND WEST[1] WATERBERG BASIN	NYLSTROOM BASIN	MIDDELBURG BASIN
WATERBERG	KRANSBERG SUBGROUP 'Late Waterberg'	Vaalwater Formation — C(2) — Cleremont Formation — C — Mogalakwena Formation ≡	 — Sandriviersberg Formation		
		— C —	— C —	—	—
	MATLABAS SUBGROUP 'Middle Waterberg'	Makgabeng Formation G — C — Setlaole Formation G	Aasvoëlkop Formation — C — Skilpadkop Formation		
		—	— C/U —	—	—
	NYLSTROOM SUBGROUP 'Early Waterberg'		Alma Formation —SW— C/U — SE[4] — U.Swaershoek U.Swaershoek	Alma Formation — C — Swaershoek Formation	 Wilgerivier Formation
		—	— U — U —	— D —	— D/U —
TRANSVAAL	LOSKOP FORMATION		Glentig Formation	Rust der Winter Formation	Loskop Formation
		—	— C —	—	— C —
	ROOIBERG GROUP		Rooiberg Rooiberg	Rooiberg	Selonsrivier Formation

TABLE 2. Lithostratigraphic units and stratigraphic relations of the early Proterozoic redbeds in the 'Waterberg' - Bushveld region, South Africa (3). (Principal sources SACS, 1980; Callaghan, 1987).

1. West only applies to Aasvoëlkop Formation
2. C = Conformable, D = Disconformable, G = Gradational, U = Unconformable
3. Not shown: Blouberg Formation, Koedoesrand Formation
4. The Alma and Upper Swaershoek Formations grade eastwards into the Sterkrivier Formation.

CORRELATION AND SETTING: SOME ISSUES AND PRINCIPLES

Prior to further consideration of the red beds whose occurrence has just been outlined, there are a number of principles and issues bearing on their correlation and setting which it will be as well to introduce:

. that a very major time gap exists between red beds - most notably the Loskop Formation - considered to represent the last stage of sedimentation and volcanicity in a shrinking Transvaal basin; and the later development of Waterberg sedimentation - evolving from an original protobasin through a major expansion phase and ultimately to a shrinking, terminal basin

. that the red beds above the Rooiberg Felsites consist of a number of unconformity-bounded sequences, the oldest sequence equivalent to the Loskop Formation, the remainder representing the Waterberg Group as defined. In this approach, the individual unconformities are regarded as representing major breaks in the sequence

. the close relationship of certain early red beds with the Rooiberg Felsites, and of others with intrusive phases of the Bushveld Complex

. the role of strike-slip sedimentation.

A key to an understanding of the suggested time break referred to above, lies in the checkered history of the Loskop Formation (SACS, 1980).

Early workers (see for example Mellor, 1907), included these rocks with the Waterberg System, forming a part of the lower division of that system. Later, these predominantly argillaceous rocks were regarded as a sedimentary phase related to the underlying Rooiberg felsites.

In 1949, Truter introduced the term Loskop, named for Loskop Dam. In addition to occurrences in the Cullinan-Middelburg area, (lower) argillaceous sediments in the Waterberg Basin to the north and argillaceous rocks in the Soutpansberg were included in the Loskop System.

At the type locality, the contact with the overlying Wilgerivier Formation (of the Waterberg) is in general disconformable, but at several localities an angular unconformity is seen. Truter regarded this disconformity/angular unconformity as representing a major time break; thus the Loskop, and Waterberg, are shown as separate systems on the 1955 1:1 000 000 Geological Map of South Africa.

Many workers were unconvinced by this separation. Truswell (1967) for example, noted that:

"... Although localities do exist where the Loskop has been folded, even eroded, before the deposition of the Waterberg, the spatial distribution of these two, with the Loskop normally marginal to the Waterberg, and always occurring associated with it, taken with the lithological similarity of these rocks fails to convey the concept of a break of sufficient magnitude for the erection of the Loskop as a System. In all probability, we are dealing not with a regional unconformity but with local unconformities which "are typically developed around the margins of sedimentary basins that rose intermittently while continuous deposition took place in adjacent areas" (Krumbein & Sloss, 1963)... "

The Loskop was deleted from the 1970 1: 1 000 000 Geological Map of South Africa; the implication was that there was no significant time gap between the Loskop and the overlying Waterberg.

Early workers (see for example Daly & Molengraaf, 1924) had shown that Bushveld granite intruded the Loskop (then referred to as lower Waterberg) and that the upper Waterberg was younger, containing pebbles of this granite. Later studies, in the Loskop area in particular (Rhodes, 1972; Coertze & Others, 1977), demonstrated further intrusive relations between the Bushveld Granite and, specifically, the Loskop.

Noting the essentially conformable relationship between the underlying Rooiberg Group and the Loskop, and that the younger Wilgerivier Formation rests on Bushveld Granite, Coertze & Others, 1977, postulated that the Loskop was the last stage of overall activity in a shrinking Transvaal basin.

Waterberg sedimentation, represented in the Middelburg Basin by the Wilgerivier Formation and in the Nylstroom area by the Swaershoek Formation (overlying the Rooiberg Group), was seen as being significantly later. On the evidence then available, a time gap of at least 180 m.y. was envisaged between the pre-Bushveld Granite Loskop Formation and equivalents and the onset of Waterberg sedimentation. On the current (1985) 1: 1 000 000 Geological Map of South Africa, these earlier red beds are shown as the uppermost part of the Transvaal sequence.

The South African Committee for Stratigraphy still holds the view that there is a major break between the termination of sedimentation (and volcanism) in the Transvaal and the commencement of Waterberg sedimentation. Johnson & Others (1989) allude to the figure of 180 m.y. between the respective units, but there is little evidence today to quantify this figure. The Rooiberg can only be said to be older than the intrusive Bushveld Complex. The basic suite of this complex is well dated at $\sim$ 2060 m.y., as are the later granites at $\sim$ 2050 m.y. (Walraven & Others, 1990). All that can be said of the age of the basal Waterberg (as defined) is that it is younger than $\sim$ 2050 m.y. The concept of a major time gap can no longer be sustained.

The concept of Waterberg sedimentation as an evolution from an original protobasin through a major phase of expansion to a terminal shrinking basin had been questioned (Cheney & Twist, 1986).

Rather than seeing a relationship between patterns of deposition and present preservation, these authors consider that the extent of the Waterberg may have been considerably greater in the past, and that the thickest remaining parts of the sequence were not necessarily depositional centres: the thick sections are preserved in structural lows, in synclines and other areas of later subsidence. With such a concept, the thickness of Waterberg sedimentation is less than suggested elsewhere and does not exceed 5 km in any one place.

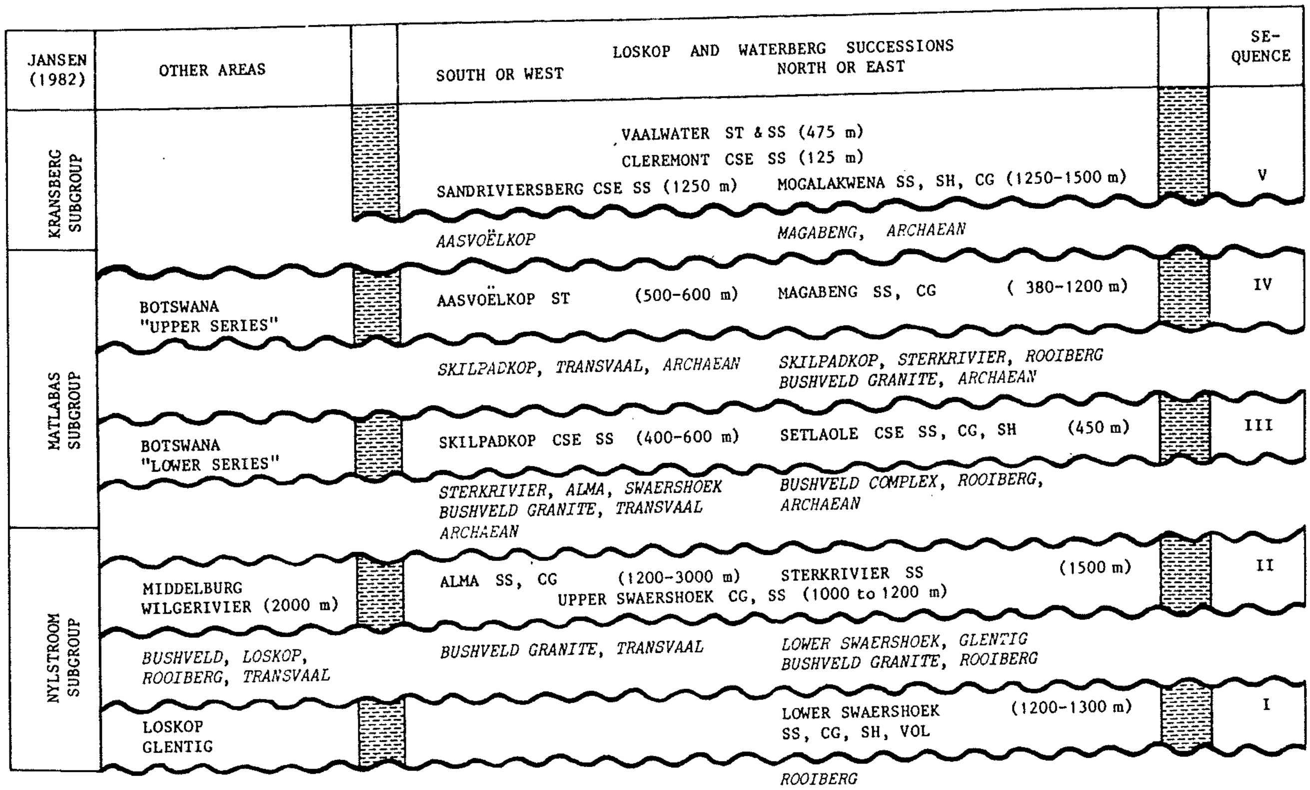

Fig. 3. Unconformity-bounded sequences in the Waterberg-Bushveld region (Cheney & Twist, 1986)

Italicised sequences represent the floor to specific sequences

Abbreviations: CSE - coarse; CG - conglomerate;
SS - sandstone; SL - siltstone;
SH - shale; VOL - volcanic

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Cheney & Twist (1986) consider that the overall succession consists of five unconformity-bounded sequences: as previously indicated, the oldest sequence (I) is equivalent to the Loskop Formation, the remainder (II - V) represent the Waterberg Group (Fig. 3).

Whereas Jansen (1982) had considered most unconformities in the Waterberg Group to be of local extent, Cheney & Twist regard them to be of much greater significance, and comment on them in some detail in relation to the following premises:

- a major unconformity is one that not only overlies associated sedimentary rocks, but also rests elsewhere on basement below
- the measure of angular unconformity between strata in an area is not a reliable estimate of the importance of the unconformity. This, that is the amount of section eliminated, can only be determined regionally
- formations which are, in fact, separated by a regional unconformity may appear to grade into one another vertically; this is likely where no angular relationship exists
- at the time of their formation, unconformities cutting continental strata are usually planar on a regional scale. Where this is not so, an unconformity showing considerable deviation from horizontality is (in the simplest case) likely to have resulted from post-depositional tectonism.

Cheney & Twist consider that because each of the sequences rests not only on one or more older sequences but also on crystalline basement, that each sequence is bounded by a major unconformity.

As is the case for all these early Proterozoic red beds, it is most difficult to establish an upper age limit of the sedimentation. Whilst recognising that the available evidence is meagre, Cheney & Twist considered that two age dates might strengthen a long time span for these sediments.

Components of the Pienaars River Alkaline Suite intrude the Wilgerivier Formation in the Middelburg area, Transvaal. The oldest of these intrusives is the Leeuwfontein grey syenite, for which Harmer (1985) reports an isochron age of 1425 $\pm$ 66 m.y. This date provides a minimum age for the Wilgerivier Formation and, by implication, for the correlatives in the main Waterberg basin further north. The possible existence of unconformity-bounded units in the Waterberg succession (Cheney & Twist, 1986) could imply a prolonged period of basin filling for the Waterberg Group. Cheney & Twist comment that, based on the present admittedly poor evidence, sequences II-V could be younger than 1420 m.y. There is no supporting evidence, however, that any Waterberg sediments are as young as 1425 m.y.

In the north of the area under consideration, the Palala granite is displaced by the Abbottspoort fault; this displacement does not affect the Makgabeng Formation (Jansen, 1976). An age determination then available for the Palala Granite of 1770 $\pm$ 30 m.y. (Burger & Coertze, 1975) provides a maximum age for that sedimentation. From these determinations, Cheney & Twist deduce that sequences II-V could be younger than 1770, sequences III-V even younger than 1420 m.y.

However, there is no genetic relationship between the Pienaars River Alkaline Suite and the red bed sedimentation; and the above age for the Palala Granite is now thought to be incorrectly based on composite samples. A single zircon has been dated at 1972 $\pm$ 62 m.y., an age suggesting that the Palala Granite approaches the Bushveld Granite in age (Brandl, 1986). Thus the available age data cannot be taken to indicate a major spread in age for the overall red-bed sequence, or indeed in support of the possible major time breaks envisaged.

The closely related distribution of the Rooiberg Felsites, the Loskop Formation (and its suggested correlatives) and the 'Early' Waterberg has long drawn comment (see for example Van Biljon, 1976; Fig. 4). Details of the relationships of these rocks, in particular with the Bushveld Granite, will be brought forward in the following section. Here a major conclusion of this review is anticipated: that the red beds of the Middelburg and Nylstroom basins relate in toto to the extrusion of the Rooiberg felsites and emplacement of Bushveld magmatism. The Otse Basin in southeastern Botswana is likely to have a similar relationship to the upper crustal intrusion of the Rustenburg Layered Suite. The point to be stressed is that the formation of these depositories relates to crustal adjustments consequent on magmatic processes.

A further conclusion is that the overlying red beds can be related very largely to strike-slip sedimentation on fundamental ENE-trending lineaments.

Comments on the red beds of the Bushveld region (the Bushveld red beds) will be presented in the ensuing section. Strike-slip sedimentation has received scant consideration and it is appropriate to provide a general background to this topic (below).

CONTINENTAL SEDIMENTATION IN STRIKE-SLIP FAULT SYSTEMS

Strike-slip motion has been shown to control the location, subsidence, shape, sedimentation and deformation of many basins (Ballance & Reading, 1980). Some such systems are related to fundamental faults with very long-standing complex histories. The mechanism is thought to have great significance: indeed, to represent an alternative to the Wilson cycle of plate tectonics, of particular applicability within continental plates and smaller basinal features (Reading, 1980).

Strike-slip faults are those whose primary motion is horizontal and parallel to the fault trace. Purely parallel movement of this kind is, however, rare: thus such transcurrent movement normally has an element of divergence, that is of extension, and is referred to as transtension; or of convergence, or compression, when it is referred to as transpression. Transtensile regimes are associated with normal faulting, basin formation and some volcanicity; transpression is marked by thrust and reverse faulting, folding and uplift.

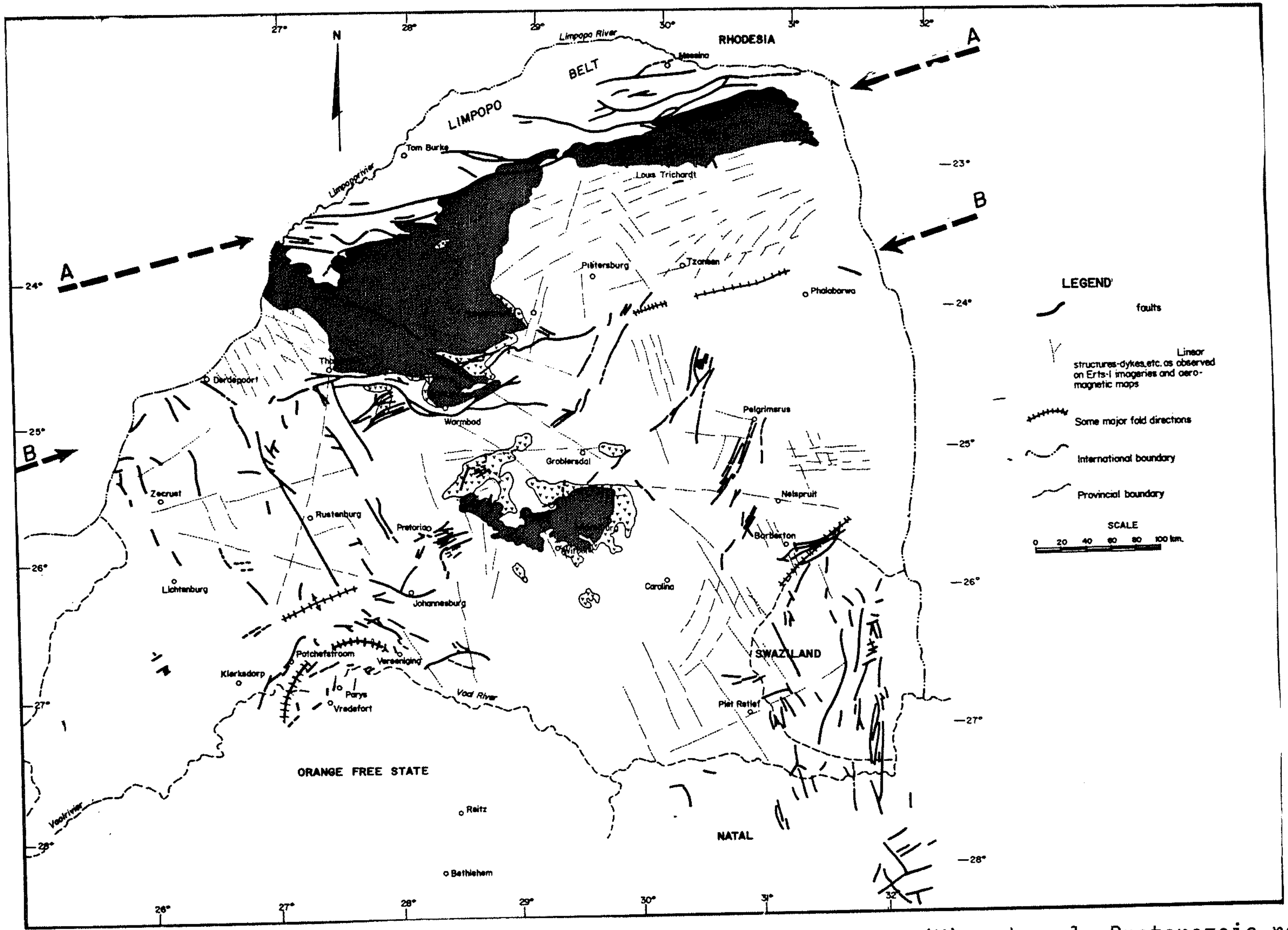

Fig. 4. Structural features, and the distribution of the Rooiberg Felsites (V) and early Proterozoic red beds in the Transvaal (van Biljon, 1976). A - Zoetfontein lineament; B - Thabazimbi-Murchison lineament. Reproduced with permission of the Geological Society of South Africa.

The faults themselves are complex: characteristically en echelon, with a range of bends, branchings, endings and offsets. Faults often die out and the motion may be taken up by an adjacent, side-stepping parallel fault.

Transpression leads to folding, thrusting and vertical uplift; transtension to the sinking of basins bordered by closely spaced normal faults.

Along a major strike-slip system, there may be quite small-scale alternate zones of extension and compression. These result from curvature along the overall fault system, the braiding of faults within the system, or side-stepping within an en echelon system. Within an anastomising system where faults converge, compression and uplift will result, divergence results in sinking.

At the end of a fault a 'pull apart' basin forms under conditions of extension, or folds and thrust may develop under a compressive regime.

Despite the dominantly horizontal displacement, the most obvious motion at any one place may be dip-slip. Large and rapid vertical movement is common.

Evidence that sediment was derived from the upthrown side is rare, except where fan deposits exist. Indeed palaeocurrents and other palaeogeographical evidence are more likely to show sediment transport parallel to the strike-slip fault rather than perpendicular to it (Reading, 1980). Because of the differential movements involved, sedimentary piles are usually several kilometres thick and have been deposited rapidly.

On land the most important initial depositional environment within a strike-slip fault-zone is lacustrine, typically bordered by alluvial fans. The fans are of restricted extent and involve the deposition of locally derived conglomerates and breccias. Bordering the strike-slip zones there may be wide alluvial plains, involving fluvial and interfluvial sedimentation. In transtension, basins form and relatively little material may be transported outside the belt; in transpression the sediment may be carried away from the source mountain belt.

Strike-slip belts are characterised by the absence, or at most low grade, of regional metamorphism: and also by sparse igneous activity; however, vulcanicity may be more voluminous where extension (transtension) is significant.

Hydrothermal gradients are usually low, but deposits of pyrite, gypsum and barite have been reported along continental strike-slip faults. In addition, many workers believe that ore bodies are located along major lineaments or crustal shears.

The concept of unconformity-bounded sequences is of considerable significance in relation to an understanding of a number of the cover sequences on the craton. The unconformity between the Chuniespoort and Pretoria Groups in the underlying Transvaal 'Sequence' represents a classic boundary between two such sequences. Cheney & Twist (1988) consider that a previously unrecognised unconformity truncates the Pretoria and Chuniespoort Groups; they refer the Transvaal strata above this surface, represented by the Dullstroom, Rooiberg, Smelterskop and Loskop, to the Bothasberg sequence.

More problematical is to regard UBS's as a panacea applicable to all the cover sequences on the craton. it is contended that it is not an appropriate concept either to the localised effects related to the Bushveld magmatism or, similarly, to strike-slip sedimentation.

By the very nature of strike-slip belts, unconformities and fold phases where they occur are bound to be limited in extent and contemporaneous with a continuous sedimentary column not very far away. In such terranes, correlations involving the supposed synchroneity of unconformities are thus of doubtful value (Reading, 1980).

THE BUSHVELD RED BEDS

The contention has been made of an early suite of red beds in the Transvaal-southeast Botswana which accumulated at or about the time of overall emplacement of the Bushveld Complex. In this section information is presented on these occurrences: subsequently the suite is discussed further.

Initially the earliest red beds, within the Rooiberg Felsites, are described; then overlying red beds in the Middelburg and Nylstroom areas; and also the occurrence at Otse in Botswana.

RED BEDS IN THE ROOIBERG FELSITES

The Rooiberg Group is a voluminous succession ($\sim$ 5000 m thick) of felsic volcanic rocks lying conformably above the Pretoria Group.

Although there are no reliable age data on the rhyolites themselves, field relations are unequivocal in showing that these rocks are pre-Bushveld Complex, > 2061 m.y. (Walraven & Others, 1990). The felsites are regarded as the immediate precursor to the plutonic episodes of Bushveld magmatism; although the original extent of the volcanism is not known this occurrence ranks amongst the largest known accumulations of silicic volcanic rocks (Twist & French, 1983).

Field studies have distinguished a lower Damwal Formation from an upper Selonsrivier Formation (Clubley-Armstrong, 1977). This distinction has relevance to a consideration of red beds, and is also thought to be significant in terms of the timing of the development of an oxygen-rich atmosphere (Twist & Cheney, 1986).

Lavas of the lower Damwal Formation are mainly dark (brown, green and black), whilst the Selonsrivier volcanics are usually pinkish red. The latter colouration results from the pervasive exsolution of dusky haematite from an original glassy matrix of the volcanics.

In the Damwal Formation pyroclastic rocks usually contain fragments of dark coloured lava, whereas in the Selonsrivier Formation red fragments predominate over black. The co-existence of different coloured clasts implies that haematisation occurred during or very soon after solidification of the flows (Twist & Cheney, 1986).

Where developed, sedimentary intercalations in the Damwal are typically buff or grey; and red colouration reflects secondary weathering. However, the intercalations of sandstone and shale, and of volcaniclastics in the Selonsrivier are red along strike lengths of several kilometres. These strata contain shallow water and some subaerial sedimentary structures (channels, ripple-marks, mudcracks), and closely resemble the widespread, overlying red beds in overall appearance.

These are the oldest red beds (> 2061 m.y.) in the Transvaal. Assuming that the pre-Gamagara weathering and erosion cycle with which the red beds of the Sishen iron-ore deposits are associated is younger (this need not be so), these red beds would be the oldest in southern Africa.

Geochemical investigations have revealed no significant differences in either major or trace elements above and below the Damwal-Selonsrivier contact. Thus the colour differences in these formations do not reflect chemical changes or control (Twist, 1985), and the possibility exists that the differences reflect change in atmospheric conditions, with oxygen becoming abundant in the atmosphere at that time.

A major change in atmospheric conditions early in the Proterozoic has long been mooted (see for example Cloud, 1976; Eriksson & Truswell, 1978). The topic is complex but lines of evidence include the presence of detrital pyrite and uraninite in the auriferous conglomerates of the Dominion, Witwatersrand, Ventersdorp and Black Reef, but not in younger sequences; the appearance of thick red-bed successions relatively early in the Proterozoic; and the disappearance of major banded iron formations at about the same time, perhaps with some overlap.

Twist & Cheney (1986) contend that the Damwal-Selonsrivier contact represents a good marker reflecting increased oxygen in the atmosphere at this specific time; and that such a time line is better pinpointed than the erosive interval between the Griqualand West and Olifantshoek noted as reflecting a similar atmospheric change in the northern Cape (Beukes, 1983). Another succession thought to span this atmospheric transition lies within the Huronian of Canada (Mossman & Harrow, 1983); an age of 2288 $\pm$ 87 m.y. is suggested for this transition; in the Transvaal the available age data are not precise enough for more specific comment.

MIDDELBURG BASIN

Strata of the Loskop Formation, and of the Wilgerivier Formation, are developed in the Middelburg Basin.

Loskop Formation

The Loskop Formation consists of some 1200 m of red bed sediment, with interbedded lavas and pyroclasts. The sediments are more argillaceous than others of these red beds; in addition to shale and siltstone, other rock types are sandstone, quartzite, felspathic sandstone, conglomerate and breccia (Coertze & Others, 1977). A prominent conglomerate at the base is composed of volcanic matrix and pebbles, particularly of quartzite and acid lava.

Acid lavas, and also epidotised, intermediate to basic lavas occur near the base of the Formation. The essentially conformable relationship with the underlying Selonsrivier Formation (of the Rooiberg Felsite Group), and the increasing number of sedimentary intercalations in the felsites, provide further evidence of a continuous succession from the Rooiberg to the Loskop.

Near Loskop Dam a porphyritic Bushveld granite intrudes basal Loskop strata (Rhodes, 1972). The porphyritic texture becomes coarser-grained and more equigranular near the centre of the body.

The outcrop of the Loskop Formation is peripheral to that of the Wilgerivier Formation (Fig. 5). The Loskop is usually flat-lying or dips at low angles; locally dips are steep.

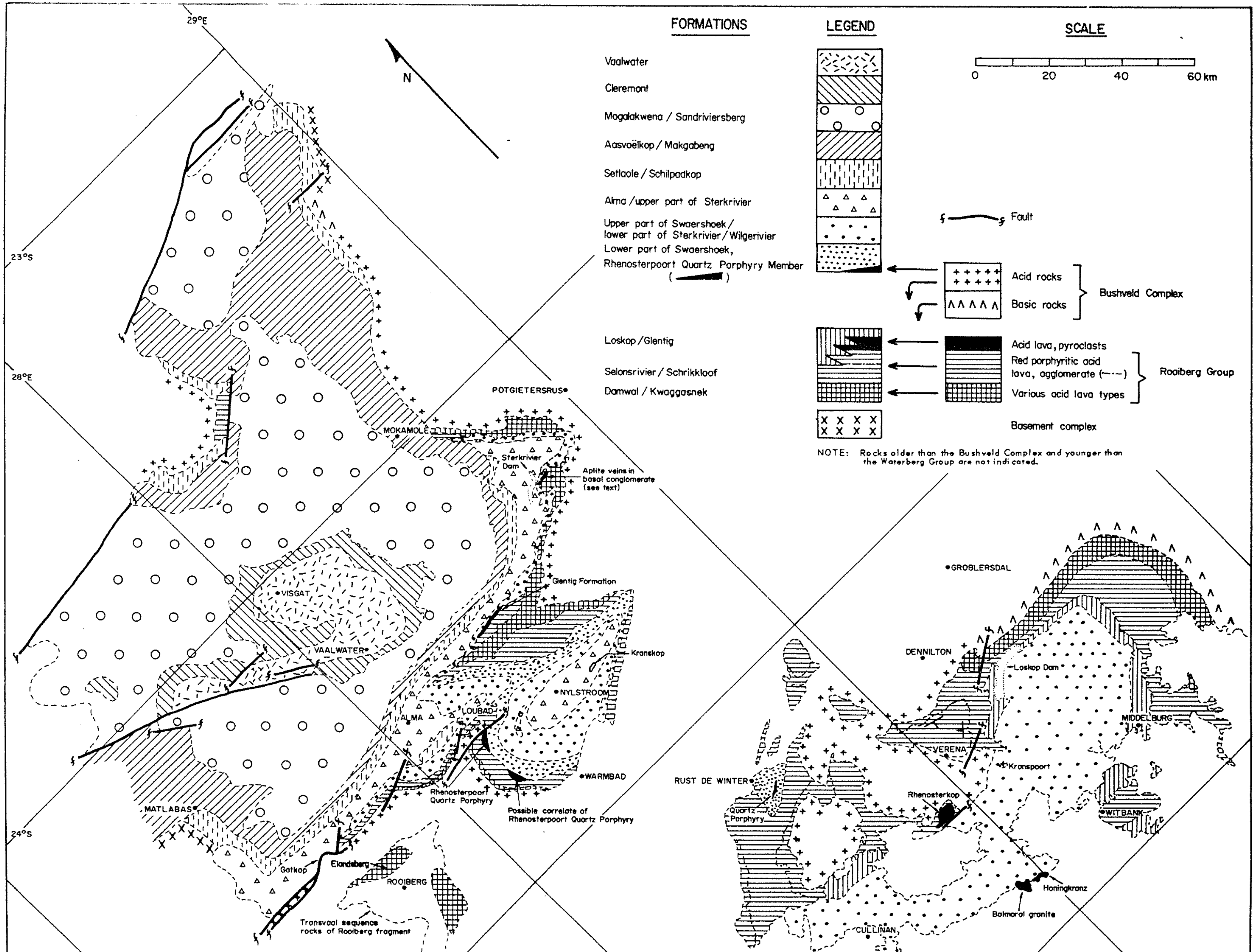

Fig. 5. The distribution of lithostratigraphic units in the Rooiberg Group, and the Bushveld and Waterberg red beds (Coertze & Others, 1977). Gross subdivisions as indicated by Coertze & Others.
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In the area north of Warmbad, a limited outcrop of sedimentary rocks, succeeded by volcanics occurs on the northern slope of the Swaershoek-berge. This is the Glentig Formation, which also conformably overlies the Rooiberg Felsites: some 100 m of red to purple argillaceous rocks, sandstone and altered lava are overlain by approximately 50 m of conglomerate; above which is up to 500 m of faintly banded quartz-felspar porphyry. The Glentig Formation and an isolated occurrence at Rust der Winter between the Middelburg and Nylstroom Basins are considered to be correlatives of the Loskop Formation (Coertze & Others, 1977).

Wilgerivier Formation

The Wilgerivier Formation is a monotonous succession of brownish to purplish medium to coarse-grained sandstone and quartzitic sandstone with few intercalations of conglomerate and shale. Although lithologically similar to the Swaershoek Formation, this succession contains no intercalated volcanics.

The maximum thickness is 2000 m. Palaeocurrent measurements from a 20 km-wide belt in the basin (Vos & Eriksson, 1977; Fig. 6) indicate a continual easterly to westerly alternation of transport irections in the fluvial sediments. Relevant structures in the rocks are planar and trough cross-bedding asymmetric or current ripples, and primary lineation.

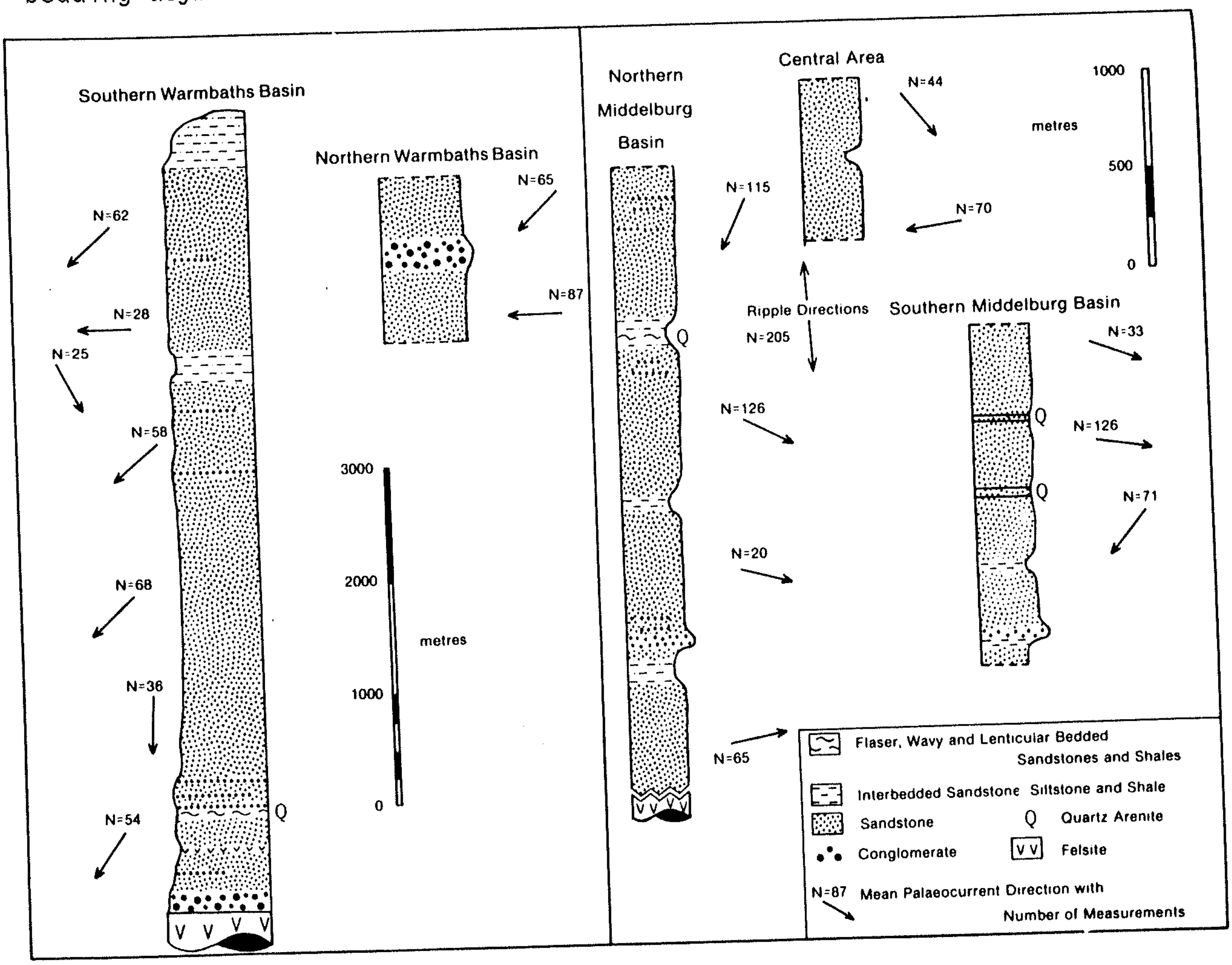

Fig. 6. Palaeocurrent means in the Waterberg and Middelburg Basins (Eriksson & Vos, 1979).
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Vos & Eriksson (1977) considered the Waterberg sediments in the Middelburg area. Within a fluvially dominated sequence they suggested that very limited tidal and wave-swash deposits occurred in embayments between fluvial fans. In a more general study of the Waterberg by these authors (Eriksson & Vos, 1979), they indicated that sedimentation occurred mainly on proximal to distal reaches of braided alluvial plans under conditions of lateral switching of the active river system associated with prolonged vertical aggradation in a gently subsiding intracratonic depository. Relatively rare quartz-arenites developed as a result of weak tidal and wave-swash reworking of the fluvial sediments.

The Formation is cut by several alkaline complexes of the Franspoort line. One of these, at Leeuwfontein, is dated at 1425 $\pm$ 66 m.y. (Harmer, 1985); note, however, that this specific complex does not cut the Wilgerivier Formation. As previously noted there is no connection between the sedimentation and this alkaline magmatism: this age, then, is no more than a truly minimal age for the sedimentation, which is generally regarded as being significantly older than the age of this complex.

NYLSTROOM BASIN

Swaershoek Formation

The distribution of the Swaershoek Formation within and immediately north of the Nylstroom Syncline is very limited, and different from that of the overlying formations.

Informally, a distinction is made between lower and upper Swaershoek.

In the Nylstroom Syncline a complete succession overlies the Rooiberg Group, but to the north only the upper portion is developed and it overlies Rooiberg lavas unconformably.

The Swaershoek Formation is a predominantly arenaceous succession consisting of sandstone, locally with interbedded shale, siltstone, pebbles and boulder conglomerate, and trachyte. The top of the formation is a persistent 5 m-thick pebble rudite.

At its maximum, the formation is 2500 m thick; it is lenticular and thins very rapidly. There are thick lutites (up to 250 m thick) in places; their origin is uncertain and may be related to the extrusion of lavas.

The lowermost beds were chaotic deposits on a very irregular topography; their deposition is thought to have included sand and mud flows involving debris flow (Jansen, 1982). Jansen also considers that contemporary volcanism may have caused heavy ash flows and lahars.

Alma Formation

The Alma Formation consists of fine-grained felspathic sandstones; sandstones and arkoses are less common. The contact with the underlying Swaershoek is gradational; thickness 700 m (these comments refer specifically to the Nylstroom Syncline).

OTSE BASIN

In general, the Waterberg in southeastern Botswana forms a thin capping of pink, red, or purple arenaceous rocks, on hills often overlying felsic igneous rocks. In the Otse area this is referred to as the Mannyelanang Formation.

A thicker sequence does occur, however, in a small area around Otse. In this setting the depositional environment was an unstable one leading to the localised deposition of various slumped, and alluvial plain deposits derived from the west (Crockett, 1971; Crockett & Jones, 1975; Key, 1983) No volcanic rocks occur.

At Otse different facies of sedimentation, that is rather than a stratigraphic sequence, are represented by the Maladiepe Hill Formation, Moeding Formation, and Ditsotswana Formation (Fig. 7.).

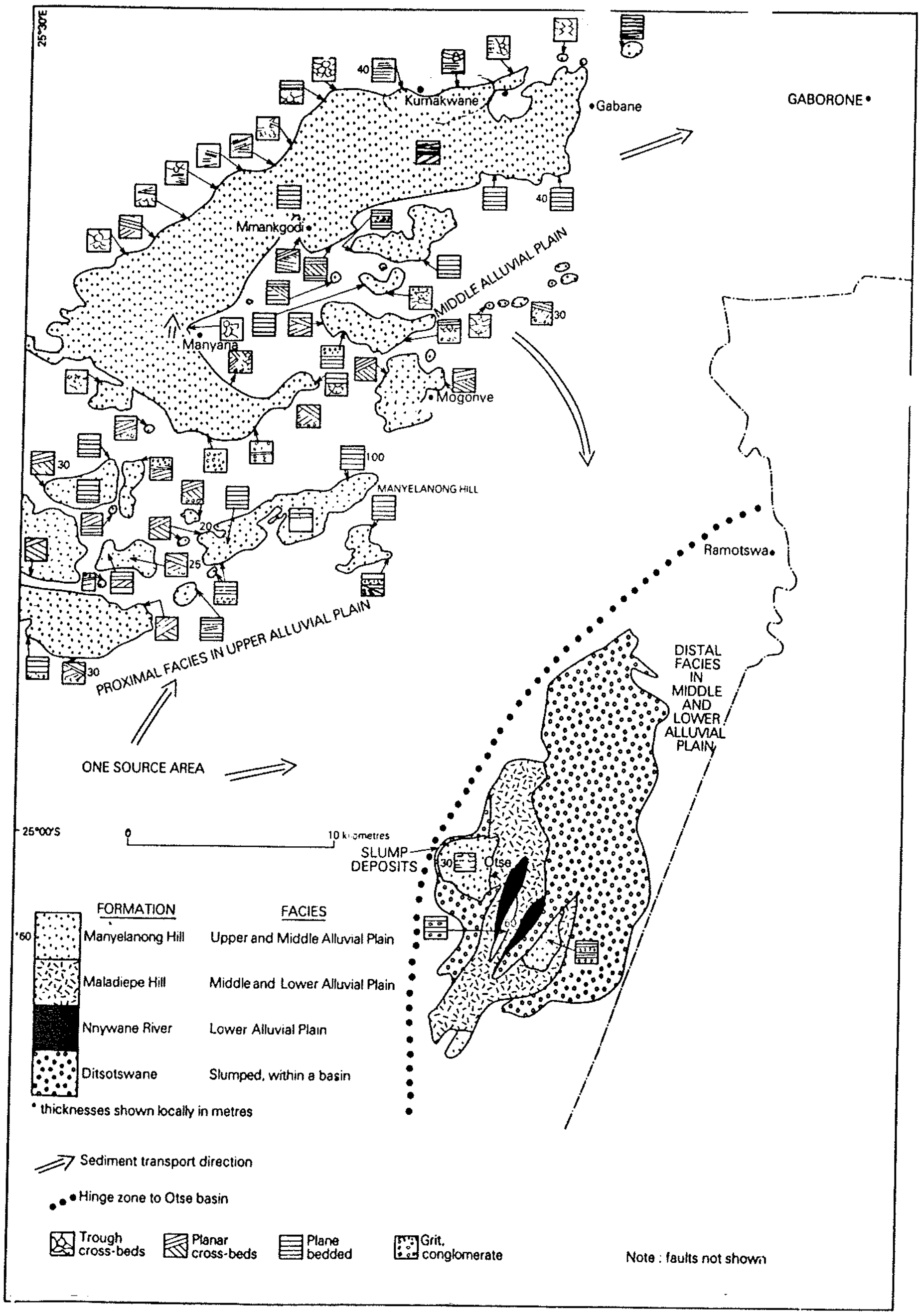

Fig. 7. Early Proterozoic red beds in the Otse Basin, and environs (Key, 1983). Reproduced with permission of the Director, Geological Survey of Botswana.

	Thickness (m)	Lithology	Environment of Deposition
Maladiepe Hill Formation	240 m	Interbedded red sandstones and shale-chert conglomerates	Mid alluvial plain
Moeding Formation	?	Red, micaceous shales	Lower alluvial plain
Ditsotswana Formation	?	Chert and dolomite (+ quartzite) conglomerate	'Slump deposit'

Table 3 . The Otse Group. Information from Key (1983)

The Waterberg strata in southeastern Botswana are generally flat-lying. However, some monoclinal structures are developed: those at Kolobeng and Moganye dip steeply parallel to ENE-trending faults. At Otse the strata dip west (and strike north-south). The inclination of the strata is taken to relate to further movement along the hinge zone referred to earlier (Crockett & Jones, 1975).

DISCUSSION

Coertze & Others (1977), and Jansen (1982) regarded the evolution of basins in Transvaal and Waterberg times as following the same pattern: of protobasins to large depositories and then to shrinking basins. Coertze & Others (1977) note that the two major sedimentary cycles are not entirely separated by the emplacement of the Bushveld Complex.

In this approach the Loskop (and its correlative the Glentig) Formation represents the final stage of the Transvaal cycle, this unit being post-Rooiberg in age but intruded by the Bushveld Granite. A major interval separates the Transvaal from the initial stage of the Waterberg sedimentation, represented by the lower part of the Swaershoek Formation in the Nylstroom Basin.

Coertze & Others (1977) and Jansen (1982) regarded the upper part of the Swaershoek Formation north of Nylstroom as a correlative of the Wilgerivier Formation in the Middelburg Basin. On lithological and structural grounds Cheney & Twist (1986) regarded the lower part of the Swaershoek Formation as equivalent to the Loskop.

Following further work in this region Callaghan (1987) has suggested that the Alma Formation represents the base of the Waterberg. In doing so he related the Alma Formation to strike-slip movement on the Murchison fault zone.

This is the interpretation followed here. Corollaries to it expressed here are that all subsequent red beds (the Waterberg red beds) are related to strike-slip faulting; and that the depositories of all pre-Alma red beds are related to upwelling and other crustal adjustments linked to the Rooiberg Felsites and the Bushveld Complex. The time sequence of pre-Alma red beds are shown in Table 4.

Two key relationships are those between the Loskop and Wilgerivier Formations in the Middelburg Basin, and the Rooiberg Felsites and Lower and Upper Swaershoek in the Nylstroom area.

The Loskop/Wilgerivier contact has been discussed previously, and it has been argued that no significant time gap exists between these units.

MOLOPO FARMS	OTSE			NYLSTROOM	RUST DER WINTER	MIDDELBURG
				Alma (1)		
		Maladiepe		Upper Swaershoek		Wilgerivier
		Moeding	BUSHVELD GRANITE	Lower Swaershoek		
		Ditsotswane				
?	MOLOPO FARMS COMPLEX		RUSTENBURG LAYERED SUITE			
			ROOIBERG FELSITES		Rust der Winter	Loskop
						Selonsrivier

TABLE 4. The determined and inferred relationship of red beds in the Bushveld region to the Rooiberg Felsites and Bushveld Complex. (1) Refers only to Nylstroom Syncline.

The relationship of the Swaershoek Formation with the underlying Rooiberg Felsites has been commented on variously. It was considered that the Swaershoek was part of the Rooiberg (Coetzee, 1969). Jansen (1969) provided evidence of a locally irregular pre-Swaershoek topography in the Rooiberg. However, over large areas the contact between the felsite and the basal Swaershoek is a disconformity (du Plessis, 1972). Note however, that there is some confusion on this point: Coertze & Others (1977) refer to a conformable or disconformable contact; Callaghan (1987) quotes du Plessis as showing that the lower Swaershoek is conformable with the Rooiberg felsites. Overall no major break has been demonstrated.

Du Plessis (1972) has shown broad conformity of the lower Swaershoek with the Rooiberg felsites, and that this 'conformity' extends to the Bushveld Granite in structures such as the Swartkloof anticline south-west of Nylstroom. Du Pleassis considers that the granite was still pliable during the deformation of the lower Swaershoek, and that the deposition of the lower Swaershoek is penecontemporaneous with the emplacement of the Bushveld Granite.

Relevant to the above, the lower Swaershoek sediments contain no Bushveld Granite clasts, whereas the upper Swaershoek, and the overlying Alma Fromation, do. It seems likely then that the shallowly emplaced Bushveld Granite was unroofed during the overall accumulation of the Swaershoek. It is envisaged that this was by rapid vertical (dip-slip) movement on the plane of the strike-slip zone of the Murchison-Thabazimbi fault.

Crockett (1969) attributed the development of the Otse basin to differential movement between a stable platform region of felsic rocks to the west and the unstable Bushveld 'basin' into which basic and ultrabasic rocks were intruded into the upper crust to the east, such movement being along a NNE-trending hinge line through Otse. Evidence of instability in this area not only may be related to sedimentation at this time, but also to earlier events in Transvaal times and to subsequent events. If this instability does relate to the emplacement of the Rustenburg Layered Suite, then this sedimentation may predate intrusion of the Bushveld Granite.

The Rustenburg Layered Suite pre-dates the Bushveld Granite in the central Bushveld granite. There may be little age difference between these intrusions, and thus the Otse red beds may well be of comparable age to the balance of the Bushveld red beds.

WATERBERG RED BEDS

Folowing on from the previous section, the Alma Formation is regarded as the base of the Waterberg red beds (Callaghan, 1986; cf. Jansen, 1982; Callaghan, 1987).

The most prominent topographic feature in the Waterberg is the extensive Waterberg Plateau in the north, a feature which slopes to the north. The plateau is bounded by steep escarpments on which the arenites of the Group are best exposed. Around the plateau are broad flats and valleys.

In the extreme south the strata may dip steeply northwards. Elsewhere dips are usually at low angles, even flat-lying.

The lithostratigraphy of the Waterberg Group is shown in Table 5 and Figure 7 (Callaghan, 1986). Callaghan follows Jansen (1982) in noting that the lower formations show distinct differences between the north/northeast and south/southwest parts of the basin. This is considered to be owing to a topographic high which ran through the centre of that basin during its early development (Callaghan, op.cit.).

South, southwestern and central part		North, northwestern and central part
Vaalwater Formation		Vaalwater Formation
Cleremont Formation		Cleremont Formation
Sandriviersberg Formation		Mogolakwena Formation
Aasvoëlkop Formation	CENTRAL	Makgabeng Formation
Skilpadkop Formation	RISE	Setlaole Formation
Alma Formation		
Swaershoek Formation		

TABLE 5. Stratigraphy of the Waterberg Group (Callaghan, 1986)

Formation	Lithology	Thickness (m)	Suggested Environment	Palaeocurrents	Notes
VAALWATER	Fine-grained felspathic arenites and arenaceous lutites	475 (maximum)	shallow sea or littoral? Possibly estuarine?		Colour mainly greyish
CLEREMONT	Medium to coarse-grained quartz-arenites	125	littoral?	NW, W, SW	Well rounded quartz, sericitic matrix
MOGALAKWENA	Medium to coarse-grained arenites and rudites	1250 - 1500; thins to north	Proximal deposits of large, braided river system	255°	Boundary between formation is arbitrary - now considered to be a single unit
SANDRIVIERSBERGE	Medium to coarse-grained arenites, some rudites	1250	More distal facies of braided river system	244°	
MAKGABENG	Fine to medium-grained arenites	380 (maximum)	Aeolian: either desert or coastal dunes?	Varied	Very large wedge-shaped planar cross-bedding
AASVOËLKOP		600 (usually thinner)	Shallow inland lake	Very varied, overall from north	
SETLAOLE	Arenites and rudaceous arenites	450	Proximal, narrow braid-plain deposit		Base of basin in northeast
SKILPADKOP	Arenites, and rudites	450 - 600	Narrow fluvial braidplain	Predominantly from north	
ALMA	Greywacke-type varied sizes	3000 thins rapidly	Varied alluvial fan deposits	Facies indicate transport north-eastwards	Also present in Nylstroom Syncline

Table 6. Lithostratigraphic units in the Waterberg red beds

Not shown - Blouberg and Koedoesrand Formations

The relationship of the lithostratigraphic units is as follows:

The Alma Formation conformably to unconformably overlies the Swaershoek Formation. The Sterkrivier Formation appears to grade laterally into the Alma Formation as well as the upper Swaershoek Formation. The Skilpadkop Formation overlies the previously mentioned units and is a stratigraphic equivalent of the Setlaole Formation in the north and northeast. These are succeeded by the Makgabeng Formation in the north and northeast and by the Aasvoëlkop Formation in the south and southwest. These formations are overlain by the Mogalakwena Formation in the northeast and by the Sandriviersberg formation in the southwest, which are themselves gradational. These in turn grade upwards into the Cleremont Formation. The Waterberg Group is terminated by the Vaalwater Formation which gradationally overlies the Cleremont Formation (Callaghan, 1986).

Generalised information on the units is shown in Table 6. Some more specific information on individual units is provided later.

The processes of red bed pigmentation have been little studied in any of the early Proterozoic red beds under consideration. However some examination has been made of the immature fluvial sediments of the Waterberg Group by Eriksson & Vos (1979), who concluded that authigenic haematite formed through the diagenetic alteration of iron-bearing detrital minerals.

In the Waterberg sandstones monocrystalline quartz is the dominant constituent. Rock fragments comprise felsite, white quartzite, chert and siltstone. Felspars are mainly orthoclase with less abundant microcline and albite. The felsite fragments and orthoclase grains are characteristically hydrated to clay minerals. This weathering led to the liberation of iron from felsite and orthoclase; this iron now occurs as haematite and limonite specks within the grains. Heavy minerals include ilmenite, sphene and pyroxene, which are in various stages of breakdown to leucoxene. Magnetite occurs at the base of many cross-bedded foresets. The matrix of these sandstones consists of opaque haematite and limonite, chert, red specular haematite laths and needles, clay minerals and authigenic quartz overgrowths.

Petrographic examination showed that iron specks and veins are evident within orthoclase and pyroxene grains, and felsite rock fragments. Magnetite is one source of iron in red beds, but in the case of the Waterberg this mineral is stable.

Iron released from silicate minerals by hydrolysis was probably transported in mildly acidic groundwaters prior to deposition on grain surfaces and along grain fractures at elevated Eh and pH, or increased concentrations of iron in solution. The precipitation was probably as limonite which subsequently aged to haematite. Eriksson & Vos (1979) note that the age of the red colouration cannot be determined accurately but probably developed at an early stage of diagenesis. The presence of authigenic quartz outgrowths postdating the pigmentation is evidence in support of this; in the complexity of the diagenetic process at least a sequence is indicated by these observations.

At several localities, examination of thin sections shows extensive alteration of ilmenite and the development of rims of anatase. The formation of the anatase, which may represent a cement, is regarded as diagenetic (Callaghan, 1987).

The Waterberg are typically red beds: the most common hue is 5R (mid red); with 10R (orange-red), 5YR (mid-orange) and 10YR (yellowish orange) making up the bulk of the colours recorded (Callaghan, 1987). However, greyish coloured rocks do occur in the Cleremont, even more so in the Vaalwater Formation.

Alma Formation

The Alma Formation overlies the 'upper' Swaershoek and is confined to a linear belt immediately north of the Thabazimbi-Murchison lineament, the so-called Alma 'trough', and to a central part of the Nylstroom Syncline. In the Alma 'trough' a thickness of 3000 m near Alma decreases rapidly in all directions. The view taken here is that the Alma Formation in the Nylstroom Syncline is coeval, but formed in a different setting.

The sediments are characteristically arkosic but grade eastwards into the non-felspathic Sterkfontein Formation.

Four facies are observed in the Formation (Callaghan, 1986): a poorly sorted rudaceous arenite with some trough cross-bedding; a subangular to angular rudite facies; a planar bedded facies, with isolated nests of cobbles and pebbles and trough cross-bedding in some places; and a lutaceous facies. These facies are interpreted as forming in streamflow, sieve, sheetflood and proximal lake deposition in a series of alluvial fans, forming a bajada or alluvial slope along a scarp on the southern side of the Murchison strike-slip fault zone. The facies are distributed towards the northeast in these fans. The trough cross-bedding in the Formation is too poorly exposed to provide statistically viable measurements.

The rudite facies is especially coarse in the Gatkop area where the Buffelshoek Boulder Conglomerate member is developed; this contains much granitic material from the adjacent Bushveld Granite - it is at times difficult to distinguish where weathered Bushveld boulders end and Alma debris begins.

Skilpadkop Formation

Overlying the Alma Formation conformably to unconformably in the south of the basin are arenites and rudites of the Skilpadkop Formation.

These sediments are thickly bedded, non-arkosic, lithic arenites and rudites. The rudites are usually pebbly, although some cobbles and boulders occur. The Formation wedges out eastwards from a thickness of 600 m.

The arenites are either massive or display trough cross-bedding. Varied palaeocurrent directions are suggested including from the northeast and west, but the measurements are sparse. Some golf-ball-size reduction spots occur in the arenites.

The sediments are thought to have accumulated on a narrow, fluvial braid plain.

Setlaole Formation

The base of the Waterberg in the main (northeastern) part of the basin (as defined) is represented by the restricted outcrops of the Setlaole Formation. The Formation grades into the overlying Makgabeng Formation.

The arenites and rudaceous arenites of this Formation are coarse-grained, arkosic to sub-arkosic in composition. A distinctive feature is the sporadic development of tuff and tuffaceous mudstone near the base. The Ngwepe Tuff is 7 m thick and has been traced for 3 kilometres.

The Setlaole Formation occupies the same stratigraphic position as the Skilpadkop Formation to the south. Like the Skilpadkop, the Setlaole Formation is interpreted as a proximal narrow braidplain deposit. Despite the apparent stratigraphic equivalence of these two formations, there is no evidence that they were ever continuous, nor is the direction of transport for this unit established clearly.

Aasvoëlkop Formation

The Aasvoëlkop Formation is an upward-coarsening sequence which occurs extensively in the west - this formation represents the total Waterberg at the Botswana border - forms a narrow strip in the south and interfingers with the aeolian deposits of the Makgabeng Formation in the northeast. It overlies the Skilpadkop Formation in the south, and, where it is developed, in the west. It is conformably overlain by the Sandriviersberg Formation. The maximum thickness developed is 600 m, but usually this unit is much thinner.

The Formation coarsens upwards from the Groothoek Mudstone member to arenites and rudaceous arenites. The arenites are typically fine-grained and immature. These beds may contain planar or trough cross-bedding; limited palaeocurrent data indicates varied directions of transport, but with an overall northern vector.

The Formation is interpreted as a shallow inland-lake. The lack of chemical sedimentation suggests that the lake is of a through-flow type (Callaghan, 1987).

Makgabeng Formation

The Makgabeng Sandstone Formation occurs in the eastern and northern area of the Basin. It conformably overlies the Setlaole and Skilpadkop Formations, and is in the main overlain conformably by the Mogalakwena-Sandriviersberg Formation, but the Formation wedges out toward the north. To the south and west, the Makgabeng grades into or interfingers with the Aasvoëlkop Formation. Callaghan (1987) considers that the maximum thickness is less than earlier estimates, and notes that in the type area, the only area where outcrop is good enough for reasonable measurement, the maximum thickness is (only) 380 m.

The Formation, which is not well-exposed, is characterised by uniform pale, greyish-red to very pale-red, fine to medium-grained arenites, which are usually lacking in matrix and display very large-scale planar cross-bedding.

This cross-bedding is best seen on the Makgabeng Plateau. The cross-beds are planar wedge-shaped; individual cross-beds are 15-50 cms thick, internally laminated and cross-bed sets are often 8 m or more in height. Other sedimentary structures are symmetrical and asymmetrical ripples, interference ripples, channels, current lineation, dessication cracks and problematical traces.

Meinster and Tickell (1975) regarded the formation as an aeolian deposit and this has been accepted by subsequent workers. There is insufficient evidence to suggest a more precise environment, and proposals have included longitudinal (Callaghan, 1987), transverse (Jansen, 1982), and coastal dune fields (Callaghan, 1986).

The Sandriviersberge and Mogalakwena Formations

These formations both consist of trough or, less commonly, planar cross-bedded medium to coarse-grained arenites and rudites. They contain frequent pebble washes. The boundary between them is an arbitrary one, and the two formations are considered to be one, the Sandriviersberge (Callaghan, 1987). The sediment becomes coarser northwards.

Coarse rudites are common towards the north, especially the northeast: boulder conglomerates are known. Three rudite zones are named members: the Tafelkop, Marken and Sessalong Conglomerate Members. The size of coarse fragments increase northwards.

The underlying Makgabeng Formation wedges out northwards, and as a result, the Mogalakwena facies unconformably overlies pre-Waterberg rocks in part.

The Sandriviersberg facies has a constant thickness of 1250 m; for much of its extent the Mogalakwena facies is 1250-1500 m thick, although it does thin to the north.

Cross-bedding is common in these rocks, in particular trough cross-bedding, but also planar cross-bedding. Average palaeocurrent directions for the two facies are 255° and 244° respectively. The Mogalakwena represents more proximal deposition in an overall large, braided river system (Tickell, 1975) flowing parallel to the fault system, that is, west-southwest. It must be assumed that a system of this size was moving towards an ocean, but there is no evidence of the nature of this marine setting.

It has been suggested that a part of the Volop Group in the northern Cape may represent a distal facies of the Sandriviersberge (Jansen, 1983). Although there is no evidence to support this suggestion at the present time, the uniform thickness of this unit, other than against its northern margin, does indicate that the Sandriviersberge is an erosional remnant of a very much larger unit.

Cleremont Formation

The Cleremont Sandstone Formation occupies the central part of the Waterberg Plateau. It is not well exposed, but the upper and lower contacts are gradational. It has a consistent thickness of 125 m.

The sandstone is distinctive: relatively light in colour - light to dark greyish mid-red, a medium to coarse-grained quartz arenite with rounded to well-rounded quartz grains. Rounded pebble layers occur uncommonly.

The sandstone is composed of quartz grains with a sericitic matrix. Extensive cementation and the formation of authigenic quartz overgrowths is common: the boundaries of the original grains being marked by a veneer of micaceous matrix which commonly contains finely disseminated iron oxides. The rock weathers distinctively to produce a pure white sand.

The sandstones may contain both trough and planar cross-bedding at various scales. The trough cross-bedding indicates sedimentary transportation towards the northwest, west and southwest. A distinctive feature is that the cross-bedding is commonly deformed by soft-sediment slumping. Horizontal lamination is also developed.

The sandstone of this unit obviously accumulated in a different setting to the underlying fluvial sediments. Jansen (1982) considered it either to represent fluvial sands re-worked in an inland basin or to have formed as littoral sand bars which may have graded into sand-dunes. Callaghan (1987) favours a littoral environment of deposition.

In places phosphate clasts occur in the arenites. Callaghan (1987) notes their angularity, with the implication of little transport and considers the phosphates to provide additional evidence of a marine environment. It is possible that the phosphates could, alternatively, come from a Phalaborwa-type (magmatic) source.

Vaalwater Formation

The Vaalwater Formation is exposed in scattered outcrops, mainly in stream beds in the centre of the basin; the stratotype is composite. The lower contact with the Cleremont Formation is conformable, but seldom exposed. The upper contact is erosional.

The formation consists of fine. felspathic arenites and arenaceous lutites. Commonly, the rocks are horizontally laminated or massive, although trough cross-bedding and ripple-marks occur. Convulute lamination and deformed cross-bedding also occur (de Vries, 1973).

The environment of deposition is uncertain. Callaghan (1987) suggests that it may represent a shallow sea or littoral environment, or even an estuary. The generally greyish colour of these sediments, and their nature, indicates that these rocks had at least a different diagenetic history to much of the red-bed Waterberg.

Blouberg and Koedoesrand Formations

Although the SACS (1980) classification shows the Blouberg as part of the Waterberg Group, it is now widely considered to pre-date the Waterberg (Callaghan, 1987). In the Blouberg area, occupying a pivotal position between the Zoutpansberg to the ENE, and the Waterberg to the south, a number of lithologies have been distinguished as members of the Blouberg Formation (Jansen, 1976). These are:

. My Darling Trachyandesite	Trachyandesite with interbedded argillaceous and arenaceous rocks
. Semaoko Grit	Grit, partly feldspathic with interbedded argillaceous and conglomeratic rocks
. Varedig Sandstone	Sandstone
. Mositone Conglomerate	Conglomerate grit, conglomerate
. Mmallebogos Grit	Grit with interbedded argillaceous rocks
. Thalalane Felspathic Sandstone	Arkose, feldspathic sandstone, sandstone with interbedded argillaceous rocks
. Manaka Arkose	Conglomeratic arkose with interbedded argillaceous rocks
. Basehla Arkose Breccia	Arkosic grit, conglomerate and breccia with argillaceous rocks at base

No succession can be established. Lithologies on the southern slopes of Blouberg are predominantly felspathic, on the northern slopes predominantly non-felspathic and partly volcanic, whilst around Lebu Mountain the rocks are mainly volcanic.

The major development of the My Darling trachyandesite member accumulated in the 13 x 13 km Lebu trough which opens eastwards. The lavas are between 500 and 800 m thick, altered and include trachytes, trachyandesites and andesites.

The isolated patches of this formation unconformably overlying basement are often folded, and overturned in places. It has been suggested that this 'incoherently' deformed succession is not related to block faulting, but rather to a compressional phase involving thrusting (Meinster, 1977). No evidence has yet been found to adduce horizontal movement (wrench faulting) but further structural analysis appears warranted.

A galena-baryte vein occurrence is known immediately south of the Blouberg, and was worked in the period about the Second World War.

The Blouberg is adjacent to the Waterberg Basin in the Makgabeng Mountains to the south; and to the Soutpansberg to the ENE. In relation to these sequences the following clear stratigraphic relations have been established (Jansen, 1976):

- the Wylliespoort Formation of the Soutpansberg Group overlies some members of the Blouberg Formation (and also Basement)

- the Sesalong Boulder Conglomerate Member, the basal unit of the Mogalakwena Conglomerate Formation of the Waterberg Group unconformably overlies some members of the Blouberg Formation (and locally also the Basement).

The Koedoesrand Formation includes epidotised (regarded as andesites) basic lavas and red beds: schistose conglomerates, grit, quartzitic sandstones, and quartz-sericite schists/phyllites (Visser, 1953). The thickness has been estimated at 1600 m. Adjacent to the Melinda Fault the Palala granite intrudes the Koedoesrand Formation (Van der Walt, 1978). Recent work has distinguished two phases: a coarse-grained hornblende-bearing grey to pinkish granite, and a coarse-grained variety containing minor amounts of mafic minerals, with a distinctive red colour. Migmatitic gneiss is now also known from this area.

The Palala granite intrudes mylonitic rocks along its southern contact, and is also affected by later mylonitization, especially along the northern contact. This granite is stanniferous, and contains Zn, Ba and fluorite.

As indicated, earlier age determinations of 1770 m.y. are now thought to be incorrectly based on composite samples, and a single zircon has now been dated at 1972 $\pm$ 62 m.y. This date approaches a Bushveld age (Brandl, 1986).

Economic Deposits

No mineral deposits are currently worked in the Waterberg.

Given below is information on known occurrences of minerals of economic interest. These relatively insignificant occurrences include alluvial heavy mineral placers and manganese deposits. There are other examples of diagenetic, possible weathering and hydrothermal occurrences. Specific mineralisation is known from the Gatkop area; this can be related to fault movement on the Thabazimbi-Murchison lineament. Information on occurrences is provided in Hammerbeck and Taljaard, 1976.

The Waterberg contains many small occurrences of manganese (Hammerbeck & Taljaard, 1976). The deposits occur as gash-vein fillings and shear zone impregnations, and as localised concentrations in the sediments. Callaghan (1987) records an occurrence in the Skilpadkop Formation (Klipspruit 457 KQ) with up to 26.5 percent MnO, chiefly as cryptomelane, in association with Ba, Cu, Zn, W and U.

Small, low-grade heavy mineral deposits occur throughout the basin, especially in the Vaalwater, Cleremont and Sandriviersberg Formations: none are of economic importance. These occurrences are titanium-bearing, and include titanomagnetite, titanohaematite and ilmenite, in association (inter alia) with zircon, rutile, monazite, and apatite. In descriptions from the southeast of the basin, Frick (1972) records two types of ilmenite-bearing grains: haematite with thin exsolution intergrowths of ilmenite, that is a titanohaematite, and, less commonly, haematite-bearing ilmenite.

At the beginning of the century, the Waterberg conglomerates were prospected for gold, but with negative results (Jansen, 1982).

Several occurrences of epigenetic copper, lead or barium minerals, often accompanied by brecciation and quartz veining, are known in association with post-Waterberg dolerite intrusives.

Callaghan (1987) records some occurrences of phosphate clasts in arenites of the Cleremont Formation. The phosphate clasts are angular and have been little transported.

In the Gatkop area, in rhyolite or granite-rhyolite-bearing rudites of the Alma Formation, the heavy mineral deposits may also contain tin and thorium: the tin as coarse cassiterites, the thorium as thorite, and in monazite and zircon.

Anomalous uranium values are known over an area at Gatkop. The uranium is as pitchblende, with some kasolite (a lead-uranium silicate); it is found in association with chalcopyrite and some galena. Callaghan (1987) suggests that the uranium/copper enrichment is related to an allitisation /monosiallisation weathering process.

High zinc concentrations have also been recorded in the Gatkop area on the farm Waterval 443KQ, especially near the Murchison fault zone. The sphaleritic zinc mineralization is found in association with fluorite, fluocerite, and zircon in a clay matrix and is regarded as hydrothermal. Zinc values increase towards the fault.

TECTONIC CONTROL OF SEDIMENTATION

Since 1975, a number of observations and comments have been made on the control of the Waterberg red-bed sedimentation.

Crockett & Jones (1975) noted the relationship of the Waterberg in southeast Botswana, and in the Transvaal to directions of faulting. The sedimentation represented fault-bounded rectilinear troughs related to a number of well-defined directions of fracturing dating from Waterberg times. The commonest fault direction was ENE (Fig. 9), and the movement on the faults was considered to be vertical.

Jansen (1982) felt that the primary control of basin formation was affected by fundamental zones of crustal weakness, in particular the ENE-trending Murchison lineament. Waterberg sedimentation was attributed to 'crustal undulations engendered by lateral subcrustal flow'.

Callaghan (1986) interpreted the Alma Formation as a series of alluvial fans... 'caused by the uplifted block on the south side of the Murchison strike-slip fault system'. Callaghan considered this fault system remained active during the accumulation of the overlying Skilpadkop Formation. Further, he relates a number of features to the Melinda Fault (on the Zoetfontein lineament) to the north. These include the presence of vertical Makgabeng strata adjacent to the fault, and frequent soft-sediment slumping in upper parts of the Waterberg.

du Plessis (1987) has re-interpreted the structure at Gatkop to the east of Thabazimbi. Previously regarded as an area of northerly directed thrust-faulting, in which older Transvaal strata were carried over Waterberg red beds, du Plessis shows, on the basis of slickensides and the identification of a positive flower structure that 'faulting is related to early Waterberg basin-forming processes and that strike-slip-controlled sedimentation (on the Thabazimbi-Murchison lineament) is indicated by pebble mis-matching'.

Structures related to wrench-faulting are complex: Y-shaped shears represent the principal displacement and develop last. During transpressive and transtensive wrench-faulting reverse and normal faults respectively flank the Y-shear symmetrically. It is these resulting structures that are referred to as 'flower structures'.

Horizontal and oblique slickenfibres on the faults bounding the strike-slip duplexes in the fault-zone at Gatkop indicates strike-slip faulting with upward movement of the duplexes, that is, the faults are transpressive and are reflected in positive flower structures.

Such a structure, from the Bobbejaanwater Quarry, 4 km south of Thabazimbi is illustrated in Fig. 10. Although the units in this structure are from the underlying Transvaal Sequence, the difficulty of interpreting such structures vis-à-vis thrust faulting is readily apparent.

On the northeast and western slope of Gatkop the felspathic grit of the Alma Formation grades into a boulder conglomerate. This rock type features prominently in the interpretation of the tectonics during deposition of this formation (du Plessis, 1987).

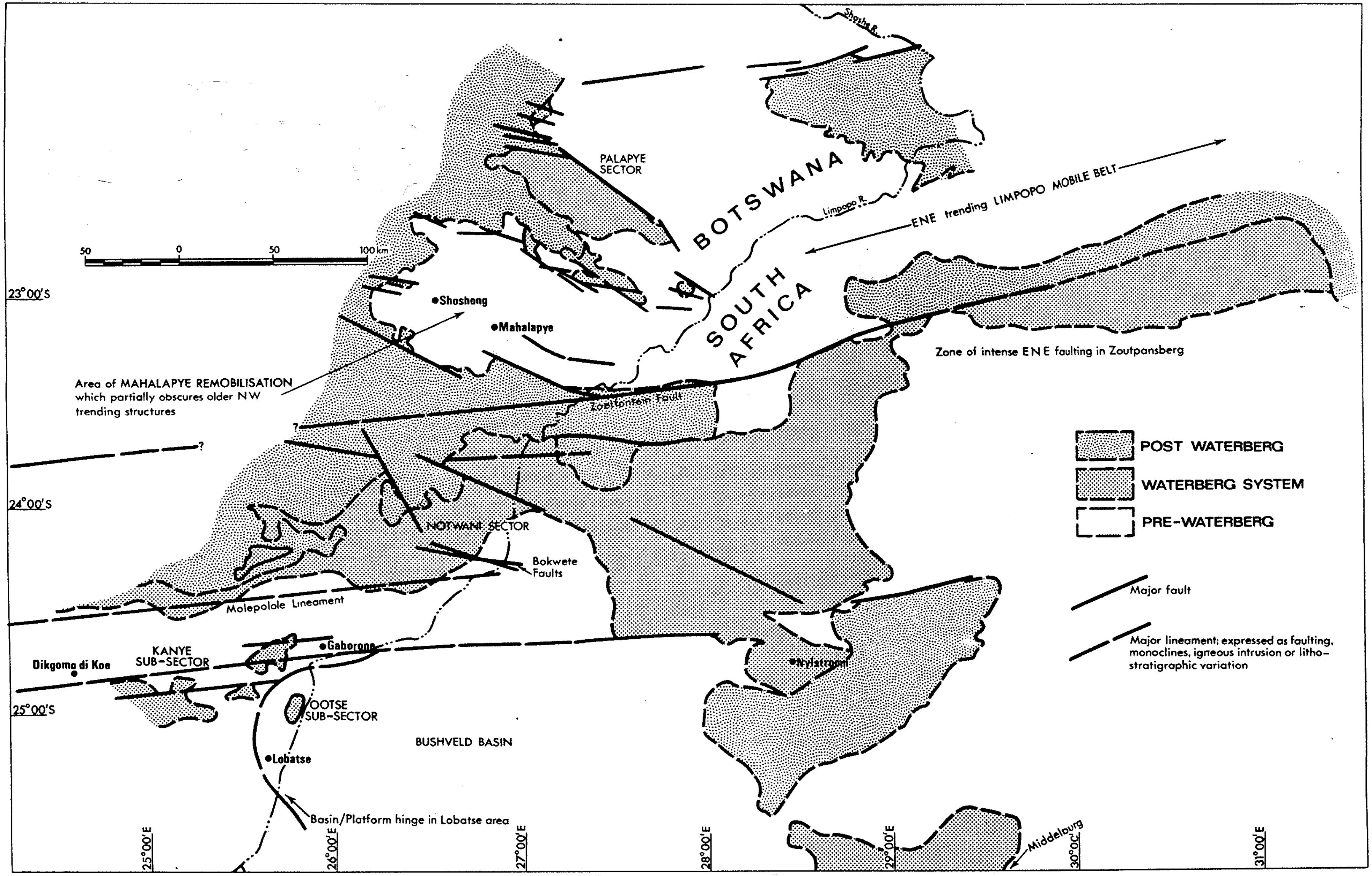

Fig. 9. Major structures in the northern Transvaal and southeast Botswana (Crockett & Jones, 1975)
Reproduced with permission of Geological Society of South Africa.

Several features of the conglomerate point to its deposition and provenance being controlled by a fault-scarp to the south:

1. The size of the coarse fragments increases from pebble to cobble southwards;
2. granite pebbles and cobbles are restricted to the vicinity directly adjacent to this fault;
3. all members of the Alma Formation grade into and interfinger with the boulder beds near Gatkop (Fig. 11). The fact that the boulder beds are stacked, interfinger with all horizons, and are not transgressive, indicate the close association with a fault scarp;
4. the thick accumulation of reddish arkose in the upper Alma Formation to the north of Gatkop is interpreted as a wedge arkose deposit related to a high angle fault scarp;
5. a sliver of Bushveld granite is present in the fault zone.

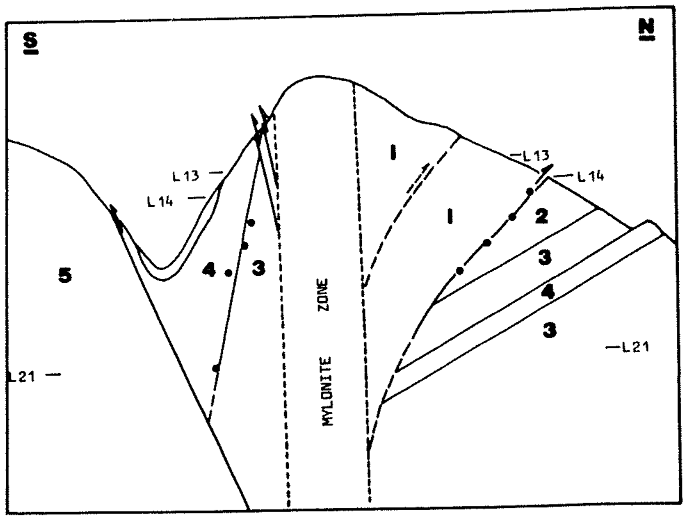

Fig. 10. Diagrammatic view of the west face of the Bobbejaanwater Quarry. Penge Formation ironstone (1) is upthrust northwards onto Hekpoort Andesite Formation (2), while Pretoria Group shale (3) and quartzite (4) is upthrust southwards onto Chuniespoort Group dolomite(5). Steep reverse faults are exposed on level 13 (L 13). Control points on the planes are indicated by large dots: these planes shown in their true attitudes, rest of diagram not to scale. (du Plessis & Clendenin, 1987)
Reproduced with permission of the Geological Society of South Africa.

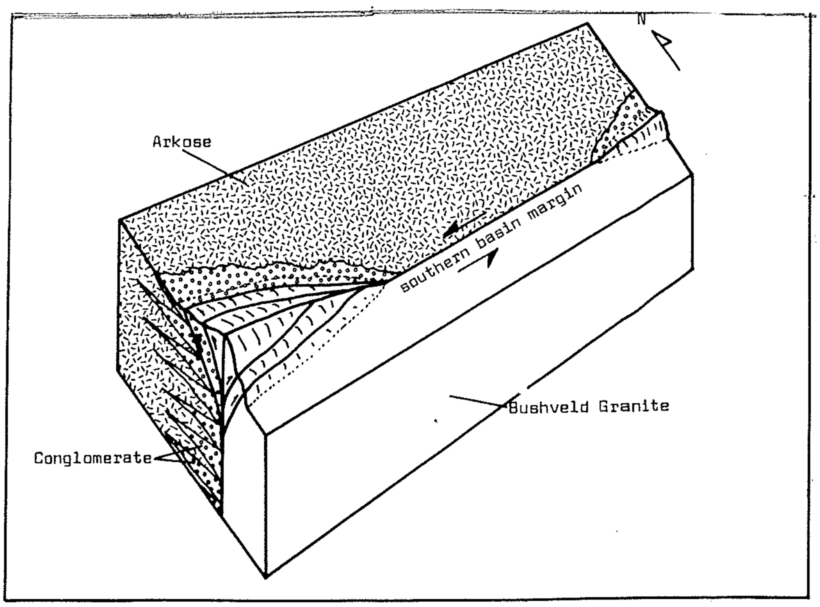

Fig. 11. Relation of facies in the Alma Formation to the Murchison fault zone (du Plessis, 1987)
Reproduced with permission of the Geological Society of South Africa.

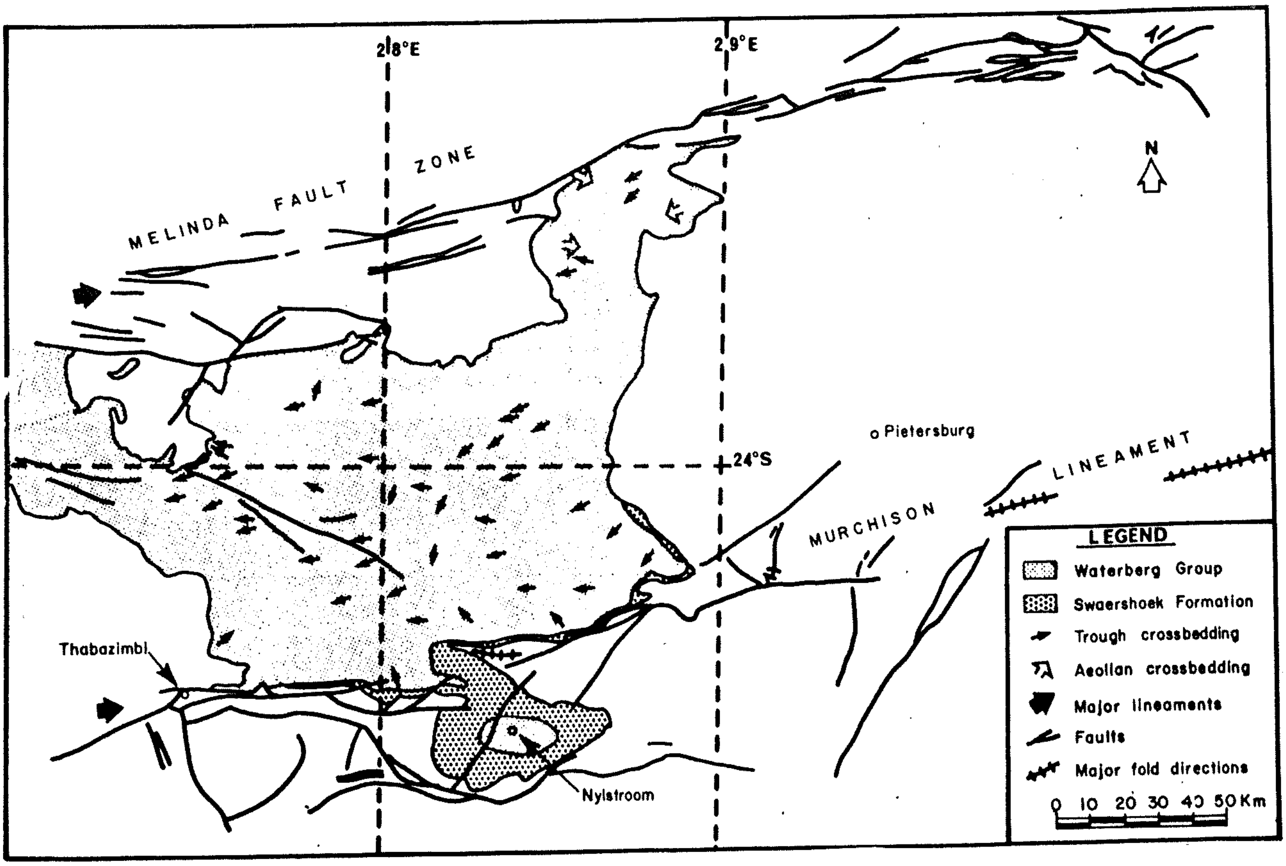

Fig. 12. Trough and aeolian cross-bedding palaeocurrent directions in the Waterberg red beds (Callaghan, 1986)
Reproduced with permission of the Geological Society of South Africa

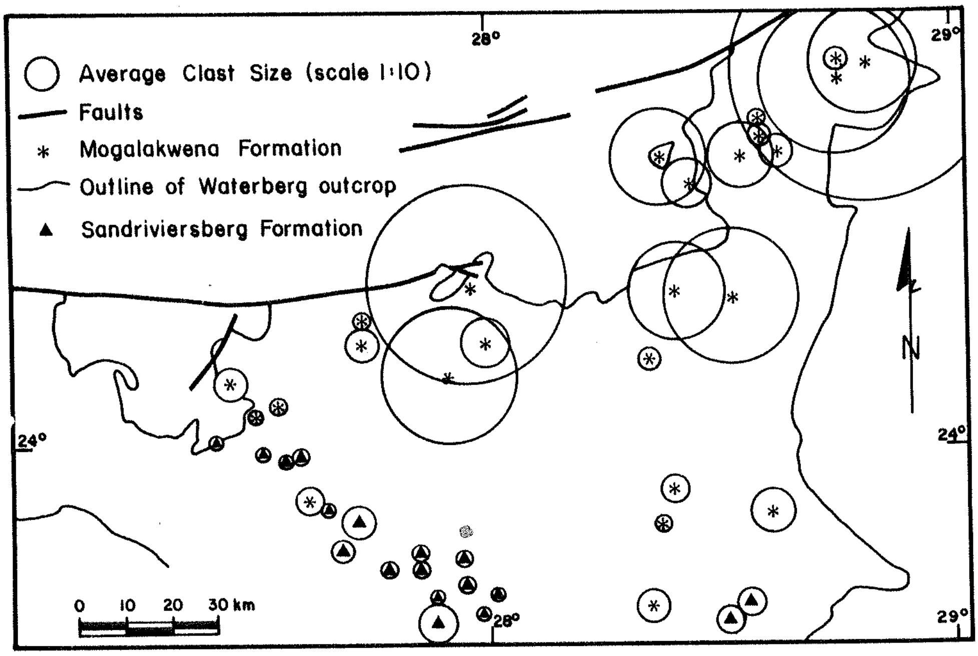

Fig. 13. Clast-size distribution in rudites of the Mogalakwena and Sandriviersberg Formations (Callaghan, 1987)
Reproduced with permission of C. C. Callaghan.

Current workers in the Limpopo Metamorphic Belt (see for example Van Reenen & Others, 1987), consider that post-Bushveld movement took place on major shears such as the Zoetfontein lineament. Transpressive strain along the Thabazimbi-Murchison strike-slip fault zone resulted in the formation of deformation lamellae in Bushveld granites near Thabazimbi (Callaghan, 1986).

The contention expressed here is that the overall Waterberg red-bed sedimentation is dominated by strike-slip faulting; related vertical movements are a part of the process. Dominant in this mechanism is movement on the Thabazimbi-Murchison and Zoetfontein lineaments: short-lived movement on the former was followed by a major side-stepping with long-standing movement following on the Zoetfontein fault to the north. Undoubtedly relating the Waterberg red beds to these two structures is a simplification: Crockett & Jones (1975) for example note that in Botswana the Molepolole lineament separates the Notwani and Kanye outcrop sub-sectors; and secondary fault directions are anticipated - and certainly occur, especially in Botswana (Fig. 9).

The difficulties in distinguishing this type of sedimentary setting have been outlined earlier. The evidence in the alluvial fan deposits of the Alma Formation at Gatkop appears unequivocal. Other relevant features include

- other than in the Alma Formation, and to some extent the Skilpadkop Formation, there is no clear evidence of north-directed sedimentation
- trough cross-bedding measurements (Fig. 12; Callaghan, 1986) essentially parallel the lineaments
- rudite fragments increase in size towards the Zoetfontein lineament (Fig. 13; Callaghan, 1986).
- very limited vulcanicity
- little or no regional metamorphism.

The Blouberg is a zone of structural complexity lying on the Zoetfontein fault; it is (in toto) not well exposed, nor fully understood. The strata include coarse clastics, and volcanics; some are folded. There is a braided fault pattern. Further structural work may show the Blouberg to be a succession of alternating zones of transpression and transtension resulting from curvature along the main strike-slip system. Its situation between the Waterberg and the opening rift of the Zoutpansberg to the east would be significant. The Koedoesrand (Formation) to the WSW may represent a similar manifestation.

Whilst much of the sediment attributed to the Waterberg Group has been transported parallel to the fundamental ENE faults, the provenance of the lithostratigraphic units is thought to have been as shown below:

NORTHERN
PROVENANCE

Zoetfontein
lineament

Blouberg

Koedoesrand

Skilpadkop/Setlaole

Aasvoëlkop (Makgabeng)

Sandriviersberg = Mogalakwena

Cleremont

Skilpadkop

Vaalwater

Alma

Thabazimbi-
Murchison lineament

SOUTHERN
PROVENANCE

THE PALAPYE RED BEDS

The Palapye Group consists mainly of clastic sedimentary rocks, frequently of red-bed type, cropping out over an area of about 3500 km² in central-eastern Botswana between latitudes 22° and 23°30'S, and longitudes 26° and 28°E (Ermanovics & Others, 1978).

Five lithostratigraphic formations have been established within the Group. The Selika, Moeng, Tswapong and Lotsane Formations form a more or less continuous outcrop in the central and northern exposures; the Shoshong Formation outcrops to the southwest as outliers on the Mahalapye Granite (Fig. 14).

The Selika, Moeng, Tswapong and Lotsane Formations show considerable lateral variation; thus informally defined members (Table 7) cannot necessarily be traced laterally for any distance (Crockett & Jones, 1975; Ermanovics & Others, 1978).

Included in the red beds of the Selika Formation is, in the east, a 120 m volcanic member of massive, dark-green amygdaloidal lavas and fissile, colour-laminated tuffs.

The Moeng Formation consists of poorly exposed shales and siltstones. The calcareous member contains several thin limestone horizons, which on occasion are oolitic.

The Tswapong Formation is a thick sequence of coarse clastic red beds. The pisolite member is a distinctive unit in which pisolitic ironstone bands are interbedded within a sequence of flaggy ferruginous quartzites. This Formation is difficult to distinguish from the Selika Formation if the overall succession cannot be established or if the intervening Moeng shales are absent.

The Lotsane Formation is extensively developed, but poorly exposed, north of Palapye. Borehole information shows that it is composed of grey, grey-blue or lilac-coloured shales, and siltstones.

The Shoshong Formation comprises a basal and discontinuously developed greywacke, succeeded by a succession of quartzites, including orthoquartzites, and ferruginous shales and a banded haematitic ironstone up to 60 m thick: above this, sandstone, siltstone, shale (locally calcareous) and local pebble conglomerates. The top of the succession is composed of impure dolomitic limestone, marble and calcareous shale.

Enormous volumes of dolerite, the Shoshong dolerite, have been intruded into the Shoshong Formation: west of 26°45' doleritic outcrop is more abundant than sediment.

The almost total geographic separation of the Shoshong Formation from the previously described Formations of the Palapye Group, and the different lithologies it contains, have, since 1950, led different authors to suggest different correlations for this unit. C. Boocock for example (in Cullen, 1961) considered, particularly having regard for the lithologies, the formation to be a correlative of the Transvaal Sequence.
However, the current field evidence from the region north of the Mokgware Hills (Skinner, 1978; Ermanovics, 1980) has been taken to suggest (sic)that the

FORMATION	MEMBER	LITHOLOGY	(Thickness)
SHOSHONG	(informal) Limestone	Dolomitic limestone and marble calcareous shale, pyrometasomatic hematite ± (?) magnetite (30 to 40 m)	
	Upper quartzite	Sandstone, siltstone, shale, quartzite, local pebble conglomerate at base (75 to 450 m)	
	Lower quartzite	Sandstone, siltstone, shale, quartzite, local basal conglomerate (25 to 350 m)	
	Greywacke	Banded ironstone, shale, feldspathic quartzite, greywacke, basal conglomerate (50 to 450 m)	
LOTSANE		Shale, siltstone, mudstone, minor sandstone and quartzite, especially in north-west (± 1 00 m)	
TSWAPONG	Grit	Quartzite, grit, siltstone	(400 m)
	Flag	Quartzite, shale	(140 m)
	Upper quartzite	Quartzite, pebble conglomerate	(140 m)
	Pisolite	Quartzite, pisolitic ironstone	(150 m)
	Lower quartzite	Quartzite, pebble and basal conglomerate	(140 m)
MOENG	Upper siltstone	Siltstone, quartzite	(27 m)
	Upper shale	Shale, siltstone	(26 m)
	Calcareous	Shale, siltstone, limestone	(50 m)
	Lower shale	Shale, siltstone	(157 m)
	Lower siltstone	Siltstone, quartzite	(75 m)
SELIKA	Upper	Quartzite, siltstone, shale	(40 m)
	Manganese	Manganiferous quartzite and conglomerate, shale	(10 m)
	Middle	Quartzite, shale, conglomerate, rare tuff bands	(700 m)
	Volcanic	Andesitic lava and tuff, shale, quartzite	(120 m)
	Lower	Quartzite, pebble/boulder and basal conglomerate	(60 m)

TABLE 7 : Lithostratigraphic subdivision of the Palapye Group

(Ermanovics & Others, 1978)

Lotsane and Shoshong Formations are equivalent in age, and that locally there is a gradational contact between the Tswapong and Shoshong Formations. It is on this basis that the Shoshong Formation is (now) regarded as an integral part of the Palapye Group.

Although no specific analysis of sedimentary environs has taken place, deposition in fluvial, littoral, lacustrine, and marine environments has been inferred; the distribution of facies suggests transportation of sediment from the north, south, west and northwest.

Localised, or low-grade iron and manganese deposits occur in the Palapye Group. Information on these is given below:

Shoshong. The Banded Iron Formation (BIF) has an average thickness of 5m, and develops grades of 25-55% Fe.
Small, high-grade lenses of manganese oxide occur in the BIF. There are small localised deposits of massive haematite and crystalline magnetite where the dolomites are intruded by dolerite; these are probably replacement bodies mobilised from Banded Iron Formations lower in the sequence.

Tswapong. Three haematite-rich horizons occur in the pisolite member. Haematite pisolites and clasts occur in a ferruginous matrix. The iron grade is 30-40% Fe. An estimated 35 million tons of this low-grade ore occurs.

Selika. The manganese member has been traced for over 50 km in each outcrop area. Psilomelane is the chief matrix component in the 5 manganiferous quartzite seams. Grades vary from 25-42 per cent Mn.

There is as yet no evidence to suggest that future discoveries would be richer or more extensive than those now known.

The Palapye sedimentation is taken to relate to vertical movement on the Lechana and Chadibe Faults (Ermanovics & Others, 1978). An ensialic graben (aulacogen) developed into which the continental debris of the Palapye Group accumulated. In the east, minor volcanism accompanied the sedimentation of the Selika Formation. Overstep took place during sedimentation so that westwards, successively younger formations rest directly on metamorphic basement. In the extreme north-west, sedimentation took place in a less confined basinal area to produce the argillites of the Lotsane Formation. At about the same time to the south the clastic and chemical sediments of the Shoshong Formation were deposited in a rim syncline surrounding the positive feature of the Mahalapye Granite (Ermanovics & Others, 1978).

The rocks of the Group are folded: the folds are broad, with a wavelenth of up to 80 km and a gentle plunge westwards. This folding indicates a late-stage compression of this region, and is a distinctive feature.

The age control on the Palapye sedimentation is imprecise.

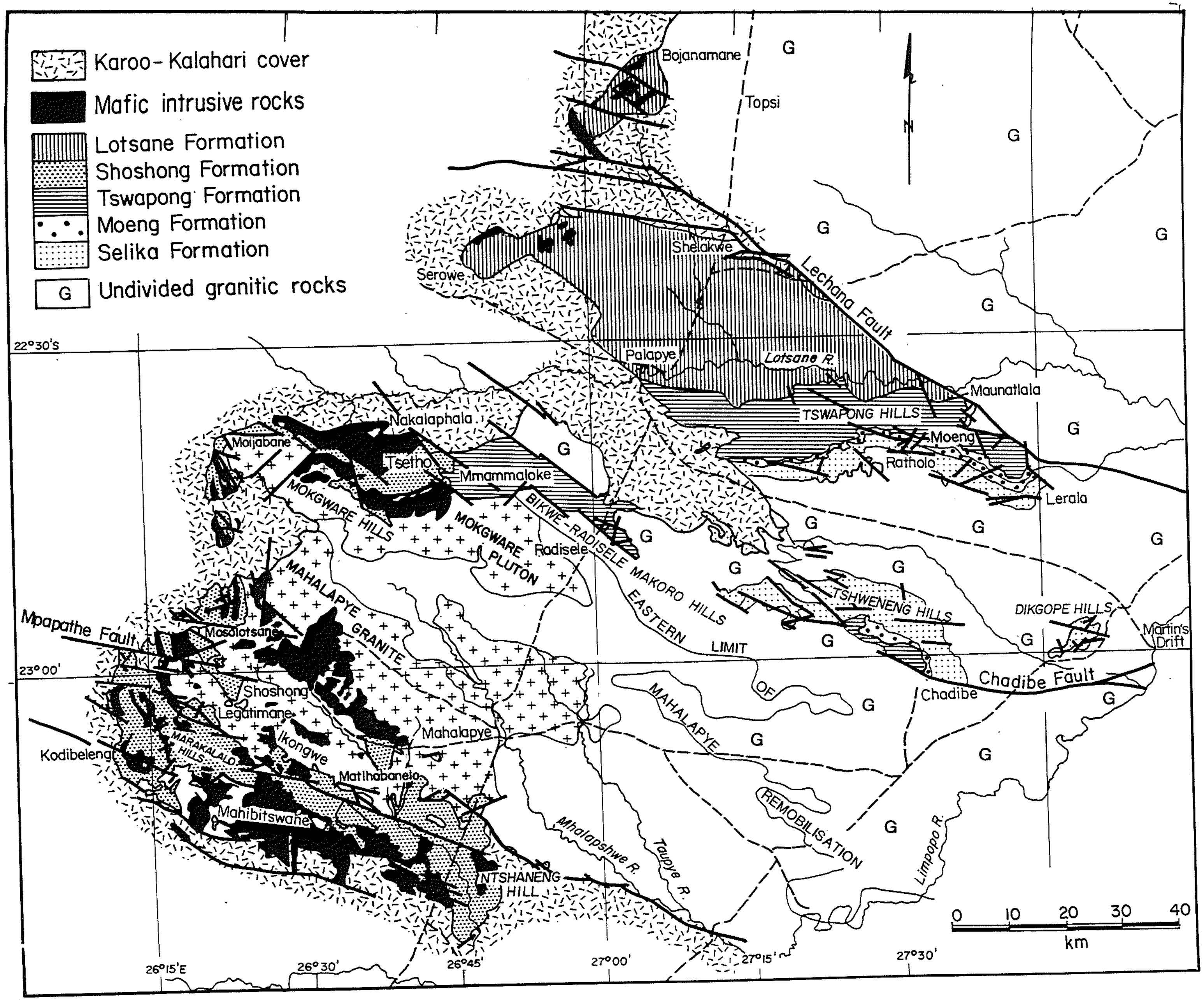

Fig. 14. The Palapye Group in southeast Botswana (Ermanovics & Others, 1978)
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The Shoshong Formation postdates the emplacement of the Mahalapye plutonic block, which represents the product of crustal mobilisation at ∿ 2200 m.y. and is not regarded as an integral part of the Limpopo Metamorphic Belt (Key & Others, 1983). The following elements are recognised in this block: the Mahalapye migmatite, the formation of which is partly shear-controlled, with dextral movement; the concordant Mokgware pluton; and the Mahalapye granite, a post-tectonic high-level sill resting above the migmatitic rocks. The Shoshong dolerite which, as indicated, cuts the Shoshong Formation, has been dated at ∿900 m.y. (dates quoted in Key & Others, 1983, attributed to C. Rundle).

The Palapye Group is considered to postdate regional uplift at ∿ 2000 m.y. (Van Breemen & Dodson, 1972) and to have accumulated at a similar time to, or coeval with, the Soutpansberg, at roughly 1800 m.y., following vertical movement (Ermanovics & Others, 1978). The age of the group may well be so, but it must be borne in mind, as has already been discussed in relation to the Soutpansberg, that the pronounced peak of Rb/Sr mineral ages in the Limpopo Belt need not be related to uplift, nor should an age of ∿1800 m.y. be accepted based on a measure of lithological correlation (and poor age data too).

The evidence that the Palapye red beds accumulated in a 'blind' aulacogen is not considered to be strong. The volcanics in the Selika Formation are relatively minor and do not of themselves provide evidence of a major rift structure. Noting the association of the Chadike Fault to the Sunnyside Shear Zone (Fig. 15), on which post-Bushveld movement is thought to have taken place (van Reenen & Others, 1987), an alternative mechanism is that, as in the case of the Waterberg red beds, the major control on sedimentation is strike-slip.

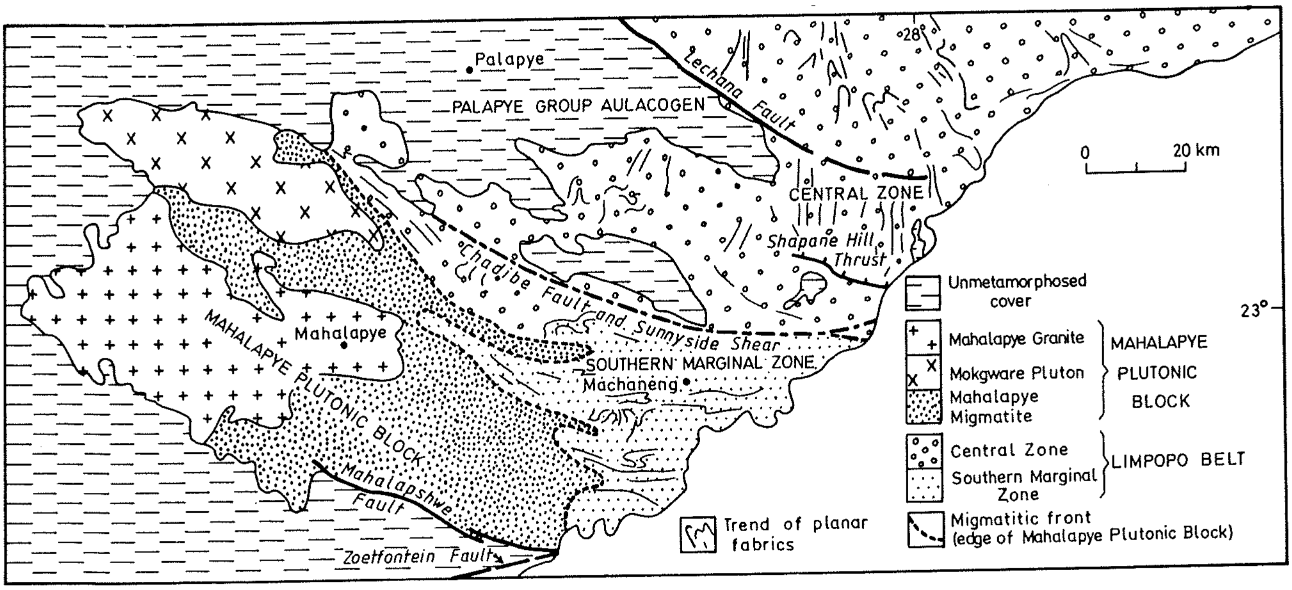

Fig. 15. The tectonic setting of the Palapye red beds (Key & Others, 1983) Reproduced with the permission of the Geological Society of South Africa.

MATSAP RED BEDS

The red beds in the Northern Cape form a major part of the Olifantshoek Sequence. This sequence is developed in a north-south strip extending northwards from near Marydale adjacent to the Orange River for some 400 kms. The following main stratigraphic units are recognised in the Olifantshoek Sequence (Fig. 16):

. Groblershoop Schist Formation
  Quartz-sericite schist; quartzite; greenstone

. Volop Quartzite Group
  Reddish brown quartz-arenite; grey quartzite

. Hartley Basalt
  Basalt, tuff

. Lucknow Quartzite Formation
  Purple and white quartz-arenite

. Mapedi Shale Formation
  Shale; quartzite; mafic lava

A correlative of the Mapedi Formation is the Gamagara Formation: this unit is exposed to the east of the main outcrop on the Maremane Dome. Unconformably underlying the Gamagara Formation in the Sishen-Postmasburg area are earlier haematite-rich, at times manganiferous beds: as is well-known these are of great economic importance.

Red beds are common, at times predominant, in the Sequence up to and including the Volop Group.

Some of the Volop Group is relatively well exposed, but in general outcrops of the Olifantshoek Sequence become sparser westwards, and northwards, as the strata disappear under a cover of Kalahari sediment. Available subsurface data has been assembled as far north as the Molopo River (Levin, 1981), and in southernmost Botswana there are outcrops in the Tsapong area and further west near the Molopo River (Boocock & Van Straaten, 1962).

The most northerly outcrops of this belt are at Jwaneng Hill in the vicinity of the Molopo Farms Complex, a sub-outcropping basic/ultrabasic body of Bushveld age (Gould & Others, 1987), which intrudes sediments and minor volcanics of the Transvaal sequence and is unconformably overlain by Waterberg rocks. The point has already been made that similar rocks also occur immediately east of the Molopo Farms Complex (but with a different, ENE strike) and thus that it is difficult to do other than correlate the Matsap red beds with the Waterberg of southern Botswana (Gould & Others, 1987). This may be so, but consideration of the contiguous Waterberg red beds - Blouberg Formation - Soutpansberg red beds in the northern Transvaal, already discussed, highlights the potential difficulties in such settings.

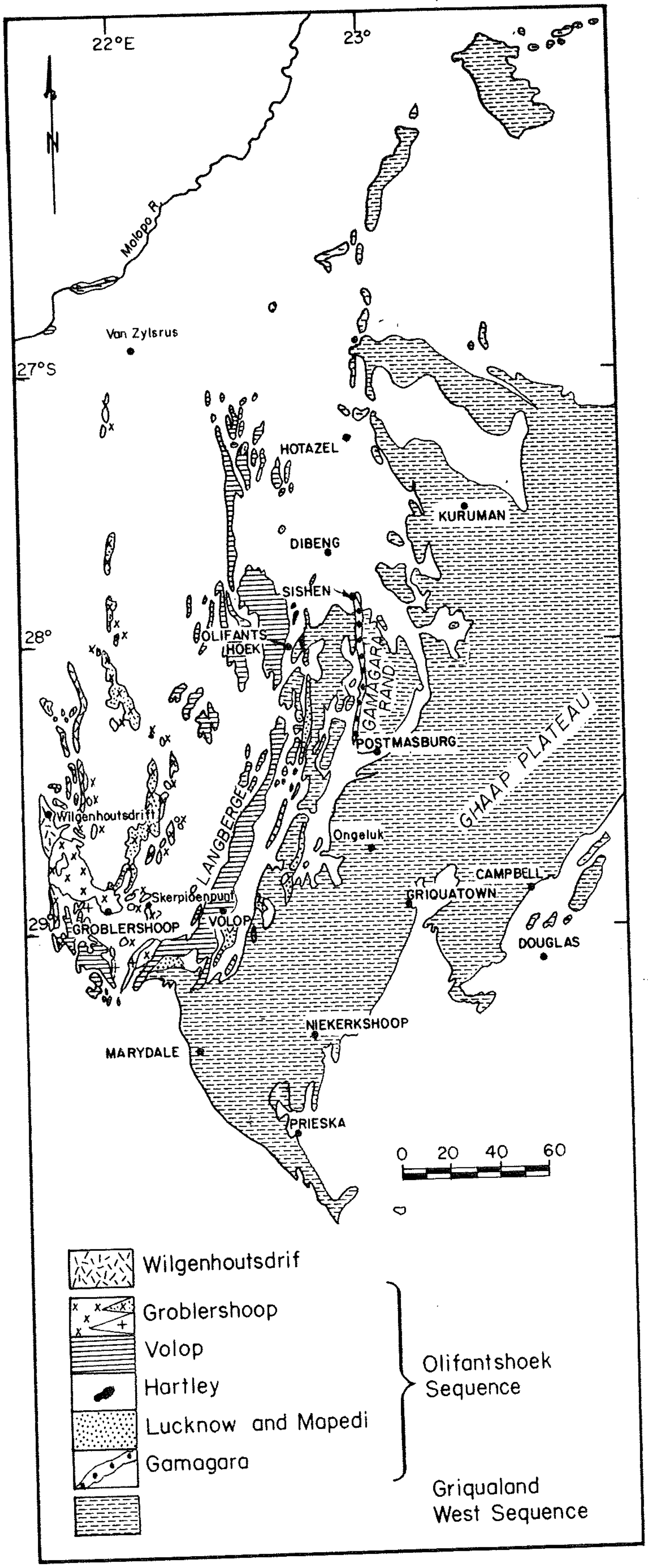

Fig. 16. The Olifantshoek sequence in the Northern Cape. Information from SACS, 1980.

Aeromagnetic trends in Botswana suggest that the Olifantshoek strata extend as far north as the westward continuation of the trace of the Zoetfontein Fault; however, the limited data available from the Kalahari Drilling Project (Meixner & Peart, 1984) has not conclusively proved this point. CKP 5A, 5A2 intersected volcanic breccia containing some quartzite clasts interpreted as Olifantshoek (Fig. 2). No holes intersected the Volop Group. The most northerly intersection of early Proterozoic sediments was at CKP 5C: siliceous shales with a dolomitised and silicified marble top are correlatives of the Griqualand West Sequence - but this is not a part of the (younger) Olifantshoek Sequence. It seems possible that the sequence cuts out, or is mostly cut out, by transverse faults south of the Zoetfontein Fault.

At the latitude of the Zoetfontein Fault, a major magnetic anomaly, the Kalahari line, distinguishes a deep magnetic basement in the west from a shallow, cratonic basement (including the Olifantshoek Sequence), to the east (Hutchins & Reeves, 1980). No Matsap, or Waterberg red beds are known north of the Zoetfontein Fault, but the Kalahari line continues northwards, as a steepening hinge zone (Pretorius, 1984).

Subsequently, the Olifantshoek Sequence has been crumpled, thrust eastwards in an orogenic phase involving east-verging folds and overthrusting. In a stratigraphic sense, the sequence represents a part of the cover of the Kaapvaal craton; in a structural sense this is a distinct fold belt, the north-south trending Kheis Belt (Vajner, 1974; Kröner and Blignault, 1976).

Broadly, but not totally, the Kheis belt is bounded on the east by relatively undisturbed Griqualand West Sequence strata. As will be outlined later, some of the Olifantshoek Sequence has, however, been thrust over underlying strata, and in southernmost exposures near Prieska, Griqualand West strata are themselves thrust eastwards.

The Kheis orogeny may or may not represent an element in the tectonic framework of the overall formation of the Early Proterozoic red beds. That it may do so provides grounds for further discussion later in this section.

The southernmost, and western margins of the Kheis Belt are further deformed by younger events (circa 1300-1000 m.y.) related to the Namaqualand Metamorphic Complex. In the foreland to this belt, specifically a part of the Upington Terrane (Stowe & Others, 1983), the Kheis structures have been refolded, turned up on end and in a final event, disrupted by dextral shearing between the Brakbosch and Brulpan Faults involving a movement of up to 140 kms. A net result of these events is an attenuation of the southern margin of the Belt (Stowe, 1984), and at least the partial obscuring of overall relationships.

There are certainly some uncertainties in the stratigraphic relations at higher levels in the Olifantshoek sequence; the existence of a number of overthrust sheets in the area, much of which is so poorly exposed, must lead to other uncertainties in the stratigraphic succession.

These uncertainties include correlations, essentially between the Groblershoop Formation and the quartzitic units of the Uitdraai, Sultanoord and Dagbreek Formations of the Upington Group in the Upington Terrane of the Namaqualand Metamorphic complex (Schlegel, 1986), and the relationship of the Groblershoop Schist Formation with the overlying volcano-sedimentary Wilgenhoutsdrif Group in the region east of Upington. Issues mentioned in this paragraph fall outside the scope of this review.

HAEMATITE-RICH BEDS BELOW THE GAMAGARA UNCONFORMITY

Below the Gamagara Formation in the Postmasburg-Sishen area, wedged unconformably between this unit and the underlying Griqualand West strata is a ferruginous chert breccia grading upwards into a distorted Manganese Iron Formation (Van Schalkwyk & Beukes, 1986). The Manganese Iron Formation, a correlative of the Asbesheuwels Iron Formation, occurs slumped into palaeo-sinkhole structures in the Campbellrand carbonates; the bulk of the Sishen iron ore is an enriched, laminated haematite ore in the Manganese Iron Formation. (The balance of this ore is conglomeratic, at the base of the overlying Gamagara Formation).

These slumped deposits represent the earliest haematitic material in the northern Cape. The development of an oxygen-rich atmosphere at about the 'Transvaal'/'Waterberg' interface has been commented on (see for example Cloud, 1976; Eriksson & Truswell, 1978). Beukes (1983) sees the erosional period in which these haematitic ores accumulated as a defined atmospheric marker (as noted earlier Twist & Cheney, 1986, consider an even more precise marker to exist within the Rooiberg Group in the northern Transvaal).

In the formation of these slumped deposits, silica was leached from the iron formations by alkaline groundwater solutions and ferrous minerals were oxidised to haematite. At the same time some iron was added to the sequence to form high-grade laminated and breccia supergene ore bodies (Beukes, 1985).

Beukes considers that the slumping, with resulting fracturing and brecciation, was an essential prerequisite for significant ferruginization: in contrast, in areas where undisturbed Asbesheuwels Iron Formation is unconformably overlain by the Gamagara Formation, ferrous minerals have been pseudomorphed by haematite, but very little leaching of silica and secondary enrichment by iron took place. He suggests that dissolution of carbonates from below the iron formation would not only have led to slumping, but also to the development of alkaline solutions ideally suited for leaching of silica from, and precipitation of, ferric hydroxides in the iron formation: silica is very soluble, ferric hydroxides very insoluble, under alkaline conditions. Ferric hydroxide may, in turn, be transformed into haematite, although the precursor to haematite platelets may also have been goethite. The transformation of goethite to platy haematite, which takes place very readily, involves a reduction in volume of about 27 per cent (Morris, 1980), resulting in a porous, high-grade haematite ore.

The large volumes of ferric compounds in the palaeosoil profile below the Gamagara indicate that the oxidation potential must have been related to a relatively high O_2 level in the atmosphere. Supergene iron ores related to dissolution of silica from, and haematization of, iron formation is well known from other parts of the world, as, for example, Minas Gerais in Brazil and the Hamersley iron ore province in Australia. More specifically, the Sishen ore deposit is very similar to the platy haematite, Whaleback-Tom Price-type deposits in Australia (Morris, 1980).

As with the iron deposits described above, there are also deposits of manganese related to the unconformity between the Campbellrand Subgroup and the Gamagara Formation on the Maremane dome. Where the Gamagara Formation rests on manganiferous dolomite of the Reivilo Formation in the centre of the dome, supergene bixbyite-rich manganese deposits are developed in the Sishen Shale Member of the Gamagara Formation - at the Glosam, Bishop and Lohatlha mines (Grobbelaar & Beukes, 1986). In this case, the manganese originally accumulated as wad on the pre-Gamagara karstic erosion surface, and was subsequently transposed to bixbyite.

OLIFANTSHOEK SEQUENCE

Mapedi Formation

At its type locality, some 20 km west of Postmasburg, the base of the Mapedi Formation is formed by a haematite pebble-conglomerate overlain by a number of upward-coarsening shale-quartzite sequences with the lowermost one only a few metres thick. The Formation is poorly exposed, contains a thick succession of amygdaloidal mafic lavas, and grades upwards through siltstones into white and purple quartzites of the overlying Lucknow Formation.

Gamagara Formation

Another red bed sequence, the Gamagara Formation, occurs further east. It occurs on the western edge of the Maremane Dome, between Sishen in the north and Postmasburg in the south. The remainder of the dome is defined by carbonate rocks of the Campbellrand Subgroup and iron formations of the Asbesheuwels Subgroup. These rocks dip gently in an arc to the north, east and south.

The Gamagara Formation strikes north-south, dips to the west and overlies the strata referred to above with an angular unconformity.

To the west of the Gamagara, older rocks of the Griqualand West Sequence have been thrust eastwards over these strata. This west-dipping thrust is the most easterly, and best exposed thrust associated with the Kheis belt.

The base of the Formation is represented by thin upward-fining cycles of the Doornfontein conglomerate member; these cycles have been interpreted as a braided alluvial fan deposit infilling surface depressions. This interpretation is based on features such as the poor sorting and rounding of conglomerate units, rapid facies varieties and the discontinuity of the conglomerate units (Van Schalkwyk & Beukes, 1986). There is an upward transition within the coarser sediments of the member to iron-formation pebble conglomerates.

The lower part of the Doornfontein member contains economic iron ores: as conglomeratic ore, consisting of partially rounded and sorted fragments of laminated and massive haematite and partially haematised iron formations, and massive ore composed of sand to silt-sized haematite grains in a matrix of platey haematite.

The remainder of the sequence is an upward-coarsening one. Red, black, and white shales of the Sishen Member are finely laminated and represent sedimentation in relatively deep and/or quiet water below wave base. The sequence shoaled upwards, as is indicated by the upward transition into trough cross-laminated, fine-grained quartzite. The upward coarsening increments of sedimentation best fit progradational delta lobe deposits. In such a delta model, the shales represent basinal muds with lenticular and wavy bedded shale/siltstone and shale/quartzite units as pro-delta deposits; and flaser-bedded quartzites and cross-bedded and laminated quartzites are interpreted as delta-front sands (Van Schalkwyk & Beukes, 1986). The upper delta-front sand unit, represented by the Marthaspoort quartzite, is better sorted than the lower ferruginous quartzite unit, possibly indicating more effective reworking by wave and/or tidal currents.

No palaeocurrent analyses are available in the Gamagara, so that the sediment dispersal patterns remain unknown. The deltas, however, were probably dominated by fluvial influences, as is indicated by the scour-based, upward-fining, coarse channel sand deposits in the upper part of the Marthaspoort quartzite, and sedimentation is assumed to have been from the east (Fig. 17).

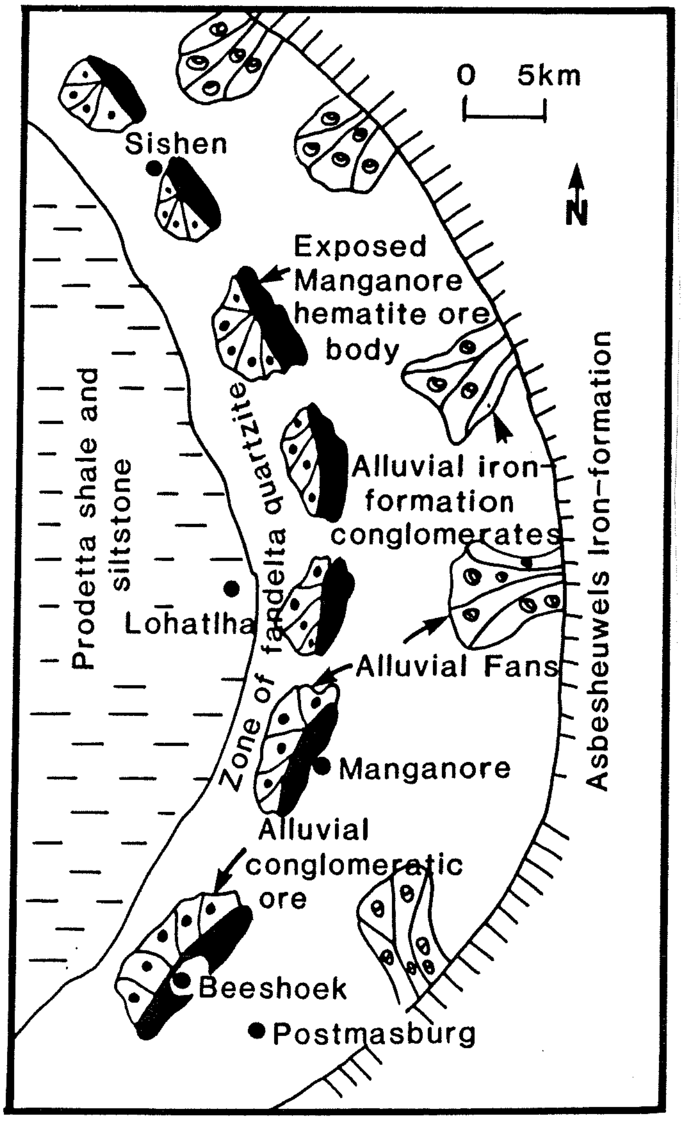

Fig. 17. The possible distribution of sedimentary environments during the early stages of deposition of the Gamagara Formation on the Maremane Dome (van Schalkwyk & Beukes, 1986)
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There are, however, distinct differences to the other red beds under discussion. Some of the red pigmentation can certainly be related to diagenetic coatings of haematite but, specifically in the Doornfontein Member, detrital haematitic material is abundant, and haematitic pisoliths also occur sporadically in the interbedded shales. Haematitic pisoliths are characteristic of lateritic soil profiles: Van Schalkwyk & Beukes (1986) consider that lateritisation may have been operative in a second-phase ferruginising of the Manganese Iron Formation, and the Gamagara ores.

A composite section of the Gamagara Formation would be of the order of 350 m thick, but actual thicknesses are much less: the Doornfontein conglomerate member is lenticular and impersistent; the thickness and lithology of the Sishen shale is also variable, and the unit thins or pinches out over palaeohighs along the unconformity with underlying Griqualand West strata; and the Paling Shale Member is commonly truncated by mylonite along the thrust fault.

Prior to 1967, it was generally accepted that the Gamagara Formation could be correlated with the Olifantshoek Sequence, and that Griqualand West Sequence strata were thrust over it from the west (see for example Strauss, 1964).

Thereafter, it was suggested (Wessels, 1967; De Villiers, 1967) that no thrusting had taken place across the Maremane Dome and that the Gamagara formed part of the (underlying) Griqualand West Sequence. In this model, the relationship between the Gamagara Formation and the overlying Griqualand West Sequence involved a number of erosional unconformities. With this interpretation, the Gamagara would represent the oldest red beds in southern Africa, probably as old as any significant red bed sequences known anywhere.

On the basis of determined major stratigraphic relations and structural dislocations, Beukes & Smit (1987) have re-established that the Gamagara Formation has been thrust over a part of theGriqualand West Sequence. The Gamagara Formation is now regarded as a small, lower part of the equivalent Mapedi Formation.

Hartley Basalt

The Hartley Basalt Formation crops out sporadically in the long, sand-filled valley east of the Langeberg. The Unit represents the middle beds or Hartley Hill Group of Rogers's (1906) Matsap Series.

The Formation is variable laterally. The following generalised sequence is based in particular on the type locality at Hartley Hill (Cornell, 1987).

Beneath the Hartley Formation lies grey quartz-arenite and brown quartz wacke. Basaltic lava flows, which occur at the base, enclose very variable thicknesses of conglomerate and agglomerate with a tuffaceous matrix. Then follows a sequence of rocks varying from white quartz-arenite through tuffaceous quartz wacke and arenaceous tuff, to lapilli tuff. An upper basalt flow follows, above which the quartz arenite component becomes dominant. This passes into the typical Fuller Formation brown quartz wacke. One or more volcanic members (of basalt and tuffaceous material) may occur higher in the sequence, i.e. in the Fuller Formation. At Pramberg to the south a quartz-felspar porphyry is developed.

Earlier definitions defined the Formation to include all volcanic material. Cornell's definition is more restrictive: he considers it to be that part of the Olifantshoek Sequence which contains more than one half by volume of volcanic material. As now defined, the formation is significantly thinner (< 300 m) than previously. Given the generally poor exposures in the area, Cornell stresses that it may be difficult to distinguish the sedimentary units above and below the basalt.

The best constrained isochron age of the basalt (Rb-Sr and Pb-Pb) is 1893 $\pm$ 48 m.y. (Armstrong, 1987). This is appreciably younger than that determined earlier - 2026 $\pm$ 180 m.y. (Crampton, 1974). It was on the basis of this older age that the Olifantshoek sequence was placed in the older Vaalian Precambrian Erathem (SACS, 1980), whereas the younger age now determined would see this sequence placed, as is the case with the Waterberg and Soutpansberg Groups in the Transvaal, in the younger Mokolian Erathem.

The Hartley volcanic rocks were erupted through fractures of the ruptured lower crust during a relatively early phase in the development of the rift in which the Olifantshoek Sequence accumulated (Cornell, 1987).

Volop Group

The Volop Group forms relatively good outcrops in the Langeberg and, to the north, in the Korannaberg. The strata are folded into a synclinorium in the Langeberg, an anticlinorium in the Korannaberg.

The total thickness of the Group is estimated at 4000 - 4500 m, but the possibility exists of some structural duplication in the sequence as a result of thrust faulting.

The unit lies above the discontinuous outcrops of the Hartley Basalt. As has been noted earlier, Cornell (1987) places minor volcanic intercalations associated with this basalt in the lowermost Formation of the Group (and in the uppermost unit below the basalt), and notes that the basalt is likely to transgress sedimentary facies, and that it need not occur consistently below the Kalahari sands.

The Volop Group is subdivided into five formations (Jansen, 1983): about half the thickness of the Group is represented by the Fuller Formation at the base, consisting of reddish brown-coloured planar and trough cross-bedded, coarse-grained quartzite. The Fuller Formation is overlain by grey-coloured, finer grained planar and trough cross-bedded quartzite of the Ellies Rus Formation. Then follows the Glen Lyon Formation which consists of reddish brown-coloured coarse-grained cross-bedded quartzite overlain by light grey medium-grained planar and trough cross-bedded quartzite of the Verwater Formation. The Top Dog Formation, which constitutes the top of the sequence, consists of white to light-grey trough and planar cross-bedded and horizontally bedded clean orthoquartzite. Contacts between all the formations are conformable and gradational; no major unconformities have been established in the Group.

The Volop Group consists of two upward-fining depositional cycles each considered to represent a retrogradational braided stream-outwash plain system. Each cycle commences with coarser grained red-bed type quartzites deposited adjacent to the source area. Erosion of the source area over a long period of time resulted in retrogradation and the deposition of finer-grained, more distal braided stream sediments over more proximal material. The lower depositional cycle is regarded as a braided stream system, whereas the upper unit was most probably deposited by large, swift flowing deep rivers that may have been similar in some respects to the present-day Brahmaputra River (Jansen, 1983). The clean orthoquartzites of the Top Dog Formation are regarded as shallow marine tidal sediments.

Palaeocurrent analysis shows one very major distinction within the Group: the sediment is derived from the north in the basal Fuller Formation, from source areas to the east in the remainder of the Group (Fig. 18).

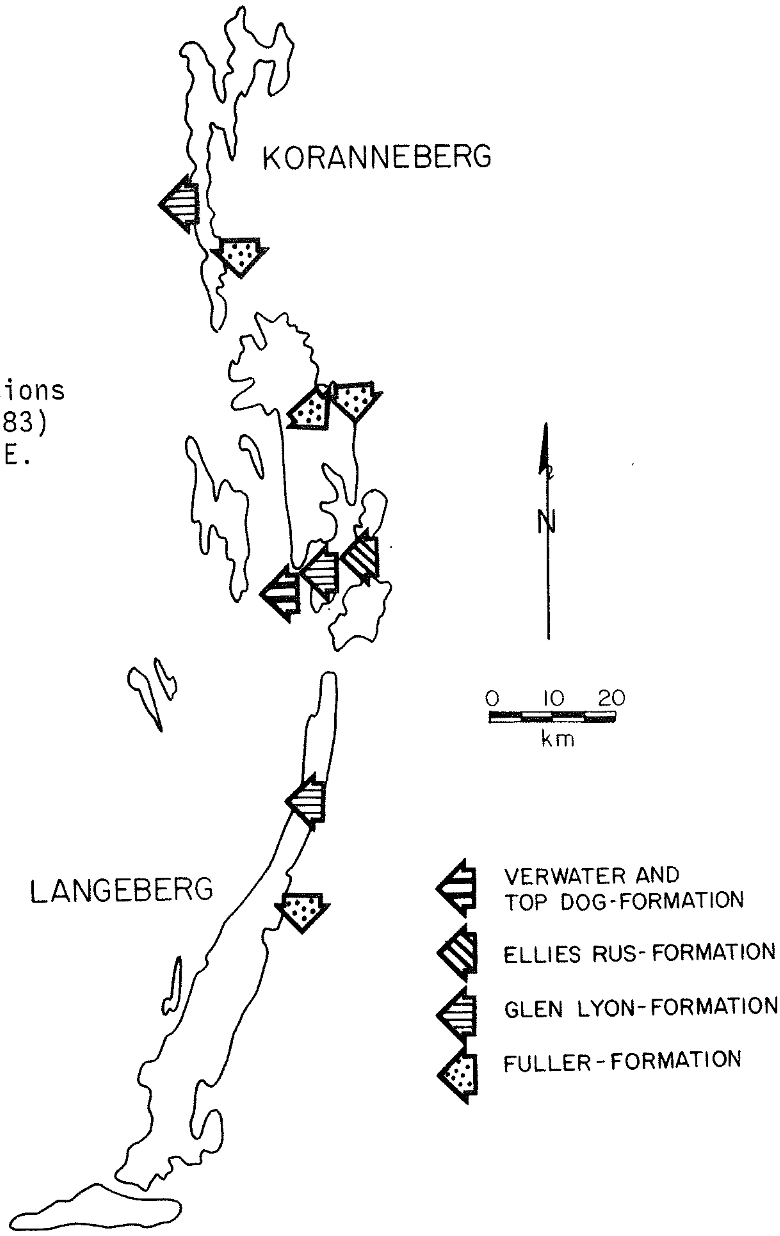

Fig. 18. Palaeocurrent directions in the Volop Group (Jansen, 1983) Reproduced with permission of E. Jansen.

Lateral facies changes within individual formations are small, the only significant ones being the predominance of horizontally bedded quartzite in the upper part of the Fuller Formation and in the southern part of the Ellies Rus Formation. The latter deposits are interpreted as sheet flow deposits on an alluvial flood plain.

No economic concentrations of minerals are known in the Group. The economic potential of the Group is considered to be very limited (Jansen, 1983). Scintillometer counts have detected traces of uranium in the Glen Lyon Formation, and very small concentrations of heavy minerals (haematite, in association with rutile and zircon) are present locally through the sequence.

Groblershoop Formation

The Groblershoop Schist Formation is composed predominantly of quartz-sericite schist. It is metamorphosed to amphibole-quartz (garnet) schist near granitic intrusions, and contains lenses and zones of mafic schists and amphibolites, and of quartzites.

Vajner (1974) regarded the contact of the Groblershoop with the Volop Group below it as tectonic, in fact on the basis of intense foliation fabrics and early isoclinal folds, he considered the Groblershoop to be older than the Volop. More recent work, however, (Botha & Others, 1976; Malherbe, 1979; Stowe, 1983) has shown that where the contact is not sheared there is a conformable sedimentary contact between the Volop and the overlying Groblershoop.

Within the predominant schists (the Skerpioenpunt Member) a number of more quartzitic units are now recognised (Schlegel, 1986).

The Boegoeberg Member is a medium to coarse-grained white to grey-white quartzite, partially orthoquartzite, containing trough cross-beds.

A second, major, distinctive quartzite is the Skeurberg Quartzite Unit, composed of orthoquartzite and interlayered micaceous quartzites and cherty nodular quartzite, with a total thickness of up to 10 m. The unit contains recumbent folds, and is the site of thrusting and folding. No sedimentary structures, or heavy minerals are seen in this unit which may represent the silicification of a previous carbonate horizon (Schlegel, 1986). This Unit overlies the Skerpioenpunt Member in the south.

In the west, this unit is underlain by the Bokpoort unit, of micaceous quartzites, semi-pelites and thin, intercalated layers of white orthoquartzite up to 30 cm thick. The quartzitic material is thinner, and a greater variety of interbedded rock types distinguishes this unit. The orthoquartzite contains some vein quartz pebbles, heavy mineral layers and low-angle cross-bedding.

The definition of the top of the Groblershoop Formation remains uncertain. Schlegel considers the purple-weathering Grootdrif Quartzite to form part of the Formation, whereas Moen (1980; 1988) correlates it with the base of the Zonderhuis Formation of an overlying Wilgenhoutdrif Group.

The contact of this quartzite with the remainder of the Groblershoop Formation is usually not exposed, but a number of features such as flaser bedding and brecciated zones suggest that it is partly or wholly of tectonic origin.

Exposures of granite are rare, but pegmatite veins in schists adjacent to the granite indicate its intrusive nature. The several masses involved are regarded as a single, tectonically dismembered batholith; the granite is thought to be syntectonic and two available ages, U-Pb ages of 1485 and 1268 from zircons (Barton & Burger, 1983) may reflect a minimum age for the first deformation event.

THE KHEIS BELT

In introducing the Olifantshoek sequence and the Kheis Fold Belt by which it is now represented, the extent of the belt was outlined. It was also noted that at the southwest margin these structures have been further deformed by younger events related to the Namaqualand Metamorphic Complex: specifically, younger Namaqualand folds refolded earlier Kheis structures.

The structures mapped in the Kheis Belt itself vary in style, intensity and orientation but are related to three main phases of deformation (Stowe, 1986). There are two phases of recumbent folding and thrusting (KF1, KF2). KF1 is developed throughout, KF2 occurs only in discrete zones. KF3, most distinctly developed in the east-central sector, is represented by upright, north-trending folds.

The region can be subdivided into overlapping, east-verging tectonic sheets by low-angle thrust (Stowe, 1986). Stowe recognises six such sheets, the uppermost of which is developed in the Zonderhuis Formation at the top of the Groblershoop Formation or the base of the Wilgenhoutsdrif Group.

Metamorphism in the Kheis Belt increases westwards; much is low-grade (Schlegel, 1986). Pelitic rocks have a phyllitic or schistose foliation that is usually defined by white mica and chlorite (Stowe, 1986). Kyanite has been recognised in the Groblershoop Formation, but it is not abundant; this may reflect pervasive retrograde sericitisation.

As has been noted the frequent red beds characterise the Olifantshoek strata up to and including the Volop Group. Above, relations are uncertain with the Groblershoop Formation. The uncertainties are even greater with the Wilgenhoutdrift Group which, at least structurally, overlies the Groblershoop Formation.

The age of the Kheis Belt is not well documented. Clearly, it postdates the Olifantshoek sediments, thus is younger than the age of the Hartley Basalt, 1893 $\pm$ 48 m.y. (assuming this age is valid). An age of metamorphism of mafic schists in the Groblershoop Formation of 1780 m.y is regarded as unreliable (Moen, 1988). If the thrust faulting in the belt does extend to the Zonderhuis Formation and this unit is part of the Wilgenhoutsdrif Group, then the thrusting is also younger than the formation and emplacement of that unit. Two U-Pb determinations on zircons from the upper rhyolite and a syenite plug intrusive into the Wilgenhoutsdrif have yielded unreliable ages of about 1300 m.y. (Barton & Burger, 1983). All that can be said is that on the present evidence, the Kheis orogeny took place between 1900 and 1300 m.y. ago.

Evidence has been led of a period of erosion below the Gamagara Formation; a presumed period of erosion between the Griqualand West and Olifantshoek sequences. However, overall the relationship between these two sequences (for which there is only evidence southwards from Olifantshoek) is one of intense deformation and movement (Coward & Potgieter, 1983).

Olifantshoek Sequence strata, not necessarily from the base of the sequence, have been thrust over or into Griqualand West strata at the southern margin of the lowermost sheet recognised by Stowe (1986) and in the Blackridge thrust, the most easterly structure recognised (Van Wyk, 1980).

The Blackridge thrust extends through the Maremane dome for 180 kms north-south. This thrust is the best documented of those in the Kheis Belt (Beukes & Smit, 1987).

The plane of the Blackridge thrust dips gently to the west at a maximum angle of 10°, except where steeper ramping occurs in more competent rock types such as iron formation and quartzite. The thrust is generally sub-parallel to bedding, and the mylonitic fault planes are generally confined to shale beds and thus may be overlooked. The underlying Griqualand West strata are little affected, and the structure may be of decollement type.

Surface structures show that the strata have been displaced eastwards for at least 35 km. Estimates of actual displacement are about 55 kms (Beukes & Smit, 1987). Movement of this order of magnitude must have given rise to substantial crustal thickening. There is, however, no evidence of the preservation of resultant sedimentary sequences directed either towards or away from the craton margin.

In the Prieska region, that is in the southernmost part of the craton, similar tectonism has affected the Griqualand West Sequence and also the underlying Ventersdorp and basement. Folding, southeast verging thrusts and shearing parallel to bedding, have led to imbricate zones and repetition of the Griqualand West stratigraphy. Coward & Potgieter (1983) relate these structures to a blind floor thrust and suggest that this structure, presumably an expression of the Kheis orogeny at a lower structural level, may continue for some distance. They speculate that this blind thrust could be responsible for many of the large, open structures in the Griqualand West Sequence of the northern Cape. If this is the case, then no phase of post-Transvaal but pre-Olifantshoek folding need be involved.

The Kheis Belt represents a thin-skinned east-verging thrust belt. (Fig.19). What cannot be demonstrated is whether this fold belt reflects the end stages of a Wilson cycle of plate tectonics, or whether the processes reflect ensialic movements.

Failure to recognise a collision zone or an ophiolite sequence can be interpreted, as has been done further north in the Magondi belt of northwest Zimbabwe (Leyshon & Tennick, 1988) as indicating that a plate tectonic model was not in operation. Alternatively, the absence of such features may simply reflect the fact that they would not be present in an eastern foreland, but would lie further west in a major belt and subsequently have been obliterated (Stowe, 1989).

This study makes no contribution to this philosophical debate, but does anticipate a relationship between the major strike-slip faulting to the north and the opening of the rift.

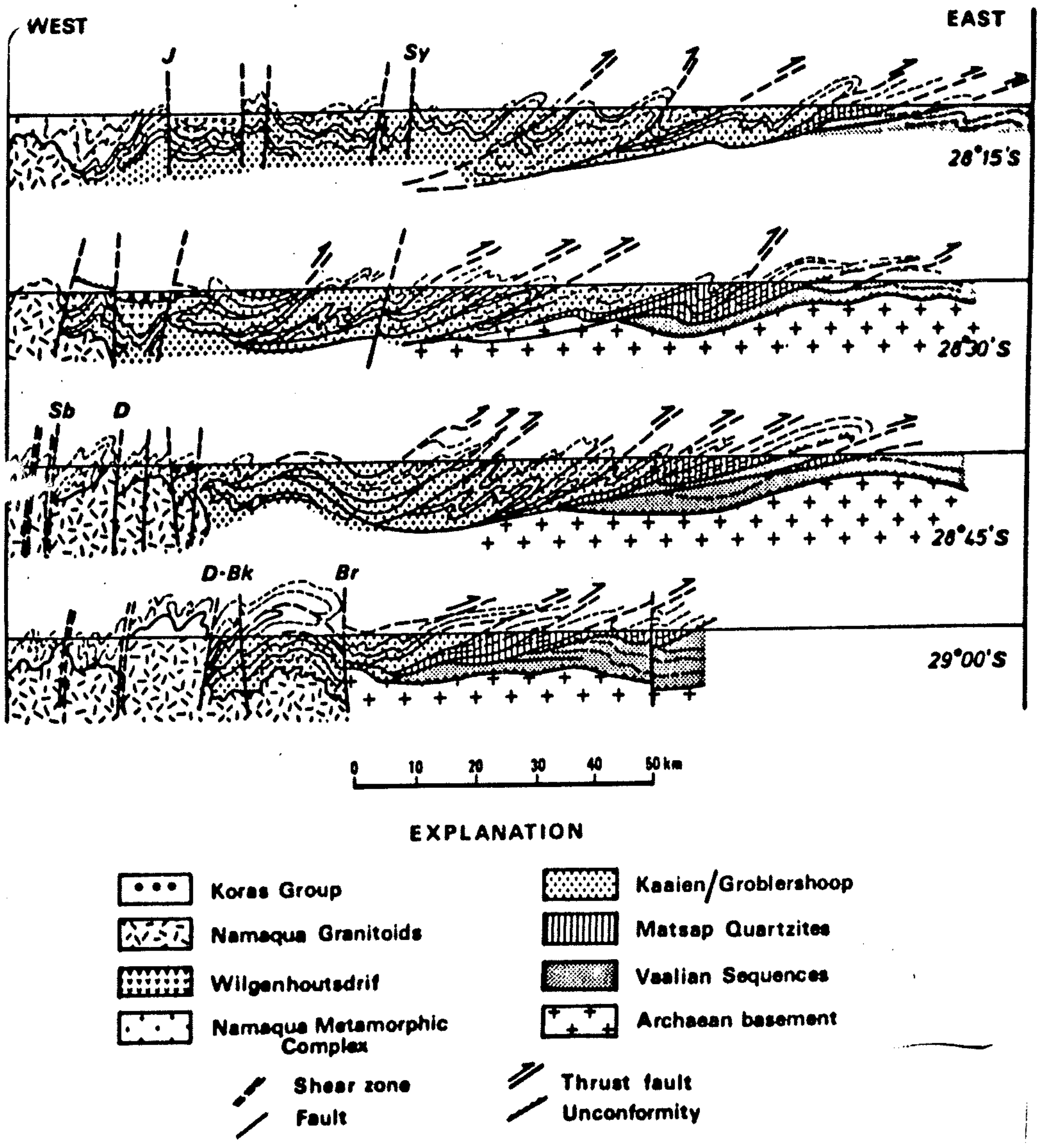

Fig. 19. East-west sections through the Kheis Belt.
(Hartnady & Others, 1985)
Reproduced with permission of *Episodes*

THE SOUTPANSBERG RED BEDS

The red beds of the Soutpansberg occur in a 200 km-long trough which opens out eastwards from a point east of Blouberg: in the east the sequence eventually disappears under a Karoo cover near Punda Milia (31°E), at which point the trough is 40 km wide.

Faulting in the Soutpansberg, especially in post-Karoo times, means that in effect a continuous undisturbed sequence does not exist. However, a major conclusion of the most recent regional work in the area by the Geological Survey (Jansen, 1975) supports the view that the sedimentary and volcanic rocks belong to a single cycle, that there are no significant unconformities in the overall sequence. This differed from an earlier interpretation (van Eeden & Others, 1955) in which the basal lavas (now Sibasa Formation) were correlated with pre-Witwatersrand strata, and a further distinction made between older Loskop equivalent sediments and a predominant Waterberg equivalent (Truter, 1949). Van Eeden & Others had postulated thrust faults to explain the presence of volcanics and supposed lower Waterberg sediments higher up in the column. Such contacts are now regarded as normal.

Thicknesses are difficult to estimate, but increase eastwards to perhaps as much as 10 000 m. Up to half of the total thickness may be volcanics, most of which are in the lower part of the succession, particularly in the Sibasa Basalt Formation.

Four major Formations are recognised in the Soutpansberg Group. Contacts between them are often gradational, and the subdivisions applied stress, in particular, the relative proportions of lava and sediment (Table 8, Fig. 20).

Sporadically developed from Louis Trichardt eastwards, a thin (< 10m) sedimentary sequence, the Tshifhefhe Formation, underlies the Sibasa Basalt. Quartzitic material in this basal unit is distinguished by its felspathic content.

The thick Wylliespoort Quartzite is extensively developed. It is thought to occur as far west as the isolated Roodepoort outlier, but it should be noted that this occurrence lies in a distinctly different NW-trending tectonic domain.

At the western limit of the Soutpansberg, the Wylliespoort Formation overlies specific members of the Blouberg Formation at Blouberg, but little can be gleaned of the absolute age of either unit from this clear stratigraphic relationship.

To the north of the main outcrop (but southwest of Tshipise), a predominantly volcanic series, a probable correlative of the Sibasa Basalt, has been distinguished as the Stayt Formation (SACS, 1980). An isolated succession on the Limpopo River (east of 31°E), the Mabiligwe Formation, is regarded as a correlative of some part of the Soutpansberg Formation; in a thin sequence, a conglomerate rests on basement, is succeeded by red-brown tuffaceous and sandy shale, and by sandstone and quartzite.

Formation	Member	Lithology	Thickness (m)
Nzhelele	Lukin Quartzite Mutale Tuff Musekwa Basalt	Predominantly white or light-coloured, brown-weathering, laminated quartzitic sandstone with interbedded shale and sandy shale Lukin Member: white quartzite Alternating reddish, brownish, or variegated shale, shaly sandstone, sandstone, and quartzitic sandstone with clay pellets and pellet conglomerate. Locally there are intercalations of tuff (Mutale Tuff member) or calcareous basaltic, andesitic, trachytic and gabbroid lavas with interbedded tuff, ignimbrite, sandstone, shale and chert (Musekwa Member)	1 000 - 2 000
Wylliespoort Quartzite	Bluebell Conglomerate Devils Gully Basalt	White, pink and light-coloured, medium-grained quartzitic sandstone and purple, brown or reddish coarse-grained sandstone, locally with interbedded pebble washes, grit, conglomerate, shale, mudstone, siltstone and lava Bluebell Conglomerate Member: bounlder conglomerate Devils Gully Member: lava, mudstone and siltstone Calcareous rocks at base in one locality	1 000 - 4 000
Fundudzi		Light-coloured quartzitic sandstone and quartzite and purple, brown or reddish sandstone, locally gritty or conglomeratic with interbedded lava, tuff, agglomerate, shale, sandy shale, and siltstone	0 - 2 800
Sibasa Basalt		Predominantly basaltic, andesitic and locally trachytic and gabbroid lavas, with interbedded tuff, agglomerate, ignimbrite, quartzite, quartzitic sandstone, grit, conglomerate, shale, mudstone and siltstone	0 - 3 300

Table 8: LITHOSTRATIGRAPHIC COLUMN OF THE SOUTPANSBERG GROUP (SACS, 1980)

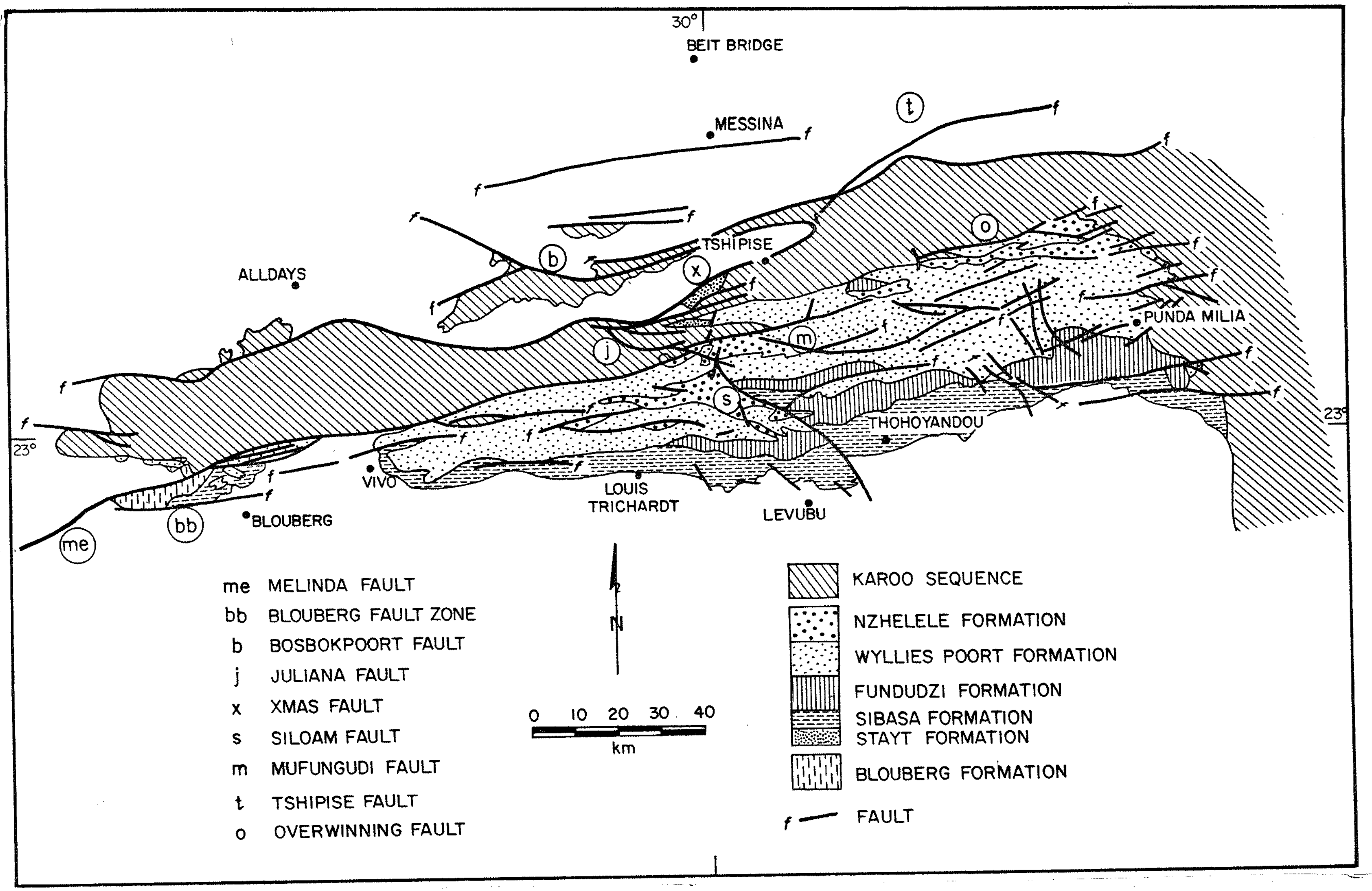

Fig. 20. The geology of the Soutpansberg red beds.

Compiled by the South African Development Trust Corporation Ltd. from available information.

No specific environmental studies have been carried out in the Soutpansberg, but the general environment is seen as alluvial plain deposition in braided streams (Jansen, 1976; Barker, 1979).

In contrast to other areas such as the Waterberg Basin, the Soutpansberg contains little coarse conglomeratic material, and significant argillaceous material towards the top of the sequence.

Up to and including the Wylliespoort Quartzite Formation, the sediments are predominantly arenaceous, and frequently channelled. Both planar and to a lesser extent, trough cross-beds are developed. A sandy, braided mid-alluvial plain is seen as the likely depositional environment for these sediments (Tankard & Others, 1982).

The sediments of the Nzhelele Formation commonly occur in thin, upward-fining cycles of up to 2m thick, in which mud-clast conglomerates are overlain by small and medium-scale trough and planar cross-beds and are capped by shales with mudcracks and raindrop imprints. The depositional environment envisaged was probably a distal, braided alluvial plain (Tankard & Others, 1982). Frequent switching of channels could have created stagnant ponds on the abandoned floodplain, leading to subaerial exposure of some sediments.

Palaeocurrent data is very limited: de Villiers (1967) and Barker (1979) consider that the available evidence indicates a predominantly northwest to northeast, that is, broadly northern source for the Soutpansberg sediments. The limited data provided by Barker suggests a northwesterly provenance.

In contrast to this, Jansen (1976) considers that this northern source applies only to certain upper parts of the sequence:

> ..." In the middle portion of the succession in the western Soutpansberg (29 measurements) are from all directions between north and southeast and in eastern Soutpansberg (9 measurements) from the west and west-northwest.... only in one of the upper divisions (39 measurements) transport directions are predominantly from the north" ...

As will be seen, the above data are used in support of different types of rift structure: Barker (1983) favours a yoked basin with a unimodal provenance in the Central Zone of the Limpopo Belt to the north; Jansen (1975) a penetrative aulacogen with a more varied source area. But more data are required before palaeocurrent analysis can contribute to an understanding of the evolution of the Soutpansberg.

Four palaeomagnetic measurements are available from the Soutpansberg (Jones & McElhinny, 1967); they provide three distinct palaeomagnetic pole positions (McElhinny, 1968). The measurements form part of a study involving other strata from these red beds to the southwest, and will be discussed in a subsequent section.

The volcanics in the Soutpansberg are thought to have formed along a linear vent system in a series of subaerial, cyclic extrusions (Barker, 1979). Individual flows are up to 2m thick. The flows are porphyritic at the base, massive in the main with amygdaloidal tops. Pyroclastic material is common, frequently occurring above the lavas: basic and acid pyroclastic rocks are present and these include agglomeratic and tuffaceous deposits, and lapilli and tuffaceous ash flows.

Tholeiitic basalt is the predominant volcanic material, with low potassium contents and high rubidium (Barker, 1979). Intrusive diabases are common too, and have a similar chemistry to the lavas. Felsic pyroclastic and lava flows are relatively abundant in the Soutpansberg (Barker, 1979). The lavas are predominantly dacitic. The pyroclastic rocks are extensively altered, but a range of textures suggests differing origins through pyroclastic flow, air fall and possibly base surge (Bristow, 1986). Alteration and low-grade metamorphism of the Soutpansberg igneous rocks is fairly ubiquitous, whereas regional metamorphic effects are not that noticeable in the associated sediments (Bristow, 1986).

The mafic volcanics are composed of relict phenocrysts of plagioclase and clinopyroxene, and a metamorphic assemblage of pumpellyite, chlorite, epidote, quartz and sericite (Crow & Condie, 1990). No metamorphic fabrics are recorded.

Very limited age data are available for the Soutpansberg Group.

- Rb-Sr whole rock errorchrons of 1756 $\pm$ 27 m.y. on lava, and 1760 $\pm$ 72, 1756 $\pm$ 17 m.y. on sills from the Sibasa Formation (Barton, 1979);

- $^{40}K/^{39}Ar$ dating of 2025 $\pm$ 23 m.y. from the Musekwa Basalt (of the Nzhelele Formation)(Burger & Coetzee, 1973).

In reviewing data from the north-central part of the Kaapvaal craton, Walraven & Others (1990), conclude that the above sill ages suggest that the Soutpansberg sediments are older than 1750 m.y.

Using geochemical tectonic discriminant diagrams the Soutpansberg volcanics plot within the subduction related fields, in particular with affinities to volcanics from continental margin areas (Crow & Condie, 1990). Noting that these lavas formed in a continental rift, Crow & Condie recognise, however, that any subduction-zone geochemical component must have been acquired during earlier subduction regimes.

The structural pattern of the trough is complex. The older the faults are, the more they are likely to be obscured and/or reactivated. Different senses of movement may be involved in specific faults.

The dominant fault pattern is ENE, parallel to the axis of the trough. The original northern boundary may be represented by the Xmas, Tshipise and Juliana faults, faults which have been reactivated in post-Karoo times (Fig. 21).

Other structural features of importance in the Soutpansberg are the duplication of the succession by strike faulting, and monoclinal tilting of the succession to the north. Karoo sedimentation and post-Karoo faulting are a part of the overall and complex history of the trough. Tilting to the north took place in post-Soutpansberg, and in post-Karoo times (Fig. 21).

The structure and, in particular, the thick sequence of volcanics indicates that the Soutpansberg is an extensional rift.

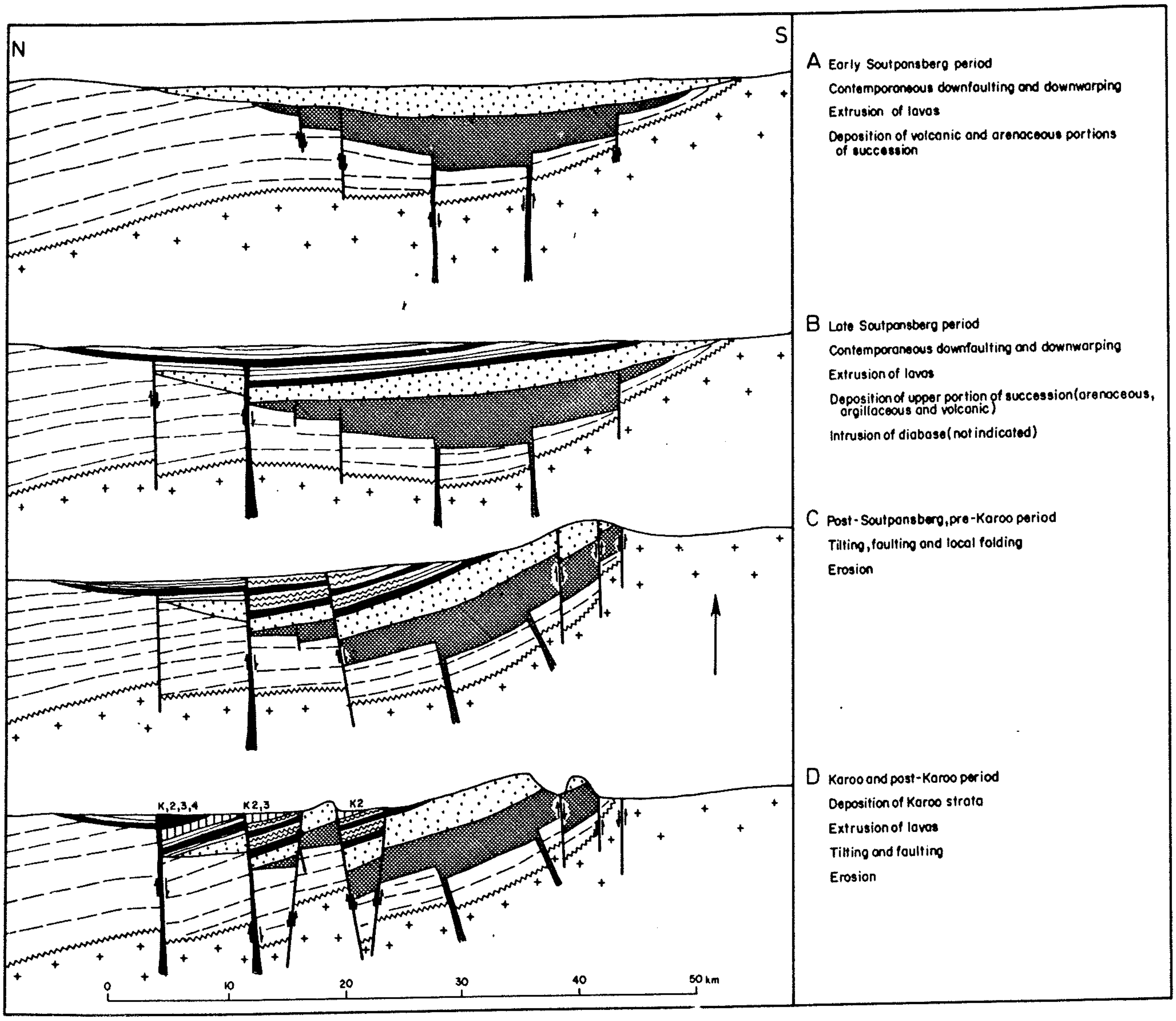

Fig. 21. The evolution of the Soutpansberg (Jansen, 1975) Reproduced with permission of the Geological Society of South Africa.

Numerous localised deposits of copper have been found in the Soutpansberg. Although there are no deposits of economic importance, there are records of copper being worked in historic times as far back as 1904 (in the Mutale region) (Wilson, 1989).

Copper occurs in two broad bands on and adjacent to the northern and southern contacts of the Soutpansberg (Wilson, 1989).

In the northern zone copper is associated with flow-top amygdales in the basal Musekwa Basalt Member of the Nzhelele Formation. Copper also occurs associated with the Mutale Tuff Member of this formation. The mineralisation is finely disseminated, and occurs as chalcocite, with native copper and chalcopyrite, and associated secondary malachite and azurite (Barker, 1979). Overall most of the mineralisation in this formation occurs in fissures and quartz veins.

The southern zone is less significant: here the copper is concentrated in joints, faults and shear zones within the Sibasa Formation Basalts, or else occurs as disseminations in epidotised lavas. Highest mean values of copper (127 ppm) have been recorded in this formation (Wilson, 1989).

Low-grade manganiferous sandstones occur in the Soutpansberg.

THE ORIGIN OF THE RIFT

It is accepted that the Soutpansberg represents a rift structure. At issue is the nature of the rift: as mentioned earlier, Barker (1979, 1983) favours a yoked basin with sediment provenance in the Central Zone of the Limpopo Metamorphic Belt to the north, Jansen (1975, 1976) a penetrative aulacogen.

Rifting need not necessarily be controlled by lateral extension: an extensional shear can govern rifting. Given the post-Bushveld transcurrent movement known elsewhere on the (adjacent) Zoetfontein Fault, such a mechanism represents another possible origin.

In further consideration of the Soutpansberg structure, it is as well to understand the terms 'aulacogen' and 'yoked basin' referred to earlier.

Aulacogens are deep, linear, graben-like features that cut cratons. The concept is a Russian one which originated in a philosophical climate in the USSR of a periodically expanding earth (Shatsky & Bogdanovic, 1961). Thus there were references to a "network of differently oriented aulacogens over the whole area of earliest platforms which have been subjected to some general horizontal extension (crawling away)".

Three types of aulacogen were distinguished: 'through' aulacogens dissecting the platform; 'penetrating' aulacogens, which penetrate into and alternate within the platforms; and 'inner' or 'blind' aulocogens, attenuating at both ends within the platform. Aulocogens were subsequently integrated into plate tectonics and are seen as failed rifts related to triple junctions, cutting into cratons at high angles to developing plate margins (Burke, 1977). Burke's comments relate specifically to penetrating aulacogens.

Aulacogens display diversity, but (a) common stages can be distinguished in a cycle and (b) aulacogen development may recur on the same site. They are characterised by the long duration of their development. They usually occur on ancient linear tectonic zones.

The aulacogen cycle commences with an incipient stage involving either the collapse of an uplifted arch or development related to gentle trough formation. It involves graben formation, steep normal faults and, often, flexuring and the development of monoclinal structures in the basement. In the main subsidence stage from 1-2 to 5-15 km of sediment and volcanics may accumulate over a period measured in tens or hundreds of millions of years. The volcanism is basic or alkaline-basic in nature. In most cases, the cycle is completed with an inversion stage involving, unlike the earlier phase, some compression. This phase may take place immediately after subsidence, or as much as 100 m.y. later, and involves either folding (and in some instances thrusting - the Benue trough is an example of this); or, less commonly, linear horst-type uplifts usually in narrow zones.

A yoked basin (Krumbein & Sloss, 1963) is a subsiding area adjacent to a complementary uplift that supplies detritus to the area of subsidence. Boundary faulting is common: such faulting tends to develop an asymmetrical cross-section in the yoked basin with the thickest deposits, characteristically wedge arkoses, tending to accumulate adjacent to the bounding fault.

Attributes cited by Jansen (1975) in support of an aulacogen concept for the Soutpansberg include:

(1) The Soutpansberg trough resembles an aulacogen in shape, size and thickness of the stratigraphic succession

(2) The trough dies out, westwards.

(3) Oblique-slip faults are common in the Soutpansberg Group. This fault pattern is characteristic of the Siberian and Canadian aulacogens.

(4) Basalts in Canadian aulacogens are characteristically tholeiitic in the lower parts of the sequence and alkaline higher up. Thick tholeiitic basalt flows characterize the lower Soutpansberg Group.

(5) Regional metamorphism and batholithic intrusions are absent in the Soutpansberg.

The younger cover to the east makes it impossible to establish whether such an aulacogen represents the failed arm of a triple junction, and if it does, whether this relates to Soutpansberg or Karoo times.

The essence of Barker's contention is that the sediment was transported from the north, and its provenance was the Central Zone (C2) of the Limpopo Metamorphic Belt. Barker suggests that the palaeocurrent evidence supports this but, as had been indicated, that evidence remains inadequate; nor are there arkose wedges or appropriate facies distributions associated with such an asymmetrical structure.

It is also most uncertain that the Limpopo Metamorphic Belt did form the provenance for Soutpansberg red beds.

The Central Zone (C2) of this belt had a complex, and unique, pre-2650 m.y. history. With the marginal zones, it was then affected by a major tectono-metamorphic event (du Toit & Others, 1983). A very major crustal thickening is envisaged with subsequent isostatic readjustment leading to deep-seated metamorphic rocks occurring at the earth's surface.

Certainly a number of workers have suggested that the Soutpansberg sediments and volcanics followed faulting after a period of continuous uplift of some 400 - 600 m.y. (see for example Barker 1979, 1983; Bristow, 1986). Whatever the scale of the uplift that undoubtedly affected the Limpopo Metamorphic Belt it is difficult to envisage an ongoing process extending for the equivalent of the duration of the Phanerozoic.

More recently it has been suggested that the isostatic readjustment (uplift) following the main metamorphism at 2650 m.y. could have taken place in less than 50 m.y. (van Reenen & Others, 1987). Such timing provides no grounds for assuming that the C2 formed the provenance for the Soutpansberg sediments 600 million years or more later.

Further information suggesting a linkage between the Limpopo Metamorphic Belt and the Soutpansberg is the number (twenty-three) of Rb-Sr mineral ages from various localities in the belt that show a very pronounced peak at 2000 m.y. (van Breemen & Dodson, 1972). However, these measurements need not necessarily reflect an uplifted potential sedimentary source at that time: Morgan & Briden (1981) comment:

> ..."Whether these mineral ages represent uplift and cooling after a long and continuous period of deep burial, or a later thermal event distinct from an earlier cooling episode is not known. What is important is that the whole area last cooled through the blocking temperature for Rb-Sr mineral systems at about 2000 m.y." ...

It is relevant to note that a $\sim$2000 m.y. event is not unique to the Limpopo Metamorphic Belt - rather the opposite: the Kaapvaal Craton is redolent with magmatic events and evidence of a thermal event at that time.

Combination of structure and stratigraphic elements has led to the definition of seven tectonostratigraphic domains in the Limpopo Metamorphic Belt, separated from each other by a variety of contacts (Watkeys, 1983). Major belts of lateral shearing involved in this are of post-Bushveld age (van Reenen & Others, 1987). The disposition of these distinct terranes is not part of the Limpopo Belt of metamorphism as such, but reflects post-Bushveld crustal adjustments along a crustal lineament. It is in relation to these later adjustments that the origin of the Soutpansberg should be seen.

On the present evidence, no relationship between the Soutpansberg rift and the Limpopo Metamorphic Belt can be sustained. The rift may well represent a penetrative aulacogen. However, some genetic linkage may yet be established between the formation of the rift and transcurrent movement on the Zoetfontein fault.

PALAEOMAGNETISM

The significance of palaeomagnetic studies in red beds was discussed in the Introduction. Only one major study has been carried out in these beds: and that was undertaken in the 60's.

In the early palaeomagnetic study, twelve samples from four localities in the early Proterozoic red beds were examined, the bulk from the Waterberg region (Jones & McElhinny, 1967). The localities are shown in Fig. 22; note that at that time more of the red beds were classified as (then) Loskop System than is the case now. The stratigraphic position of the samples are given below.

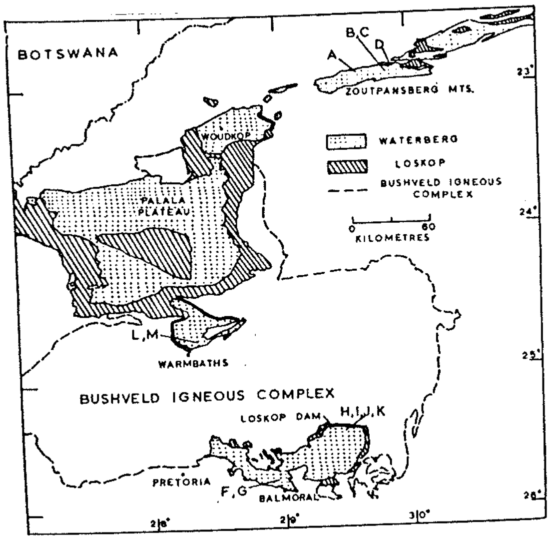

Fig. 22. Sampling sites of palaeomagnetic study by Jones & McElhinny (1967).

The Loskop/Waterberg distinction follows terminology followed at that time.

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Zoutpansberg	Warmbaths	Balmoral	Loskop Dam
	Present Upper Limits		
			500 metres
A(29°, 332°E of N) (near top of succession)		700 metres	H(9°, 191°E of N)
? obscured by faulting	???		300 metres
C(24°, 5° E of N)		F(5°,due north)	I(17°,170°E of N)
100 metres		80 metres	300 metres
B(24°, 5°E of N)	L (flat lying)	G(5°,due north)	J(56°, 194°E of N)
?	200 metres	800 metres	200 metres
D(22°, due north)	M (flat lying)		K(59°,183°E of N)
200 metres	200 metres		300 metres
	Base of System		

McElhinny (1968) groups these sites as follows:

Group 1 - A.H
Group 2 - B,C
Group 3 - F,G,L,I
Group 4 - D
Group 5 - J,K,M

and suggested some correlation with the stratigraphy:

> ..."Sites J,K and M lie near the base of the succession in one group {and show reversed polarity} and sites A and H, near the top of the succession, form a separate group. Sites F,G,L and I, which are certainly higher horizons than J,K and M, form a third group". ...

At that stage, it was probably appropriate to regard the Loskop and Waterberg as a simple and unitary succession, wherever developed. It is now apparent that with the exception of the samples from the Zoutpansberg, all the others are from the Loskop Formation (Loskop Dam) or the Early Waterberg (Balmoral, Warmbaths).

Morgan & Briden (1981) provide a further pole position for red beds immediately underlying the acid volcanic horizon at Rust der Winter. It will be recalled that this is regarded as a correlative of the Loskop Formation (Coertze & Others, 1977), although the age determination of the volcanics, 1790 ± 70 (Oosthuyzen, 1964) appears too young (Walraven & Others, 1989).

The pole position determined for this occurrence lies reasonably close to McElhinny's 1968 Group 5, has the same south-seeking polarity, and in addition lies close to the palaeomagnetic pole position of the Bushveld gabbro.

Morgan & Briden conclude that the correlation of the Rust der Winter sediments with the 'Lower Waterberg' appears to be safe. They also consider that the Rust der Winter may not be extrusive, and suggest that the 1790 ± 70 age on the 'felsite' "can now be regarded as a minimum age not only for the Ruster der Winter site but also for its palaeomagnetically correlated lower Waterberg equivalents".

Additional studies on the acid volcanic horizon at Rust der Winter failed to isolate a stable remanence for this material, although some indication of a direction almost identical to the direction of the present Earth's field (Turner's (1979) Type C) is suggested.

Other available data (Fig. 23) includes that on the Bushveld granite (Gough and Van Niekerk, 1959; Hattingh, 1989), on the Limpopo Belt and Phalaborwa (Morgan & Briden, 1981). The rocks of the Limpopo Belt have two stable magnetisations: Component A is thought to have been acquired during cooling of the belt at about 2000 m.y., Component B a later chemical remanent magnetism of unknown age. The Phalaborwa studies indicate two bursts of igneous activity suggested at circa 2000 and 1900 m.y.

The very limited palaeomagnetic results from the Waterberg can mostly be related to an APW curve for Africa crossing northern Africa obliquely from ENE to WSW between 2050 and 1850 m.y. (Jones & McElhinny, 1967; Morgan & Briden, 1981; Hattingh, 1989) (Fig. 23). This path includes McElhinny's Group 5, 4 and 3; and 2 (not shown by the above authors); and the Rust der Winter measurement.

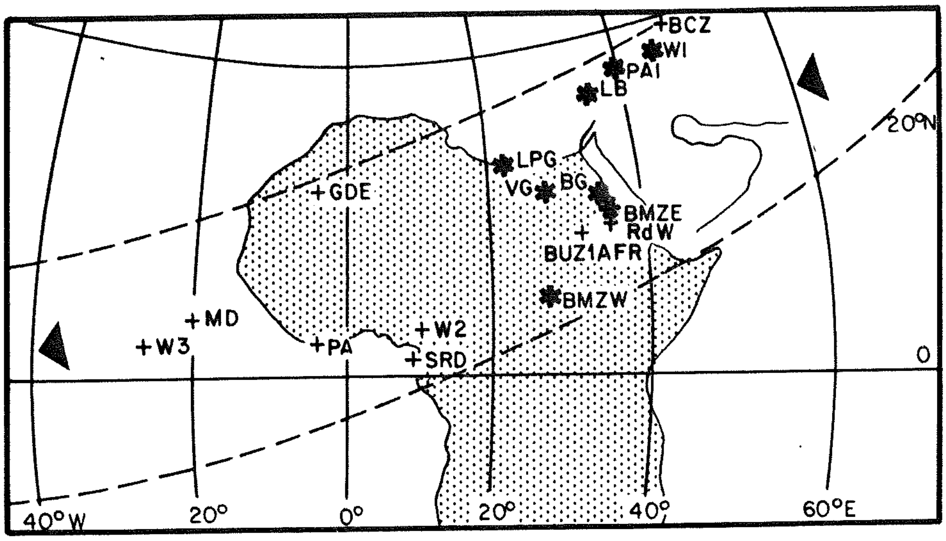

Fig. 23. Apparent Polar Wander path for Africa with selected pole positions from southern African data for the period ~2070-1700 m.y.

W1-W3 - Waterberg sediments: groups E,D & C in McElhinny, 1968;
VG - 2006±47 m.y., Vredefort Granophyre (Hargraves, 1970);
LPG - 1980 ± 39 m.y., Limpopo gneisses (Morgan & Briden, 1981);
LB - 1840 ± 276 m.y., Losberg Gabbro (Hargraves, 1970);
GDE - 1985 ± 73 m.y., Great dyke extensions in Limpopo Belt, (Jones & Others, 1975);
SRO - 1876 ± 68 m.y., Sand River dykes (Morgan, 1985);
PA - 2012 ± 19 m.y., Phalaborwa syenite and dolerites (Morgan & Briden, 1981);
MD - 1811 ± 20 m.y., Mashonaland dolerites (McElhinny & Opdyke, 1964);
BG - Bushveld Gabbro (Gough & Van Niekerk, 1959);
BCZ - 2054 ± 11 m.y., Critical Zone, Bushveld Complex;
BMZE - 2049 ± 152 m.y., Main Zone, eastern Bushveld Complex;
BMZW - Main Zone, western Bushveld Complex;
BUZ1AFR - Upper Zone, Bushveld Complex. For specific references to BCZ, BMZE, BMZW, BUZ1AFR by Hattingh, see Hattingh (1989);
RdW - 1790 ± 70 m.y., 'extrusive' in Rust der Winter Formation.

* - N poles; + - S poles

Modified after Hattingh, 1989; reproduced with permission of *Tectonophysics* and P.J. Hattingh.

McElhinny's Group 1 (single measurements from the Soutpansberg (A) and Loskop Dam (H) does not lie on this path: this pole position lies near to Scandinavia - northwest USSR. In relation to available information, these measurements suggest an age of ~ 1100 m.y. (Fig. 24).

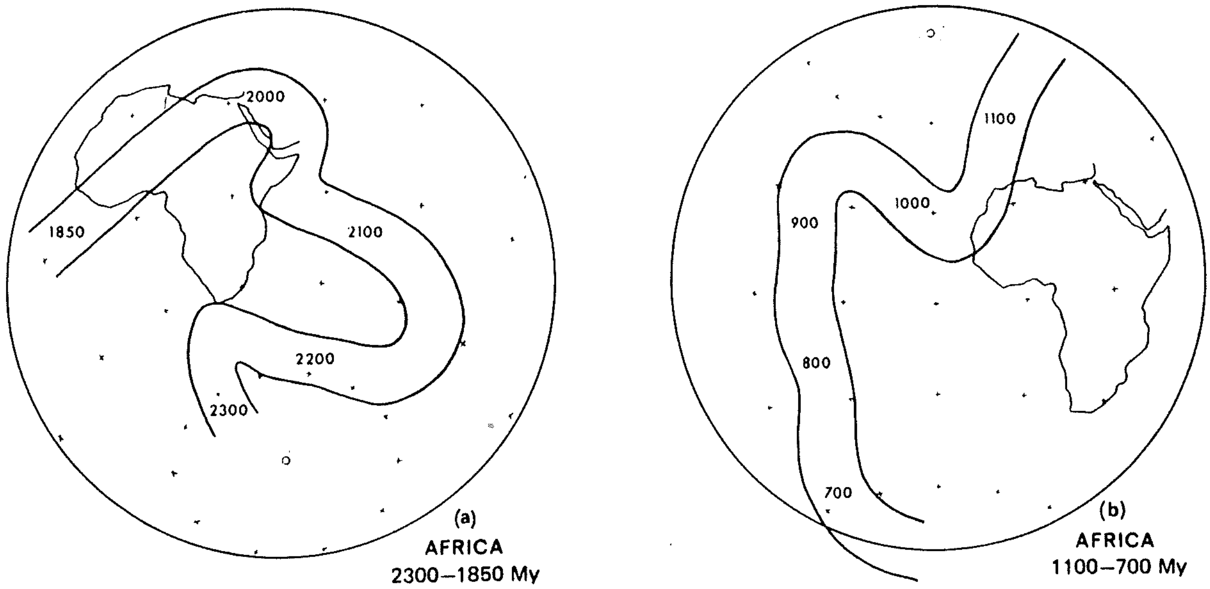

Fig. 24. Precambrian Apparent Polar Wander paths for Africa. From Morgan & Briden, 1981. Reproduced with permission of *Physics of the Earth and Planetary Sciences.*

Such an age approximates to that of the Umkondo Group in southeastern Zimbabwe. The Umkondo has been seen as a correlative of the red bedsunder discussion by some (see for example Stowe, 1989). If these red beds are $\sim$ 1800 m.y. old, then such a correlation is not acceptable either on the basis of age or of lithology.

U-Ar ages for doleritic material cutting the Umkondo have ranged from 590 - 1785 m.y. Given the potential for argon loss the oldest age, of 1785 $\pm$ 90 m.y. for hornfels bordering a sill in the Umkondo was often accepted. However, recent Rb-Sr data from Umkondo dolerites and contained lavas in the sequence are interpreted to indicate an age for these intrusives and contemporaneous extrusives of 1080 $\pm \frac{140}{25}$ m.y. (Allsopp & Others, 1989).

The Umkondo does contain very limited red beds, but overall is distinctly different to the Waterberg: it comprises continental, marginal marine and marine sediments deposited on an east-sloping continental margin (Button, 1977); and the strata and adjacent floor rocks were to be caught up in the distinctly later Zambezi (Pan-African) tectonic movements (Kröner, 1977).

Palaeomagnetic work in the 60's distinguished two directions of magnetisation in the abundant dolerite/diabase intrusives in northern, central and eastern Zimbabwe (McElhinny & Opdyke, 1964): a distinction between the sills intruding the Umkondo group (later a similar direction was determined for Umkondo lavas) and those elsewhere, referred to as the Mashonaland dolerites. Subsequently, it was shown that the numerous post-Waterberg, pre-Karoo diabasic sills in the northern part of the Transvaal and eastern Botswana had a mean pole position not significantly different to the Umkondo dolerites.

The Waterberg palaeomagnetic study referred to earlier (Jones & McElhinny, 1967) can be related to these pole positions, and that of the Bushveld Gabbro. Sites close to the base of the succession yielded directions like those of the Bushveld Gabbro, some intermediate horizons similar to the Mashonaland dolerites, uppermost samples gave directions close to that of the post-Waterberg diabases.

From the above, it is perhaps possible that the overall Waterberg accumulated over a period of 1000 m.y. but this seems unlikely: the specific palaeomagnetic measurements in the Loskop area alone, in the Loskop Formation and the early Waterberg Wilgerivier Formation, are, successively, similar to Bushveld, Mashonaland dolerite and post-Waterberg diabase directions. A similar spread in age would be required for the Zoutpansberg.

It appears prudent to accept the comment of Allsopp & Others (1989) that the earlier Waterberg pole directions should be viewed as preliminary; further interpretation of the measurement from Site H (and A) may show it (them) to represent a subsequent modification of an earlier pole position. Certainly there is a need for further examination of the significance of the $\sim$1100 m.y. event on the Kaapvaal Craton.

A major detailed palaeomagnetic study of the 'Waterberg', and of related intrusives, linked to an understanding of its diagenesis, would be of considerable value. It should be sufficiently comprehensive to test the stratigraphic relations as known, and the uncertainties that still exist. The implication of an APW across northern Africa at the time of Waterberg sedimentation requires specific palaeoclimatic consideration: low-latitude desert conditions would be precluded.

ECONOMIC CONSIDERATIONS

The red beds of the northern Cape contain very major iron and manganese deposits at and immediately below the base of the Olifantshoek Sequence in the Sishen-Postmasburg region: in the Manganese Iron Formation, and as reworked deposits at the base of the overlying Gamagara Formation.

For the rest, as has been described, little of economic value is known in these red beds. The issue is one of whether or not the suite has been adequately researched. It is felt that any further studies in these early Proterozoic red beds, whether directly related to exploration or of a more basic nature, should consider, or in some instances consider further, possible models that may be of economic significance. These include:

. consideration of possible mineralisation related to the unconformity at the base of the suite; and on other major unconformities within the sequence, if they exist. Such mineralisation could involve uranium, and also platinum group elements (PGE) and gold;

. further consideration of copper in basalts;

. awareness that a major type of copper deposit is found in association with red beds. In virtually all cases this occurs at a (first) marine transgression over red beds, or in intercalated non red beds within red bed sequences;

. noting that whereas virtually all gold, and uranium, found in fluvial conglomerates comes from rocks that are older than the earliest red beds (and younger than granite greenstone terranes), nevertheless in a solitary example, of the Tarkwaian of southwest Ghana, gold is found in this setting in association with haematite;

. the Olympic Dam copper-uranium-gold deposit in South Australia was regarded as an unusual type of sediment-hosted mineralisation associated with haematite. Continuing work has now shown that the deposit has resulted from explosive brecciation and hydrothermal alteration. Thus a relationship between this type of deposit and red beds no longer exists.

It is axiomatic for virtually all types of ore deposit that the basic elements of formation (namely source, transport and trap) be represented. Equally, a range of specific factors involving origin, structure, lithology and so forth, control the occurrence of individual deposits. Notwithstanding these points, it may be of value to make some general comments about the types of deposit referred to earlier.

RED-BED COPPER

Main types of sediment-hosted stratiform deposits are base metal: they include red-bed copper, sandstone-lead and carbonate hosted lead zinc. The deposits occur in tectonically active intracratonic settings, commonly in fault-controlled basins.

Differences in weathering lead to the formation of red-bed copper and sandstone-lead deposits. Copper is released early during mild chemical weathering of the more readily altered mafic minerals in rapidly deposited, immature, rift-generated felspathic red-bed sandstones. More intense and prolonged chemical weathering, brought about by stable tectonic conditions leading to peneplanations, give rise to transgression onto the craton; the development of relatively more mature quartzitic sandstone is required to release lead through the breakdown of felspar. The zinc-dominant nature of the carbonate-hosted deposits is paralleled by the corresponding zinc-dominant composition of sea water. The zinc dominant effect could be enhanced by any groundwater passing through (Bjørlykke & Sangster, 1981).

Copper deposits associated with red beds are known to extend back in time to more than 2000 m.y. (Bowen & Gunatilaka, 1977); that is, as far back as the red beds themselves.

In such deposits, as in virtually all types of ore deposit, it is necessary that the basic elements of formation, namely source, transport and trap be represented.

For a stratiform copper deposit of this type to form, it is desirable for the relevant basin to contain large volumes of mafic volcanics, particularly tholeiitic basalts which have several times the average crustal values for copper: appreciable volcanics are developed towards the base of stratigraphic sequences under consideration in specific settings.

Most of the copper deposits of this type occur either at the time of the first marine transgression over red terrestrial successions or within marine (and, in some instances, lacustrine) sequences associated with red beds (Gustafson & Williams, 1981). A marine transgression occurs at the top of the Matsap red beds and (as correlated) at the top of the Palapye red beds; and a marine influence is inferred at the top of the Waterberg red beds.

The consistent feature of such deposits is a situation in which basinal groundwater could move to a shallow site of sulphide deposition.

Typically, such deposits are closely associated with gypsum, anhydrite or carbonate pseudomorphs after evaporite minerals. These evaporites are regarded as a likely source for the vast amount of sulphur that becomes fixed as sulphide in the deposit, and also for the high salinity required to move the metals at relatively low temperatures. There are numerous cases in the literature of the late identification of evaporites (see for example Gustafson & Williams, 1981). Although evaporites are not known in the early Proterozoic red beds under discussion, this could reflect that they have yet to be identified.

Most authorities agree that reduction of brine sulphur is the key mechanism for sediment hosted stratiform copper deposits. The various features of the deposits suggest that the brines were sulphate-rich and in equilibrium with haematite, quartz, felspar, illite and carbonates, and that they migrated along permeable sediments to sites of mineralisation. At these sites, some form of organic matter or previously deposited pyrite was encountered with the aquifer.

Arid (oxidising) environments are also desirable because the result favours the concentration of soluble chlorides, and sulphur as evaporitic sulphate for the suite under consideration. The admittedly limited palaeomagnetic data suggests relatively high latitudes of formation for these red beds.

COPPER IN BASALT

The Keweenawan Basalts of North Michigan have been worked for copper for a century. In this province copper, and zinc, were mobilised during prehnite-pumpllyite facies metamorphism (Jolly, 1974). Copper, averaging 70 ppm was leached from epidotised or more metamorphosed rocks in a zone of dehydration in the lower part of the section; it was precipitated in hydrated rocks near the upper part of the section. The boundary between dehydration and hydration commonly occurs in the prehnite-pumpellyite low-grade facies of metamorphism. Native copper is the common copper mineral.

It will be recalled that numerous uneconomic occurrences of copper have been found in the Soutpansberg: as dissemination in volcanic rocks, or as fillings and replacements in brecciated lavas and sediments - usually adjacent to major faults. A single consistently mineralised horizon occurs in the Nzhelele Formation; the mineralisation in this horizon consists of chalcocite, with native copper and chalcopyrite.

Barker (1979) notes that the Soutpansberg volcanics are ubiquitously altered: alteration minerals present are chlorite, epidote, quartz, sericite, and minor calcite, leucoxene, and actinolite. Stoljan (1974) had reported pumpellyite from near the base of the Zoutpansberg volcanics.

In a recent study Crow & Condie (1990) record that the metamorphic grade in the Soutpansberg volcanics is low: associated with relict phenocrysts of plagioclase and clinopyroxene is a metamorphic assemblage of pumpellyite, chlorite, epidote, quartz and sericite.

Further studies of the association of metamorphism and copper mineralisation appear to be warranted in the Soutpansberg.

UNCONFORMITY-RELATED URANIUM DEPOSITS

A specific class of uranium deposit occurs in the Proterozoic. Such deposits are found close to major erosional unconformities during a commonly occurring orogenic period about 1800-1600 m.y. ago. The classic examples of this class are the ore bodies of Cluff Lake, Key Lake and Rabbit Lake in northern Saskatchewan, Canada, and those of the Alligator Rivers area in the Northern Territory of Australia (Taylor & Cameron, 1980).

In the Alligator Rivers uranium field in Australia, the ore bodies have a number of features in common:

- all occur in the Cahill Formation;
- they are located in breccia zones;
- they are contained in rocks that underwent low-temperature retrogressive metamorphism; and
- they occur in proximity to granitoids containing 2-6 times the normal uranium abundance for granitoids.

Hoeve & Sibbald (1978) favour a diagenetic-hydrothermal origin for the Athabasca Basin deposits in Saskatchewan, Canada. This involved post-depositional oxidation and leaching, considered to have continued for a lengthy period after deposition. Ore deposits were formed under conditions of deep burial at elevated temperatures and pressures by interaction of a uraniferous oxidised aquifer with reducing, graphite-bearing metamorphic rocks of the basement floor. The large-scale convection required for such an interaction may have been induced by mafic magmatic activity coeval with the episode of mineralisation.

It will be recalled that Cheney & Twist (1986) consider that the red bed sequence in the Waterberg region is represented by five unconformity-bounded sequences (I - V). They note that the uranium deposits known elsewhere occur where sulphide or graphite-rich basement rocks are overlain by former aquifers and comment:

> ..."Because the Bushveld Complex rocks on the eastern margin of the succession and in the Villa Nora area on the north are only sparsely sulphidic, economically these margins are the least promising. Sequence IV, which is above the basal unconformity in the south-western portion of the succession, is also an unfavourable target because it consists of the silty Aasvoëlkop Formation. The remaining margins of the succession appear to have a greater economic potential: these margins have thicknesses less than 2 km, which need not preclude mining any future discovery..."

The thesis of Cheney & Twist is that the unconformities they describe represent very major breaks: whereas Sequence I was pre-Bushveld Granite in age, Sequences III - V might even be younger than 1420 m.y. In effect, they regard each unconformity as a basal unconformity.

GOLD IN FLUVIAL DEPOSITS

At the beginning of this century, the Waterberg conglomerates were prospected for gold, but with negative results (Jansen, 1982).

The most important source of gold is in, and related to, fluvial conglomerates; in addition, there has been substantial production of uranium from this host rock. With a single exception, these gold and uranium-bearing quartz-pebble conglomerates and associated coarse-grained arenites come from successions ranging in age from 3000 to 2200 m.y. Rudites in earlier Archaean-type greenstone belts have not been worth exploitation, and from about 2200 m.y. the development of iron formations, and somewhat later, of red beds, terminated the metallogenic epoch in which the auriferous and uraniferous conglomerates were formed.

The younger exception is the Tarkwa Goldfield in southwest Ghana. At Tarkwa, gold occurs in a conglomeratic zone at the base of the Banket Group. This Group lies within the overall Tarkwaian Supergroup, which overlies Birrimian schists. Intrusions into the Tarkwaian have been dated at 1968 $\pm$ 49 m.y. (Hirdes & Others, 1987).

The conglomeratic zone is represented by upward-fining cycles which, taken with the presence of trough cross-bedding, erosion channels, ripple-marks, and clast-supported gravels, suggests that the environment of deposition was a coalescing series of fluvial fans (Pretorius, 1981).

The pebbles of the conglomerate are predominantly of vein quartz. The matrix is of quartz and haematite, with ilmenite, magnetite and rutile. In the trough cross-bedding, the foresets are frequently delineated by detrital haematite grains. Haematite and other iron oxides also occur in disseminated form in the matrix.

The payable conglomerates are channel-lag gravels or reworked channel and point-bar rudites into which sand filtered, such sand carrying detrital gold, haematite, black sands as well as other heavy minerals.

A key point to note in this unique deposit is that, unlike the older pyrite-rich conglomerates such as the Witwatersrand, there is a close relationship of iron oxide, as haematite, with gold. At Tarkwa, the general case is the higher the quantity of haematite, the richer the gold concentrations.

The available evidence suggests that the Birrimian schists represent the source of the gold, but the source of the haematite has not yet been determined.

Contemporary studies of red beds sees their reddening as a diagenetic process linked to circulating groundwaters. The special significance of the Tarkwa Goldfield to this review is unrelated to diagenesis, but lies in the co-existence of detrital haematite and gold in fluvial deposits at a time when early red beds, such as those under consideration, accumulated. Further studies in conglomeratic fluvial environments in those early Proterozoic red beds would appear to be worthwhile, if only to characterise these widely scattered environments more precisely.

OLYMPIC DAM DEPOSIT

The Olympic Dam copper-uranium-gold deposit is estimated to contain in excess of 2000 million metric tons of mineralised material with an average grade of 1.6 per cent Cu, =.06 uranium oxides, 0.6 g/metric ton of gold, as well as silver and rare earths.

The deposit occurs under a cover of younger Precambrian to Cambrian sediments on the Stuart shelf in South Australia. A factor in its discovery was a relationship to a gravity and magnetic anomaly. At the time of its discovery in 1975 the deposit was thought to occur in a thick sequence of predominantly sedimentary breccias within the Olympic Dam Formation. The sequence was affected by pervasive haematite, chlorite and sericite alteration. Below the Olympic Dam Formation was assumed to be a granite equivalent to that occurring outside the mineralised area.

The deposit was regarded as localised within a northwest-trending trough or graben, arched across a northeast axis. Strike-slip and dip-slip faults occur. The graben-fill sediments were thought to have been deposited in an arid, subaerial environment during rifting or strike-slip faulting.

The paragenesis of the deposit is complex, but haematite is the most abundant ore mineral. It formed rock breccia and by replacement in the matrix, haematite formed both before and after the copper sulphides chalcopyrite and bornite (Hagni & Brandon, 1989). The gold formed at about the same time as the copper sulphides, and the uranium accumulated later.

The occurrence of stratification and of graded breccias had led early workers to conclude that the host breccias had formed as detrital sediments. Thus Roberts & Hudson (1983) concluded that Olympic Dam was a very unusual type of sediment-hosted mineralisation with features that can be explained by low-temperature syngenetic or diagenetic processes. Further work, and mine development, has now shown that the Olympic Dam breccias are of hydrothermal origin, formed by a combination of explosive brecciation and hydrothermal alteration and replacement (Oreskes & Einaudi, 1990). The breccias occur as dyke-like bodies in fractured granite, and formed by progressive hydrothermal brecciation and iron metasomatism.

The origin as now proposed removes this type of deposit from consideration in relation to red beds.

The inferred close temporal and spatial association with specific granitic and related volcanic suites, seen more clearly in the southeast Missouri Province, has yet to be demonstrated in these early Proterozoic red beds. Nevertheless, the relationship between the high-level Bushveld Granite and limited coeval volcanism near the base of the Bushveld red beds merits some consideration in this regard.

TECTONIC FRAMEWORK

The Early Proterozoic red beds on the Kaapvaal craton are a suite of continental sediments. Many, but not all, are fluvial. In virtually all of them the reddening is thought to be of diagenetic origin.

A focus of this review has been to establish whether or not an overall framework exists for these sediments. In the text differing views on individual areas have been presented. What follows does not necessarily repeat such comment, but reflects a preferred - at times derived - assessment of the framework in which these rocks accumulated.

The oldest and best dated red beds within these red beds occur as intercalated sediments within the upper part of the Rooiberg Felsites, in close association with the Bushveld granite in the Middelburg and Nylstroom basins or suites, and at Otse in close proximity to the Rustenburg Layered Suite. Sedimentation of these Bushveld red beds is controlled by surface, near surface and upper crustal adjustments related to Bushveld magmatism. The Bushveld red beds include strata currently classified as Loskop Formation (and its equivalents) and much currently referred to as the 'Early' Waterberg. The clear spatial relationships of these rocks (van Biljon, 1976) and the field relations - including relationships to the Bushveld Granite - suggest that these red beds accumulated over a relatively short period. Breaks in the sequence are regarded as being of only local significance.

The Rooiberg Felsites, Bushveld Granite and the ultrabasic/basic lobes to the east and west of it, and the sub-outcropping ultrabasic/basic Molopo Farms Complex further west in Botswana and the extreme northern Cape represent major manifestations of magmatic activity $\sim$ 2000 m.y. ago.

In the northern Cape there is a very large gravity anomaly, the Groenwater anomaly; it is some 230 km in length, a maximum of 60 km wide. It is possible that the anomaly is related to denser material within the sedimentary sequence on the margin of the craton, but the original postulate of a large mafic intrusive body some 5 - 8.5 km thick at a depth of less than 20 km (Smit & Others, 1962) cannot be discounted.

If such a body does exist, then the balance of these Early Proterozoic red beds lie immediately outside a magmatic arc, and do so at the western, northwestern and northern margin of the Kaapvaal craton, as presently exposed (Fig. 25).

There is other evidence of widespread magmatism, in some instances of a thermal event, scattered through the craton at this time. This includes

- The Losberg, and other small intrusions in the Vredefort region, is of similar chemistry to the Bushveld Complex, and the Brandfort gabbro (OFS) may be of similar age (Coetzee & Kruger, 1989)

- The Vredefort event is 'more or less contemporaneous with the Bushveld activity' (Walraven & Others, 1990)

- Alkaline complexes occur at Phalaborwa (eastern Transvaal) (Eriksson, 1984), and at Schiel (northern Transvaal)(van Reenen & Others, 1987)

. There is evidence of a thermal event at this time: the effect is widespread in the Limpopo Belt (Van Breemen & Dodson, 1972), in the Witwatersrand, and the Onverwacht (Barberton) cherts (information from Coetzee & Kruger, 1989).

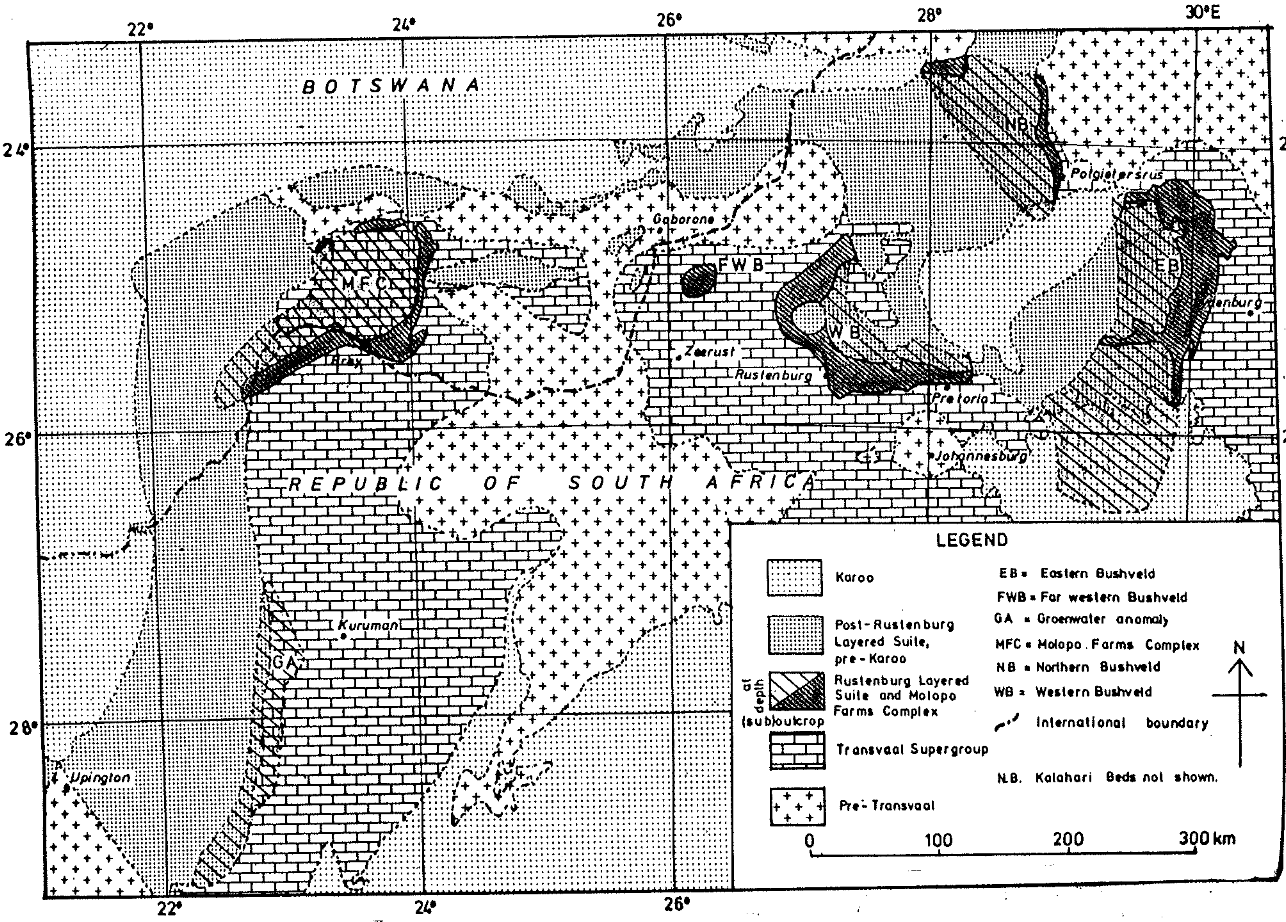

Fig. 25. Sketch map showing extent of suites correlated with the basic/ultrabasic rocks of the Bushveld Complex. (Gould & Others, 1987).

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It must be assumed that a mantle hot spot existed at this time. and the implication is of at most small relative motion of such a hot spot and the relevant continental crust. The array of igneous rocks generated reflects specific conditions.

It is at present uncertain what role this most marked Eburnian magmatism may have played in relation to the crustal movements that took place immediately afterwards.

Hatton (1988) suggests that there is evidence that the Bushveld Complex and related events formed near a plate margin: the high-magnesian nature of the Rooiberg felsites is geochemically similar to volcanic arc granites: the parental magma to the lower part of the Rustenburg Layered Suite is boninitic, a rock type found in Pacific island arcs; and the amount of olivine in the Lower and Critical Zones of the Rustenburg Layered Suite decreases southwards, suggesting a southward thickening of the lithosphere away from a plate margin. Hatton & Sharpe (1990) have suggested a model in which the felsites, and the Bushveld Complex, result from the collision of a spreading centre with the northern edge of the Kaapvaal craton: there is, however, as yet no tectonic evidence to support such a model.

White & McKenzie (1989) feel that even if mantle plumes do not drive rifting the resultant uplift is of assistance in helping subsequent movement along. Possibly the arcuate nature of the basic/ultrabasic intrusives indicate an eccentricity in the plume, and if so, this may be a more specific factor leading to crustal movements.

The west-southwesterly movement of the palaeomagnetic pole at this period is well documented. Hatton (1988) comments that prior to this the pole migrated S-N-S across the north pole. Subsequent sinistral movement on major faults is a reflection of the ensuing ENE movement of the craton.

There is evidence that the Bushveld Granite was rapidly unroofed, and that initial Waterberg red bed sedimentation (Alma Formation) was controlled by strike-slip sedimentation on the Murchison-Thabazimbi ENE-trending lineament; these early Waterberg red beds were shed northwards from this fundamental fault, but a major side-stepping soon took place with the left-lateral movement continuing on the Zoetfontein lineament to the north: the bulk of the Waterberg red beds were shed from this northern zone.

An inherent feature of strike-slip terranes is differential vertical movement. As a result, unconformities are bound to be limited in extent, and contemporaneous with sedimentation in adjacent areas. Correlation on the basis of such unconformities becomes inappropriate. The differential vertical movements give rise to both provenance and depository.

No relationship is envisaged between uplift of the Limpopo Metamorphic Belt and the Waterberg red bed sedimentation.

The present disposition of the Limpopo Metamorphic Belt is as a number of domains, each containing structural and stratigraphic elements and separated from each other by a variety of contacts. The most recent belts of shearing recognised are of post-Bushveld age. The Palapye red beds are associated with such a shear zone, the south-east trending Sunnyside Shear. Strike-slip faulting on this zone may also, as was the case with the Waterberg red beds, have controlled the Palapye red bed sedimentaiton. The mobilisation of the Mahalapye granite is a related phenomenon.

The Blouberg, lying on the Zoetfontein fault at the northern margin of the Waterberg red beds, is a zone of structural complexity. The strata include coarse clastics, and volcanics; some are folded. There is a braided fault pattern. It is felt that further structural work may show the Blouberg to be a succession of alternating, narrow zones of transpression and transtension arising from curvature along the main Zoetfontein strike-slip system.

From work on the sub-outcropping Molopo Farms Complex, Gould (1987) makes the point that it is difficult to do other than correlate the Waterberg with the Matsap red beds of the northern Cape. The view taken here, that the setting for many of the overall red-bed sediments resulted from crustal movements following major Eburnian magmatism, is in accord with this. But the setting of the Matsap red beds is distinctly different; and precise correlations between settings unlikely and inappropriate.

Within the Matsap red beds significant vulcanicity low in the sequence (within the Mapedi Formation, and as the Hartley Basalt), indicate attenuation and rifting of the crust. Some sediment transportation was parallel to the north-south rift, later the provenance lay east. In a broad sense rifting was either contemporaneous with the strike-slip movement (and Waterberg red bed sedimentation) to the north, or resulted from it. Even if the latter option prevailed, some overlap between Waterberg and Matsap sedimentation may still have occurred.

Subsequently the Matsap red beds and related sediments were thrust eastwards in a poorly dated but major thin-skinned orogenic belt, the Kheis Belt. What cannot be demonstrated is whether this fold belt reflects the end stages of a Wilson cycle of plate tectonics, or whether the processes reflect ensialic movements.

The Soutpansberg red beds accumulated in a linear feature. Understanding this complex region is compounded by later Phanerozoic re-activation : Karoo sedimentation and post-Karoo structure is superimposed on earlier events.

Up to half of the Soutpansberg sequence is composed of basaltic lavas. The structure opens eastwards and (now) disappears under younger Karoo rocks; further east it is presumed to have been re-worked in the younger Mocambique belt.

The Soutpansberg lies adjacent to the Zoetfontein lineament. The presence of the thick volcanics indicates that the structure was an extensional rift at this time. The date at which this rift formed is enigmatic, but there seems logic in regarding it as a crustal adjustment subsequent to earlier strike-slip faulting on this fundamental fault.

The Soutpansberg is a deep, graben-like feature and was described as an aulacogen by Jansen (1975). Subsequently penetrative aulacogens were incorporated into plate tectonic concepts, and seen as failed arms cutting into cratons from triple junctions. It is difficult to provide specific evidence or or against the Soutpansberg rift in relation to plate tectonics, particularly given the superimposed Karoo and post-Karoo events.

The view taken here is that the Waterberg , Palapye red beds, and the Matsap red beds accumulated in structurally controlled terranes (strike-slip faulting for the Waterberg and Palapye; rifting for the Matsap); and that the movements concerned took place immediately or soon after uplift related to an Eburnian magmatic phase.

Less certain is the age of the Kheis orogeny, and the Soutpansberg aulacogen (Fig. 26). One very tentative speculation could even see the Soutpansberg as a crustal adjustment related to the Kheis. Further age determinations may shed more light on the sequence of events in these Early Proterozoic red beds; a systematic palaeomagmatic study would be of very considerable value.

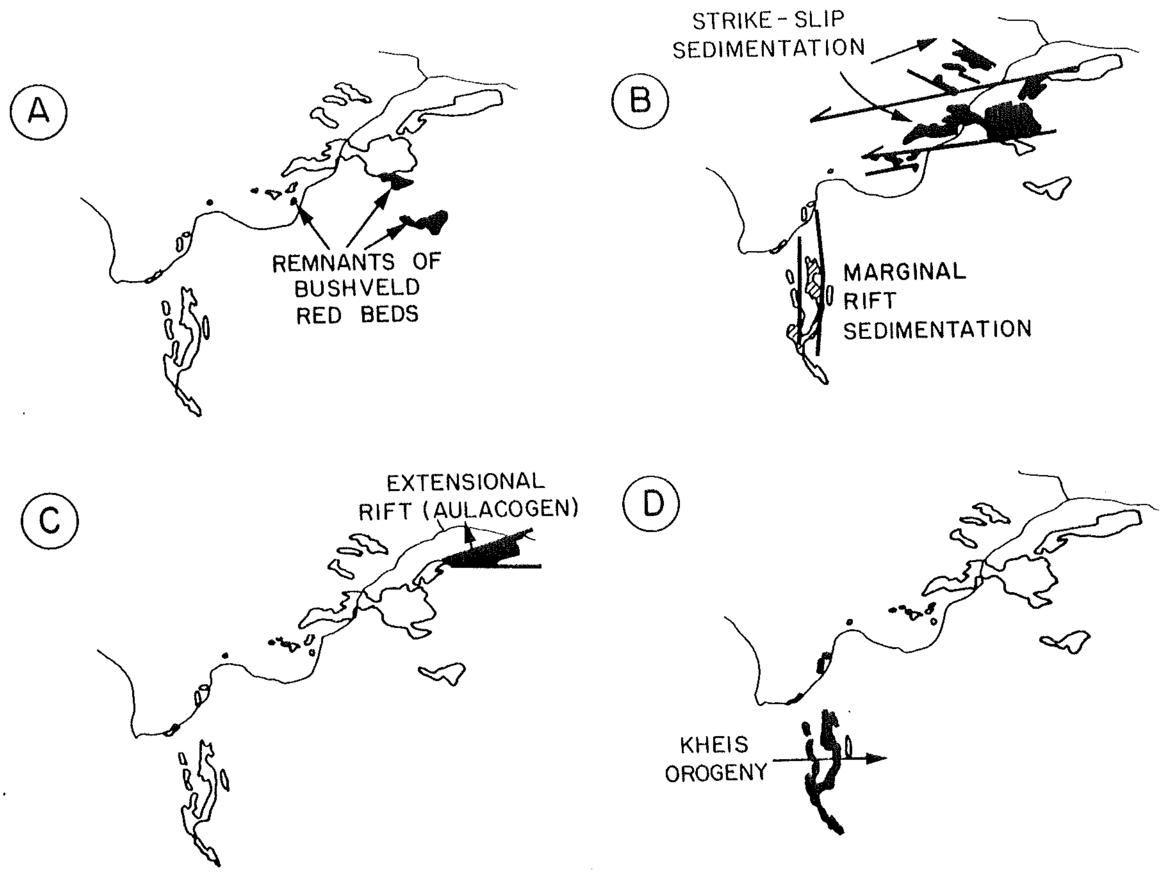

Fig. 26. A possible sequence in, and affecting, the Early Proterozoic red beds

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