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"Synthesis of Proterozoic Data as a Prerequisite"
for Tectonic and Thermal Modelling"

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"Synthesis of Proterozoic Depositional Data as a Pre-requisite for Tectonic and Thermal Modelling"

The proposal was to analyze, from published data, the tectonic environment of Proterozoic sedimentary deposition. We found that we could do even better because unexpectedly usable Archean data were available from the Kaapvaal province of South Africa and Archean data are more relevant to early crustal genesis. We analyzed these with a view to establishing whether and how the early crust (3.1 to 2.5 Ga) of one of the world's oldest continental objects behaved differently from continental crust in later times.

Our results showed that both the Pongola (3.1 Ga) and Ventersdorp rifts (2.6 Ga) behaved in no measurable way differently from younger rifts and our results are embodied in papers in the press in "Geodynamics" and "Tectonophysics" (preprints attached).

Our studies have also indicated that the Witwatersrand Basin (contrary to earlier interpretations) represents a typical foreland trough. This is discussed to some extent in our Tectonophysics paper, but resources were not sufficient to enable us to completely analyze the data and we plan to pursue this issue further, for example, by seeing how the Witwatersrand Basin flexed under the load of mountains on its northern flank.

Kevin Burke

Synthesis of Proterozoic Depositional Data as a Pre-requisite for Tectonic and Thermal Modelling

There are two basic varieties of intracontinental sedimentary basins: the rift type and the foreland basin type. Although these two basin types show contrasting tectonic, sedimentary and thermal histories they may show confusingly similar subsidence histories (Figure 1) (Bally, 1982). Both show rapid initial subsidence for fundamentally different reasons. In the rift model (McKenzie, 1978) this first phase of rapid subsidence is produced by isostatic compensation of thinned lithosphere, while in the foreland basin model it is caused by flexure of the lithosphere in response to peripheral loading. The second, slower phase of subsidence is produced by thermal relaxation of the lithosphere in the rift case, while in the foreland basin the slowing of subsidence is related to decreasing load as erosion in the peripheral mountains increases and compression declines.

Because the amount of thermally controlled subsidence in Precambrian sedimentary basins has been used to estimate the thickness of the early lithosphere (e.g. McKenzie et al., 1980) it is important to attempt to determine exactly what is the thermally controlled subsidence. For these reasons it is worth trying to synthesize all depositional and geologic data before tectonic, subsidence and thermal modelling for the basin in question is attempted.

Examples exist of thermal and subsidence modelling with incomplete consideration of depositional and tectonic data (Bickle and Eriksson, 1982; McKenzie et al., 1980). The subsidence history of several early Precambrian basins has been related to initial thinning of continental lithosphere followed by a more widespread slower subsidence due to thermal relaxation of this lithosphere. Examination of the regional geology indicates that several of these basins have not formed in rifts, but in foreland basins, invalidating the most simple thermal and tectonic models. It is proposed to analyze the depositional histories of these basins in terms of both basin models, so that more accurate thermal and tectonic models can be made for the early Precambrian lithosphere, yielding information on local geology and placing constraints on the secular evolution of the lithosphere.

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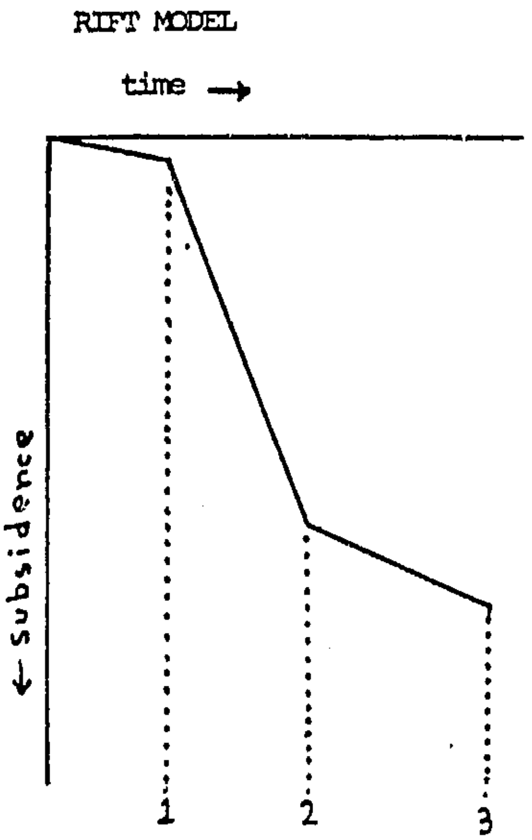
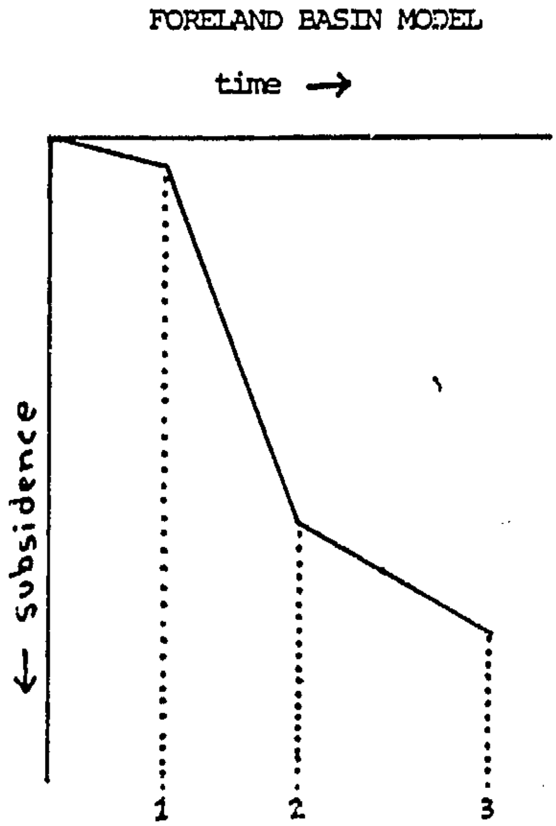
Figure 1





RIFT MODEL

1→2 fast initial phase caused by isostatic compensation of thinned lithosphere

2→3 slow phase caused by thermal recovery of lithosphere to original thickness

FORELAND BASIN MODEL

1→2 fast initial phase caused by rapid loading and flexure of lithosphere

2→3 slow phase caused by reduction of load of peripheral mountains

The Pongola Structure of Southeastern Africa: The World's Oldest Preserved Rift?

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Abstract

Rocks of the Pongola Supergroup form an elongate belt in the Archean Kaapvaal Craton of southern Africa. Because these rocks exhibit many features that are characteristic of rocks deposited in continental rifts, including rapid lateral variations in thickness and character of sediments, volcanic rocks that are bimodal in silica content, coarse, basement derived conglomerates and thick sequences of shallow water sedimentary facies associations, we suggest that the Pongola Supergroup was deposited in such a rift. The age of these rocks (approximately 3.0 Ga) makes the Pongola structure the world's oldest well-preserved rift so far recognized, and comparison of the Pongola Rift with other rifts formed more recently in earth history reveals striking similarities, suggesting that the processes that formed this rift were not significantly different from those that form continental rifts today.

Introduction

The Pongola Supergroup outcrops in a semi-continuous linear belt in South Africa and Swaziland. Figure 1 shows the location and present outcrop pattern of the Pongola Supergroup, as well as suggested (approximate) margins of the Pongola rift basin. The present extent of the basin is approximately 275 km x 100 km, implying a minimum depositional area of 27,500 square kilometers, although much of its original volume has been removed by erosion, destroyed by granitic intrusions and buried under later cover.

Figure 2 shows a geologic map of the Pongola Structure, with stratigraphic sections of the Supergroup. There are great variations in stratigraphic thickness of the Pongola sediments and a large proportion of volcanic rocks is preserved in the structure. These features suggest to us that the Pongola Supergroup rocks may have been deposited in a continental rift environment. Sawkins (1982), expanding on observations by earlier authors, for example Wilson (1968), Dewey and Bird (1971), and Burke (1977), has described characteristics of rocks deposited in ancient rifts and our modified list is as follows:

(1). Rift deposits not subsequently incorporated in an Atlantic-type continental margin would not (in general) be highly deformed or metamorphosed.

(2). Lateral thickness variations can occur over short distances and may exceed several kilometers in magnitude.

(3). Volcanic and intrusive igneous rocks are often bimodal in silica content.

(4). Faulted margins and linear rift trends might be discernable.

(5). Subaerial and subaquatic sediments might both occur.

(6). A thick, coarse basal clastic section, in many cases containing volcanics and related intrusive rocks, is overlain by a generally thinner, finer sedimentary section usually poor or lacking in volcanics.

We review the rocks of the Pongola structure in the light of these criteria in order to determine whether or not the Pongola Supergroup was deposited in a rift, and perhaps obtain a better understanding of the nature of the Archean continental rifting process.

Stratigraphy and Sedimentary Associations of the Pongola Supergroup

The Pongola Supergroup (Humphrey, 1912) has been divided into two subgroups; the Nsuzi (formerly Insuzi, S.A.C.S., 1980) which rests non-conformably on older Archean granites and greenstones, and the Mozaan Subgroup which oversteps successively lower units of the Nsuzi Subgroup towards the southeast (Button, 1981). Locally, in the northern outcrop area, a paleoregolith up to 7.5 meters thick is developed between the older granites and the Nsuzi Subgroup (Matthews and Scharrer, 1968). This soil is thought to have formed by *in situ* subaerial chemical weathering of the underlying biotite-microcline granite. The paleoregolith is now composed of sericite, quartz and vermiculite, with the amount of sericite diminishing towards the top which, along with the development of a crude stratification may suggest incipient reworking of the soil (Button *et al.*, 1981).

The Nsuzi Subgroup

The Nsuzi Subgroup consists predominantly of interfingering basalts and rhyolites, basaltic andesites and dacites, along with minor sedimentary intercalations. These volcanic rocks have been interpreted as showing both tholeiitic and calc-alkaline affinities (Tankard et al., 1982) and are bimodal in silica content. The Nsuzi Subgroup is best developed in the north along the Pongola River (Fig. 2) where it attains a thickness of 6100 (Fig. 3). It thins towards the south to a thickness of 1800 in the Wit Mfolozi inlier (Figs. 2 and 3). This substantial variation in along strike thickness can be attributed to differential subsidence during deposition, although the relatively thin nature of the Nsuzi Subgroup farther south in such localities as Nkandla (Figs. 2 and 3) may, at least in part, be attributable to erosion that may also have removed the entire Mozaan Subgroup, which is absent in this area. The base of the Nsuzi Subgroup is marked by an up to 850 thick series of rapidly alternating argillaceous, arenaceous and arkosic grits, quartz arenites, conglomerates and breccias (Button, 1981). The immature nature of some of these fluvial sands and the presence of feldspars indicates that they were derived, at least in part, from a local, uplifted, probably granitic source terrane. These sediments resemble fanglomerates and are derived from older Archean basement that perhaps formed shoulders to an active rift. Some of the more mature quartz-arenite and granule-conglomerate sediments that occur in the upper part of this lower unit (Matthews and Scharrer, 1968) may have been transported along the rift axis in an internal drainage network, as is common in the early stages of rift development (for example: as in the Awash valley of the Ethiopian Afar). Hobday and Von Brunn (1976) have demonstrated a dominantly granitic source for the arenites and argillites interbedded with the Nsuzi lavas and have suggested, as a result of applying a model by Klein (1971), that they were deposited by tidal currents with under a high tidal range. In some places the arenites wedge out against paleotopographic basement highs (Matthews, 1967) that may have been induced by syndepositional basement faulting (Button, 1981). Basalts of the Nsuzi Subgroup are of both tholeiitic and high magnesium varieties (unpublished analyses of Matthews, mentioned in Button, 1981). The basalts have been deuterically altered to an assemblage of turbid plagioclase, chlorite, epidote, calcite, leucoxene and opaque phases, and contain amygdules of chlorite, quartz, calcite, epidote and pyrite (Matthews, 1967). Rhyolitic horizons are abundant in the Nsuzi Subgroup with thicknesses of individual flows typically less than 10 meters. Flow top and gas streaming structures are common (Tankard et al., 1982). DuToit (1954) described abundant amygdaloidal andesites and tuffs from the lower Nsuzi valley. The Nsuzi volcanics have been interpreted as showing greenschist facies metamorphism and deuteric alteration and the mobility of those elements used to discriminate between calc-alkaline, and any other igneous rock series, suggests that the present composition of the Nsuzi volcanics does not necessarily closely correspond to their original chemistry. This conclusion is supported by the large scatter of Sr, Ba, and Fe_2O_3 on Larsen and crystallisation index diagrams (Tankard et al., 1982). The calc-alkaline affinities are in any case not strongly developed; on the basis of current data we do not regard the composition of the volcanics as incompatible with the interpretation of the Pongola structure as a rift.

A banded shale zone locally interbedded with the Nsuzi lavas (Fig. 3) was considered by Matthews (1967) to be a tidal flat deposit. The presence of these shallow water deposits about 4 km stratigraphically above the alluvial facies of the basal sand (see Vryheid-Piet Retief section, Fig. 3) indicates a minimum value for the subsidence that must have occurred in the Pongola basin. The tidal facies contains graded beds with small scale ripple drift, cross-laminated siltstone, shale, and lenticular channel deposits that display basal lag concentrates and fining upwards sequences. Another tidal or shallow marine deposit occurs in the Nsuzi Subgroup above another zone of bimodal volcanics. It contains carbonate cemented quartz arenites (Matthews, 1967), various other carbonates including aphanitic, oolitic and pisolitic dolomites, shales, tuffaceous sandstones, breccias and lenses of granule conglomerates. Also present are some shale clasts, partially silicified intraclast dolomites, sandstones with herringbone cross-laminations, fenestral textures in dolomites and domical stromatolites (Von Brunn and Mason, 1977; Mason and Von Brunn, 1977) which have been interpreted as representing a tidal flat environment of deposition. Other cycles with sand and mudstone are present throughout the Nsuzi Subgroup and have been interpreted to have formed from muddy tidal flats prograding over shallow water sands (Hobday and Von Brunn, 1976; Von Brunn and Hobday, 1976; Von Brunn, 1974).

The persistence of shallow water sedimentary environments throughout the eruption and deposition of the Nsuzi Subgroup volcanic rocks indicates that subsidence was penecontemporaneous with deposition. The subsidence was accomodated by normal faulting, which has been locally demonstrated to be of pre-Mozaan age in the Nkandla area (Matthews, 1967). We suggest that the Nsuzi Subgroup represents a rapid initial phase of subsidence and filling of the Pongola rift basin because it has characteristics similar to those of many more recent initial phase rift deposits, including: bimodal volcanic rocks, rapid variations in thickness and extent of sedimentary facies, syndepositional basement faulting and irregular basement topography, linear outcrop trends and (at least locally) faulted margins, thick sequences of shallow water facies deposits and immature basement derived fanglomerates.

The Mozaan Subgroup

The Mozaan Subgroup consists predominantly of alternating shales, quartz arenites, conglomerates and minor iron formation. The interaction of three sedimentary environments; braided alluvial plain, tidal flat and offshore shelf, has been discerned in the deposition of the Mozaan sequence (Watchorn, 1980) which is over 4600 m thick in its northern outcrops (Figs. 2 and 3; see also Matthews and Scharrer, 1968). Only 700 m are preserved towards the south, but much of the original thickness of the southern sections appears to have been removed by erosion, possibly during uplift associated with the Natal collision (see below).

The unconformity at the base of the Mozaan Subgroup truncates successively lower units of the Nsuzi Subgroup towards the southeast, and in general the area of Mozaan outcrop extends considerably farther east than the outcrop area of the Nsuzi Subgroup (Fig. 3). Tankard et al. (1982) have inferred that the Mozaan sediments may also extend a considerable distance westward of the area of Nsuzi outcrop (Fig. 4), but in that area they are now covered by Karoo rocks. The geometry of the Pongola Supergroup shows some resemblance to the Steers Head, or Texas Longhorn condition, a feature that typifies the two stage stretching, and thermal recovery development of a rift system (Burke, 1979; McKenzie, 1978). Bickle and Eriksson (1982) have suggested that the Mozaan sediments represent the thermal subsidence phase of the Pongola basin. Sediments deposited during the thermal subsidence phase are expected to differ from those of the initial stretching phase by slower accumulation (and subsidence) rates, a lack or scarcity of volcanics, and a larger area of deposition.

Pyritic conglomerates are developed near the base of the Mozaan group (Button, 1981) and similar conglomerates persist as well-defined thin horizons throughout the entire thickness of the subgroup. DuToit (1954) described conglomerates from the Denny Dalton area that are auriferous and uraniferous, also containing clasts of vein quartz, striped and black cherts, green quartzite, lava and pyrite. Near the base of the Mozaan Subgroup in Swaziland quartzite greatly exceeds shale in abundance, while throughout most of the subgroup they occur in nearly equal amounts (Hunter, 1963). Fourteen quartzite-shale cycles have been recognized from the northern outcrop areas (Button, 1981). Many of the shales are ferruginous and in some cases iron formations have developed, typically consisting of magnetite + tremolite + actinolite + quartz + chlorite with or without spessartine assemblages (Hunter, 1963). Banded iron formations consist of: (1) alternating iron oxide and red jasper mesobands (Strauss and DuPlessis, 1956); (2) oolites with chert nucleii surrounded by siderite (Beukes, 1973) or, (3) magnetite rich shales alternating with chlorite rich mesobands (Hunter, 1961; Beukes, 1973). The banded iron formations sometimes display intricate sedimentary slump fold structures (Beukes, 1973) which may indicate tectonic instability of the environment during deposition. Van Vuuren (1965) has noted some low grade manganese deposits associated with the iron rich parts of the Mozaan Subgroup, concluding that the manganese enrichment is most pronounced along faults.

In general, the shales of the lower Mozaan Subgroup typically containing an andalusite + pyrophyllite assemblage are more aluminous than are the upper, iron rich shales. Local contact metamorphic effects in the lower Mozaan Subgroup may have accentuated the apparent alumina concentration, locally forming andalusite - sericite or pyrophyllitic phyllites and sillimanite bearing quartzites (Beukes, 1973; Hunter, 1963).

Von Brunn and Hobday (1976) have attributed patterns of cyclic sedimentation in the Mozaan Subgroup to tidal flat deposition on an east-west trending shoreline. The orientation of the shoreline, which could possibly be related to the progradation of a shallow marine "delta" down an elongate (rift valley) depression, is indicated by the southeast directed trough cross-beds and the east-west preferred orientation of pebble long axes and ripple crests. The shallow tidal environment was hypothesized by Von Brunn and Hobday (1976) because of the presence of herringbone cross-strata, double crested and flat topped ripples. Cyclic sedimentation patterns between lower, middle and upper tidal flat sediments were noted by these authors. The lower tidal flat deposits consist of cross-laminated arenites that grade up into an alternating arenite-argillite mid tidal flat facies. The middle tidal flat deposits commence with flaser bedded arenites which are succeeded by wavy and lenticular bedded arenite-argillite alternations. These contain numerous dessication cracks, ripple marks and mud clasts indicative of tidal reworking. The upper tidal flat deposits consist of mudstones and graded siltstones displaying abundant mudclast microbreccias and dessication features. Locally associated with the upper tidal flat deposits are jaspillitic iron formations that may have accumulated in small episodically flooded depressions (Button, 1981).

The Usushwana Intrusive Suite

The Usushwana Intrusive Suite was named by Hunter (1950) and has recently been dated at 2871 +/- 30 Ma (Sm-Nd mineral/whole rock isochron, Hegner et al., 1984). In the Mhlambanyatsi area (Fig. 2) the Usushwana complex occurs as a northwest striking 40 km long x 6.5 km wide steep sided dike that parallels major faults in the granitic basement (Tankard et al., 1982; gravity data of Burley et al., 1970). This large dike is linked by a sheetlike mass that extends along the base of the Pongola Supergroup to a second northwest striking large steep sided dike complex that is also paralleled by major faults in the basement. Both the dikes and sheet contain xenoliths of Pongola rocks.

The Usushwana complex is composed of various mafic phases including hypersthene gabbro, quartz gabbro, olivine pyroxenite and serpentinised ultramafic rocks. A compositional layering is locally displayed. The earliest phase is a coarse grained olivine pyroxenite that was intruded by quartz gabbro, disrupting the compositional layering (Tankard _et al._, 1982). A thick granophyre layer commonly overlies the quartz gabbro, and can be observed northeast of Piet Retief (Fig. 2). Granitic dikes up to 20 meters wide containing plagioclase, quartz, epidote and biotite commonly fill joint planes in the gabbroic rocks. Although no economic grade mineral occurrences are known from the Usushwana complex, some disseminated chalcopyrite and pyrrhotite layers are fairly well developed near Mhlambanyatsi (Tankard _et al._, 1982).

Intrusives of the Usushwana complex also occur as numerous dikes and sills throughout the entire thickness of the Pongola Supergroup. Humphrey and Kriege (1932) recognized 10 diabase sills that intrude the Mozaan Subgroup preferentially along quartzite-shale contacts. Hunter (1963) notes that the amygdaloidal "basalt" in Swaziland is very similar to the amygdular diabase sills east of Piet Retief. Since these sills are part of the Usushwana complex, it is possible that the "basalt" in Swaziland may be the extrusive equivalent of the Usushwana complex, or alternatively it may be a large sill from which the cover has been eroded.

A postulated evolution for the Usushwana complex is (Hunter, 1970) that magmas appear to have preferentially intruded along faults that were formed during the initial rifting episode. Numerous diabase dikes lithologically similar to the main Usushwana complex intrude the basement complex, and are commonly oriented subparallel to the basin margins. The Usushwana intrusions have destroyed much of the original contact relationship between the granitic basement and the Pongola Supergroup. Tankard et al.,(1982) note the many striking similarities of the Usushwana complex to the Great Dike of Zimbabwe, which is considered to represent an outcropping example of "the axial dike" of a continental rift (Burke and Whiteman, 1973). If the Usushwana igneous complex is related to the Pongola structure, we may be seeing an episode of renewed rifting postdating a thermal subsidence event recorded in the Mozaan sequence. Such an event might make tectonic interpretation of stratigraphic thicknesses difficult.

Age of the Pongola Structure

Rocks of the Pongola Supergroup have recently yielded isotopic agges of 2940 +/- 22 Ma (U-Pb zircon concordia intercept, Hegner et al., 1984) and 2934 +/- 114 Ma (Sm-Nd whole rock isochron from basalt-rhyolite suite, Hegner et al., 1984). These dates are slightly younger than earlier reported ages of 3083 +/- 150 Ma (Rb-Sr whole rock method, Burger and Coertze, 1973) and 3030 +/- 90 Ma (U-Pb from zircons, recalculated from Burger and Coertze, 1973).

Granitic basement to the Pongola Supergroup has been reported to have yielded a Rb-Sr whole rock isochron of 2995 +/- 140 Ma (eg., Hunter, 1974 "quoting" data of Allsop et al., 1962). However, Allsop et al., (1962) ascribe this isochron to a Swaziland G4 granite that intrudes the Mozaan sediments in southwestern Swaziland (Hunter, 1957, 1963), providing a minimum rather than a basement age for the Pongola Supergroup. It is therefore likely that the basement to the Pongola Rift is of an older ag[illegible] [illegible]ssibly that of the "older" granites and gneisses in Swaziland which have yie[illegible]ed a Rb-Sr isochron of 3367

+/- 300 Ma (Allsop et al., 1962). Until future studies resolve relative ages of Pongola rocks and the Swaziland G4 granites, it is only possible to state that the time of initial Pongola deposition and rifting was approximately 3.0 Ga ago.

The Usushwana Intrusive Suite, which we suggest represents a renewed rifting event in the Pongola Structure, has yielded a Sm-Nd mineral/whole rock isochron of 2871 +/- 30 Ma (Hegner et al., 1984), which is in fair agreement with an earlier age determination of 2813 +/- 30 Ma (Rb-Sr whole rock, Davies et al., 1969). Ages quoted in this section have been recalculated (where necessary) using the decay constants of Steiger and Jager, 1977.

Metamorphism and Convergent Tectonics

The Pongola Supergroup has been metamorphosed to greenschist facies and is only mildly deformed, except in the south, where it was involved in at least two orogenic episodes. In the Vryheid-Piet Retief area just south of Swaziland Pongola beds are gently folded producing dips of 0 to 30 degrees and a dominant west-northwest fold trend direction (Button, 1981). Although another episode of gentle folding has produced mild basin and dome structures (DuToit, 1954; Button, 1981); each of the folding episodes was only associated with a minor amount of shortening and no complex structural relationships appear to have formed. In the Amsterdam and Mhlambanyatsi areas the general structural disposition of Pongola strata is a synclinal form, trending and plunging to the southeast, and intruded by steep sided mafic Usushwana dikes (Fig. 5).

Pongola rocks in the southern inliers around Nkandla are strongly folded and metamorphosed, with the amount of deformation increasing southward towards the Natal-Ntingwe thrust front, which marks the boundary between the Kaapvaal and Natal crustal provinces (Matthews, 1959). The Natal Province is an allochthonous terrane emplaced upon the southern portion of the Kaapvaal Craton as a series of nappes at a suture zone about 1.0 Ga years ago (Matthews, 1981). The Tugela and Mfongosi suites of the Natal Province represent, respectively, the lower and upper parts of an ophiolite suite obducted northward onto the Kaapvaal Craton along the Makasana basal decollement zone (Tankard et al., 1982). The truncated nature of the rocks of the Pongola structure at the Natal thrust front suggests that the Pongola rift originally continued farther southward, and may now be partly buried under the Natal allochthons.

Discussion

The Pongola Supergroup shows many features that are typical and diagnostic of rocks deposited in continental rifts. Its age ranges from approximately 3.0 to 2.87 Ga, making it the world's oldest well-preserved rift so far recognized. This identification extends the occurrence of rifts considerably further back in time than recognised by Burke et al. (1976) and Burke (1977) and suggests that it is more likely that the scarcity of old rifts is an accident of preservation rather than indicative of a radical secular change in planetary tectonic behavior from the Archean to younger times. Linear outcrop trends along with locally faulted margins enable the original width of the rift to be roughly estimated at 50-70 km. In numerous places where the basin margins are not observed to be faulted, the contact between Pongola strata and older basement is the site of mafic intrusions (Usushwana Suite). These dikes (and sills) have been preferentially intruded in places that are inferred to have been zones of weakness developed during the initial stages of Pongola rifting. Bickle and Eriksson (1982) have estimated a stretching factor of 1.4 to 2.2 for the Pongola basin using thicknesses for the sediments and volcanics deposited during postulated initial stretching and thermal subsidence phases. Intrusion of the Usushawana rocks late in the rifting episode indicates additional extension. Because of difficulties in quantifying the amount of Ushushwana related extension and because the lithospheric thickness before any Pongola rifting remains unknown we prefer not to make a numerical estimate of extension or stretching. However, the overall thicknesses and size of the Pongola structure are quite typical of Phanerozic continental rifts, for which a stretching factor of about 2 is often appropriate. The evidence from the Pongola rift suggests that Archean rifting was not significantly different from continental rifting later in the earth's history. There is persuasive evidence of an ocean closing event about 1 Ga ago in the Namaqualand orogen at the southern end of the Pongola rift, but it is not yet clear whether the initiation of the Atlantic-type continental margin on the southern edge of the Kaapval Craton (which forms the northern side of the Namaqua orogen) was contemporary with or much younger than the Pongola rifting event.

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Figure Captions

Figure 1. Southern Africa showing the location of Pongola Group outcrops and the suggested shape of the 3.1 Ga old Pongola Rift in South Africa and Swaziland.

Figure 2. Geologic map of the Pongola structure with stratigraphic sections from various localities. The rift is bounded by thrusts in the south and splits in two around basement horsts in the north (based in part on the work of Button, 1981).

Figure 3. Stratigraphic columns for the Pongola Group (modified from Button, 1981). Inset shows the approximate locations of these columns, which show features typical of rift stratigraphy.

Figure 4. Sketch map showing the distribution of Nsuzi sediments in the Pongola structure which are interpreted as rift-fill (Fig. 4a). Also shown are our suggested rift margins, which probably joined an Atlantic-type continental margin.
The Mozaan sediments are distributed more widely than the Nsuzi Group rocks and have been interpreted as thermal subsidence phase deposits (Fig. 4b). (Based on work of Tankard _et al._, 1982).

Figure 5. Sketch cross-sections of the Pongola rift and Usushwana intrusive suite (a) S.W. of and (b) N.W. of Mhlambanyatsi in Swaziland modified after Hunter (1970).

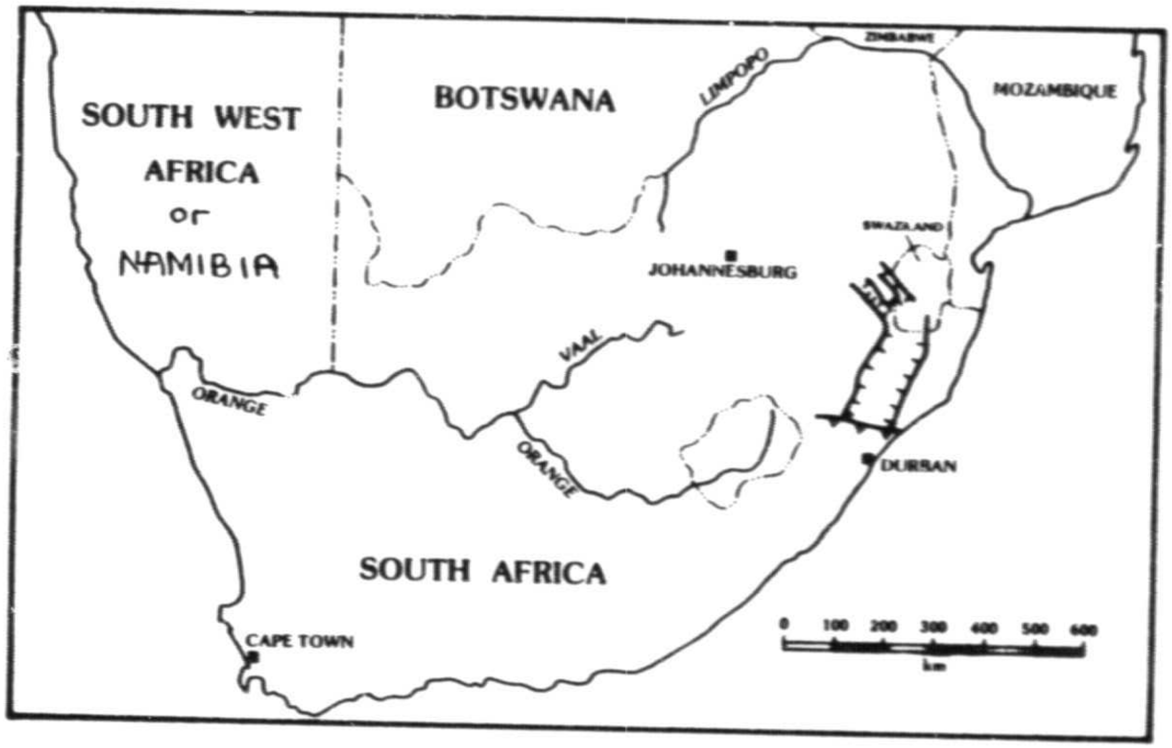
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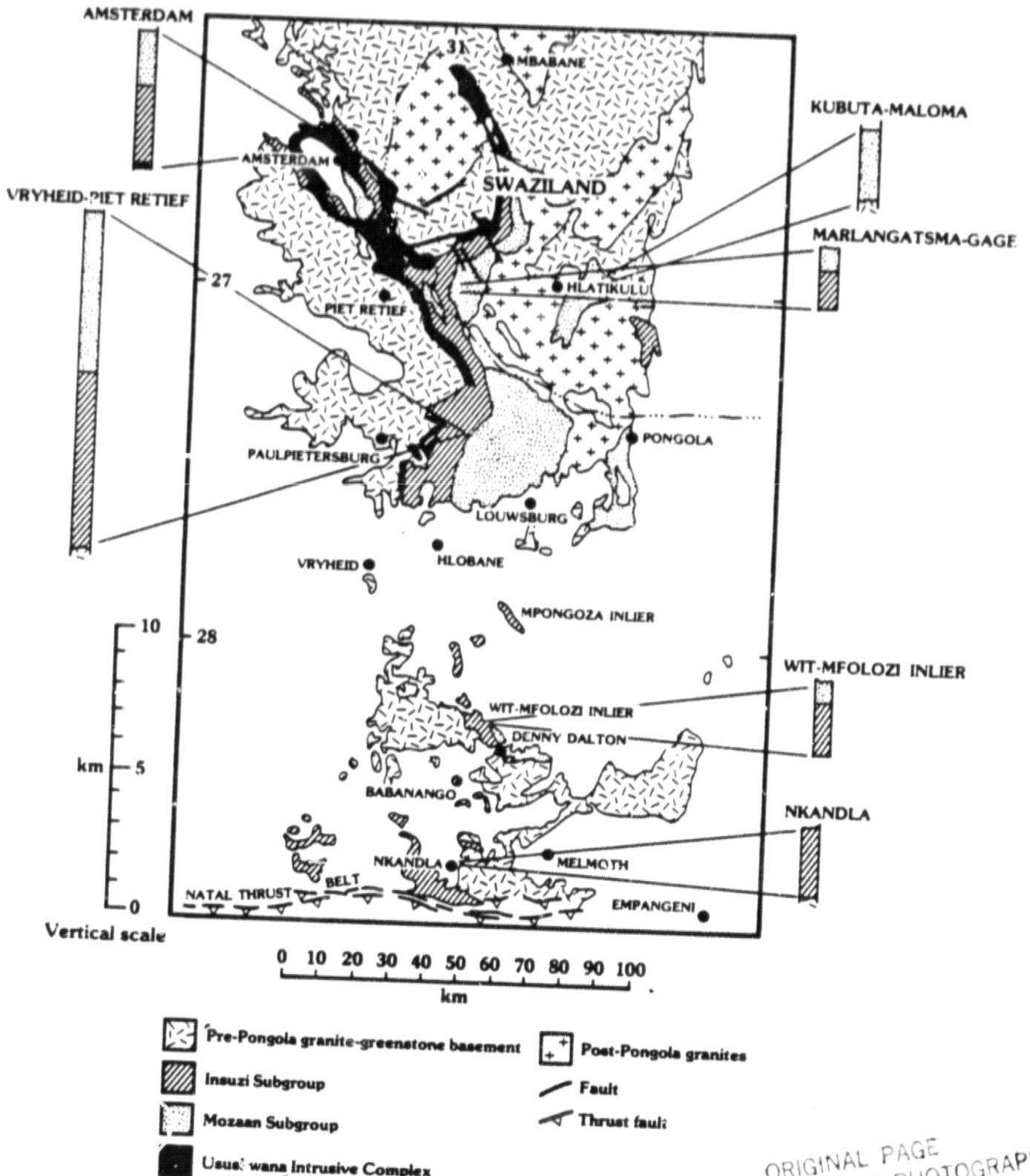
Fig 1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LOUWSBURG
HLOBANE
VRYHEID
MPONGOZA INLIER
28
WIT-MFOLOZI INLIER
WIT-MFOLOZI INLIER
DENNY DALTON
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NKANDLA
NKANDLA
MELMOTH
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km
Pre-Pongola granite-greenstone basement
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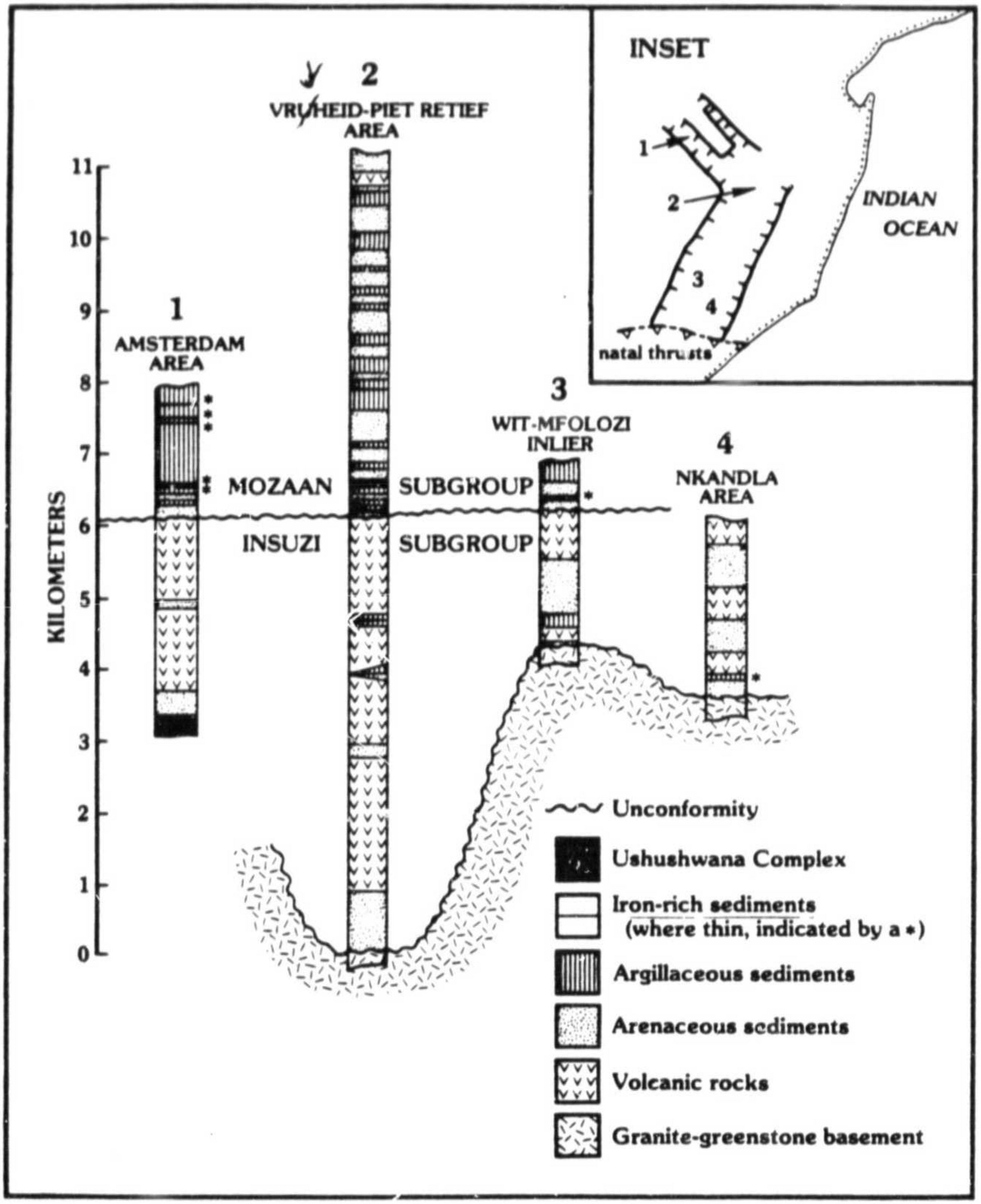
Mozaan Subgroup
Usushwana Intrusive Complex
Post-Pongola granites
Fault
Thrust fault

Fig 2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MOZAAN SUBGROUP
INSUZI SUBGROUP
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Unconformity
Ushushwana Complex
Iron-rich sediments (where thin, indicated by a *)
Argillaceous sediments
Arenaceous sediments
Volcanic rocks
Granite-greenstone basement

Fig 3

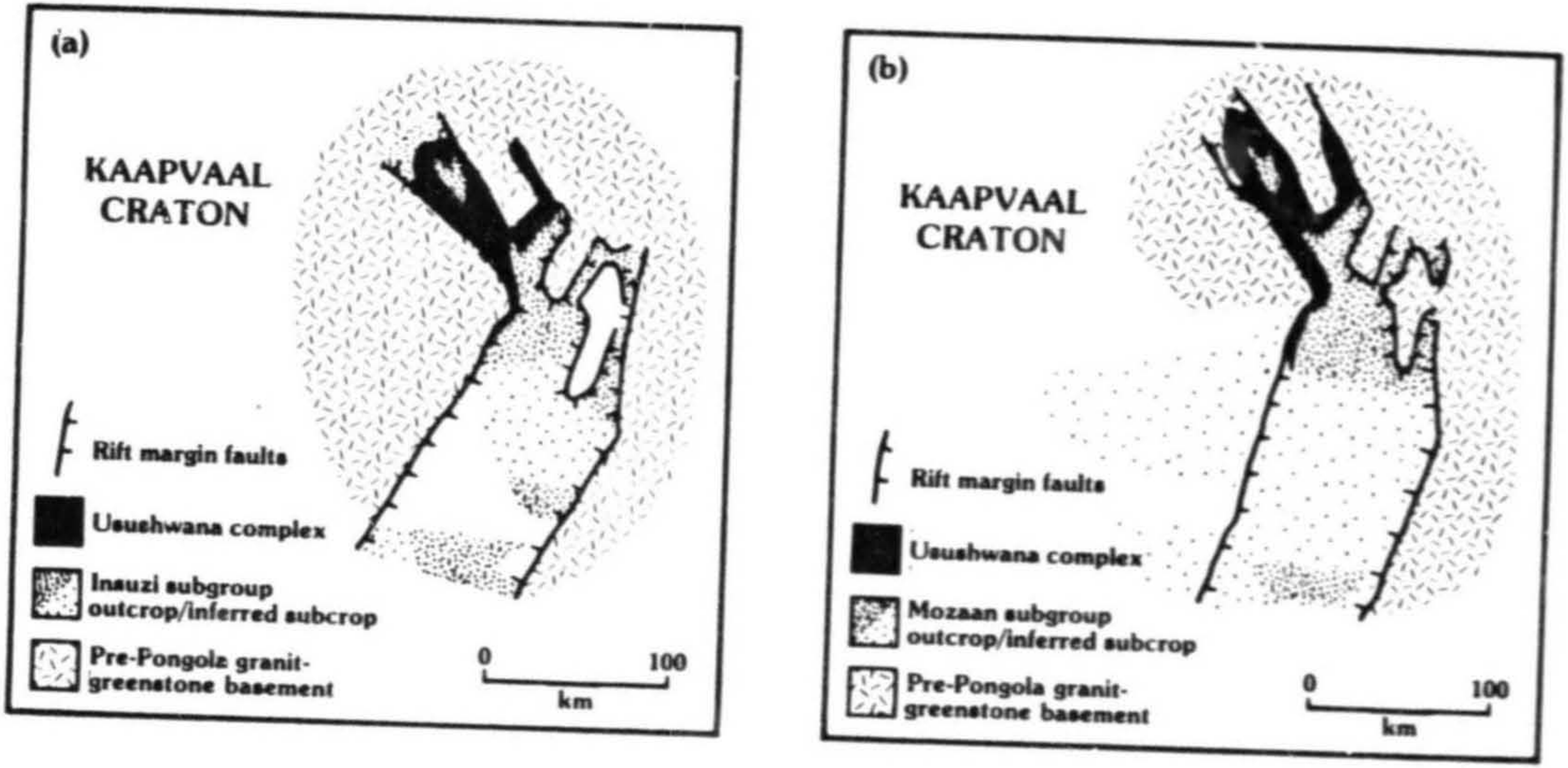
(a)
KAAPVAAL
CRATON
Rift margin faults
Usushwana complex
Insuzi subgroup
outcrop/inferred subcrop
Pre-Pongola granit-
greenstone basement
0
100
km
(b)
KAAPVAAL
CRATON
Rift margin faults
Usushwana complex
Mozaan subgroup
outcrop/inferred subcrop
Pre-Pongola granit-
greenstone basement
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Fig 4

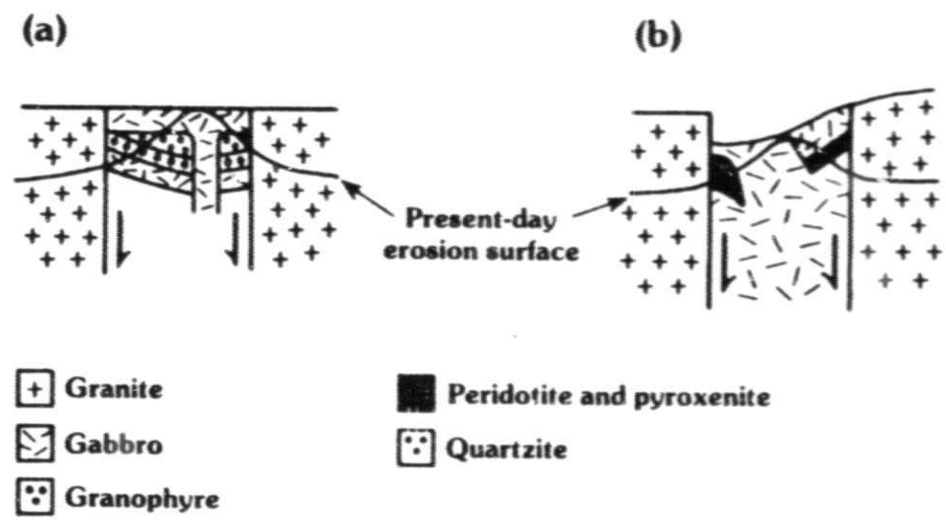

Fig 5

BLACK AND WHITE PHOTOGRAPH

IS THE VENTERSDORP RIFT SYSTEM OF SOUTHERN AFRICA RELATED TO A CONTINENTAL COLLISION BETWEEN THE KAAPVAAL AND ZIMBABWE CRATONS AT 2.64 Ga AGO?

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ABSTRACT

Rocks of the Ventersdorp Supergroup were deposited in a system of northeast trending grabens on the Kaapvaal Craton approximately 2.64 Ga ago contemporary with a continental collision between the Kaapvaal and Zimbabwe Cratons. We suggest that it was this collision that initiated the Ventersdorp rifting. Individual grabens strike at high angles toward the continental collision zone now exposed in the Limpopo Province where late orogenic left lateral strike-slip faulting and anatectic granites are recognized. We relate the Ventersdorp rift province to extension in the Kaapvaal Craton associated with the collision, and see some analogy with such rifts as the Shansi and Baikal Systems associated with the current India/Asia continental collision.

Introduction

Sedimentary and volcanic rocks of the Ventersdorp Supergroup were deposited in north to northeast trending grabens that define a broad elongate belt on the Kaapvaal Craton of southern Africa (Figure 1). Ventersdorp rifts are estimated to extend over 200,000 km^2, although much of this is covered by more recent strata (Tankard et al., 1982).

The Ventersdorp Supergroup rests conformably on Witwatersrand strata at such localities on the Rand as Benoni, Evander and in the Welkom Goldfields (DuToit, 1926, p. 73; S.A.C.S., 1980; Tankard et al., 1982). Elsewhere, for example along the northwestern margin of the Witwatersrand basin between Krugersdorp and Klerksdorp there is up to 5° of angular discordance between the two Supergroups, while in still other localities Ventersdorp strata onlap older units and in places rest on the Archean granite-greenstone basement. The Ventersdorp Supergroup exhibits rapid lateral facies variations, volcanic rocks that are bimodal in silica content, irregular basement topography and linear faulted basin margins. For these reasons we suggest that the Supergroup may have been deposited in a continental rift environment, and we here attempt to characterize the nature of this Archean continental rifting event by a review of the Ventersdorp history.

Stratigraphy

In an analysis of outcrop and borehole data Winter (1976) devised a regionally correlatable lithostratigraphic classification of the Ventersdorp

Supergroup. This work eliminated much previous confusion about the complex Ventersdorp stratigraphy, usually attributable to lack of good exposure, interfingering of lithologies, pinching out of units and progressive onlap of the Ventersdorp strata. Winter (1976) divided the nearly 8 km thickness of the Ventersdorp Supergroup into the Klipriviersberg Group, the Platberg Group and the overlying Bothaville and Allanridge Formations (Pniel Series of DuToit, 1907, p. 93). [see figure 2].

Klipriviersberg Group

Rocks of the Klipriviersberg Group form the lowest part of the Ventersdorp Supergroup. The Group is a maximum of 1830 meters thick (S.A.C.S., 1980), and consists of a monotonous sequence of alkali-rich tholeiitic flood basalts (Tankard et al., 1982) [Table 1]. This group is correlated with the lower volcanic stage of the Orange Free State Goldfield (Winter, 1976), and is also well developed in the Cape Province, Transvaal and southeastern Botswana. From base to top the Klipriviersberg Group consists of the following six formations (Winter, 1976):

(1) Westonaria Formation, with the local development of the Meredale Member at its base.

(2) Alberton Porphyry Formation

(3) Orkney Formation

(4) Jeannette Agglomerate Formation

(5) Lorainne Formation

(6) Edenville Formation

Meredale Member

The Meredale Member of the Klipriviersberg Group is composed of a sequence of mafic komatiitic high Mg basalt flows alternating with more normal basaltic horizons that also have high Mg contents. The Meredale Member is well developed at the base of the Ventersdorp Supergroup in the vicinity of Johannesberg, while in other areas "talcose tuffs" thought to represent paleosols developed from mafic rocks (Westonaria Formation) may be lateral equivalents of the Meredale Member (Cawthorn et al., 1979). The mineralogy of the basalts (summarized from McIver et al., 1981), consists essentially of stubby tremolite-actinolite laths in a chloritic groundmass perhaps representing a devitrified glass. Accessory minerals include chlorite, clinozoisite, epidote and plagioclase microphenocrysts. Some acicular tremolite-actinolite needles in the groundmass have been suggested to have replaced original spinifex textured olivine or pyroxene needles, formed by the supercooling of komatiitic liquids (McIver et al., 1981; see see their Figure 5.2).

Upper Klipriviersberg Group

Since the six Formations that comprise the rest of the Klipriviersberg Group differ markedly only in their textures lavas of the Westonaria, Alberton Porphyry, Orkney, Jeannette Agglomerate, Lorainne and Edenville Formations are here discussed together. These lava flows are commonly wedge shaped and of uneven thicknesses related to filling surface irregularities on underlying Witwatersrand strata. The Westonaria Formation is developed only in the deepest grabens, and overlying formations progressively onlap the rift margins. The remarkable chemical homogeneity (both lateral and vertical) of these basalts, coupled with the scarcity of interbedded sedimentary horizons suggests rapid effusion (Tyler, 1979). Faulting was contemporaneous with extrusion, as Tyler (1979) demonstrated in west central Transvaal for an thickened section of Klipriviersberg lavas interbedded with boulder conglomerate and basaltic breccia. Visser et al. (1976) describe angular granite-pebble conglomerates and basement relief of up to 800 meters from the lower Ventersdorp Supergroup (Figure 3). In many areas of northern Cape Province the clastic wedges in these fault troughs were progressively rotated by up to 80° before deposition of the Bothaville Formation (Visser et al., 1975-76). Steepening of talus deposits next to basement horsts is consistent with the contention that fault activity was contemporaneous with the deposition of the lower Ventersdorp Supergroup. The normal faults formed during this extensional episode may have served to localize eruptive centers of the Klipriviersberg lavas, as the flows are often thicker at, or confined to regions near these faults (Figure 6). Crockett (1971) has demonstrated that these flows were, at least locally, fed from gas blasted fissures.

Individual basalt flows within the Klipriviersberg Group may be porphyritic, amygdular (chert filled), or massive in texture although local agglomeratic horizons have been noted (S.A.C.S., 1980). Flow top breccias are commonly developed and volcaniclastic beds become increasingly more abundant towards the top of the group (Crockett, 1971). Mineralogically, the basalts are composed of a felted mass of calcic plagioclase microlites and subhedral clinopyroxene prisms, commonly altered to chlorite. Small magnetite grains are scattered throughout, and accessory minerals include quartz, chlorite, calcite and pyrite (Tyler, 1979). Some of the vugs in the basalt are filled with a mixture of quartz, calcite and radiating needles of fibrous pumpellyite, mantled by a cherty material containing chlorite and magnetite inclusions (Tyler, 1979).

The Klipriviersberg lavas show a remarkable chemical homogeneity both through their 2000 meter thickness and over a huge areal extent (> 155,000 km^2). This, and the scarcity of explosive volcanics led Tyler (1979) to classify the Klipriviersberg lavas as flood basalts, an attribution which is supported by both chemical analyses (Table 1) and field relations. When compared to the "average" continental tholeiite, basalts of the Klipriviersberg Group appear to be enriched in K_2O and depleted in Al_2O_3, although the K_2O enrichment may be attributed, at least in part, to later greenschist facies metamorphism. Wyatt (1976), noting the general low alumina, high total iron, intermediate Mg and low to intermediate silica content of the upper Klipriviersberg lavas described them as alkali-rich tholeiitic basalts.

Platberg Group

The Platberg Group represents an upward fining sequence of sediments and lavas, consisting of the Kameeldoorns, Makwassie and Rietgat Formations (Figure 2). The Platberg Group rests conformably on the Klipriviersberg lavas in the central parts of the Ventersdorp Basin (Winter, 1976); elsewhere it lies on steeply dipping Witwatersrand strata or pre-Witwatersrand rocks.

The basal Kameeldoorns Formation (regarded as equivalent to the Klippan Formation of the Welkom Goldfield) was deposited mainly in fault troughs (S.A.C.S., 1980) and displays rapid lateral facies changes and complex interfingering of lithologies. The distribution and character of the Kameeldoorns sediments was determined by proximity to active (normal) fault scarps. Massive boulder conglomerates, talus breccias and debris flow deposits are concentrated near horst blocks, while lithologies typically grade laterally through arkoses and siltstones into lacustrine shales and carbonates, in some places less than 10 km away from the fault scarp (Winter, 1976). The boulder and cobble conglomerates range from chaotic non-stratified clast supported wedges to matrix supported debris flow deposits containing subrounded to rounded clasts that frequently reach diameters of one meter. Clast and matrix lithologies strongly reflect the nature of the adjacent horst block; for example, in the New Kameeldoorns Graben conglomerates consist predominantly of fragments of Klipriviersberg volcanics clearly eroded from the adjacent horst (Winter, 1976). In areas where the adjacent horst block contains no earlier Ventersdorp strata the Kameeldoorns clast lithologies may also be correlated with the source area; for instance, near the Wesselbron "granitic arch" the conglomerates are arkosic, containing pebbles of mafic metavolcanics, graywacke and other pre-Ventersdorp metamorphic rock fragments (Hatch, 1903; Visser et al., 1976). In the vicinity of the Taljaardstam Fault conglomerates contain an increasing proportion of quartz arenite and shale pebbles towards the top of the Formation, reflecting erosion through Ventersdorp lavas and into Witwatersrand strata in the adjacent horst block (Buck, 1980). All of these features are characteristic of modern day fanglomerate, talus scree and alluvial fan deposits, formed as wedges prograding into adjacent subsiding grabens.

Shales of the Kameeldoorns Formation are black, finely laminated and occur in siltstone-mudstone upward fining couplets generally less than 2 cm thick. They commonly display dessication cracks, ripple cross-laminations and shale intraclasts (Buck, 1980). Some of these lacustrine deposits contain laterally linked, low relief (< 5 cm) hemispheric stromatolites. Rare thick cross- and horizontally laminated oolite beds up to two meters thick also occur in basin centers and on the sides of half grabens distal from fault scarps (Buck, 1980; and Figure 4).

The Makwassie (Quartz Porphyry) Formation overlaps the Kameeldoorns Formation, progressively lapping onto the Klipriviersberg Group, Witwatersrand Supergroup and older Archean terranes. Although locally absent the formation has been described as over 2100 meters thick in the western Transvaal. However, the recognition of a thick pseudotachylite (?) zone separating similar lithologies (Winter, 1976, p. 42) suggests at least local repetition. Makwassie lithologies include abundant quartz porphyries, andesites, dacites, felsic ash flow tuffs, green-gray feldspar porphyritic lavas and non-porphyritic volcanics interbedded with sparse volcaniclastic conglomerates, coarse sands and flinty shales (Winter, 1976; Tankard et al., 1982), with the proportion of quartz porphyry to other lava types decreasing away from eruptive centers. Most of the eruptions that produced the Makwassie Formation were subaerial, as indicated by the numerous flow structures, agglomerates and graded tuffs, although some possible pillow lavas (Winter, 1976) suggest that subaqueous volcanism occurred as well. Massive to granular porphyritic and variolitic felsites of rhyolite composition occur interbedded with volcaniclastic sediments over a large area in the Transvaal State (Tyler, 1979; Tankard et al., 1982). Sometimes these pyroclastic units that are correlated with the Makwassie Formation display an extensive layer with 15 cm lithic (porphyritic felsite) clasts, fragments of older lithic units and bands of bipolar fusiform lapilli (Tyler, 1979). Agglomerates of the Makwassie Formation are of two kinds; the first displays no stratification and consists essentially of ill-sorted angular blocks of siliceous volcanic material contained in a fine-grained matrix. This type is considered to have originated from an explosive vent (Tyler, 1979). The second type of agglomerate is composed of "chaotic, subrounded heterolithologic clasts in a schistose shaley matrix" (Tyler, 1979, p. 142), and may represent a laharic breccia. Crockett (1971) has noted that the explosive agglomerates were confined to volcanic vents (usually less than 10 meters in diameter and cutting through older formations) during the early stages of Makwassie volcanism, and that the style of volcanic activity changed from these early vertical eruptions to horizontally directed eruptions in the later stages of Makwassie effusion, which spread agglomerates over larger areas. Geochemically, the Makwassie felsites roughly correspond to a rhyoltic composition (Table 1) except that they appear to be highly potassic and relatively deficient in Na_2O, MgO and CaO. The apparent Na_2O deficiency has been considered a result of alteration (Tyler, 1979).

Sedimentary rocks of the Makwassie Formation are typically very immature poorly sorted packages of arkosic or volcaniclastic debris. In the western Transvaal a 300 meter thick lens of coarse, ill-sorted, poorly packed volcaniclastic conglomerates and sandstones accumulated as an upward fining series of channel-lag and point-bar deposits (Tyler, 1979).

The top of the Makwassie Formation grades into the basal (Garfield) member of the mixed volcanosedimentary Rietgat Formation, which may be up to 1300 meters thick (S.A.C.S., 1980). Greenish-gray "andesitic" to dacitic lavas of the Rietgat Formation are usually porphyritic to extremely porphyritic, with short and lath-like feldspar phenocrysts (Winter, 1976). Many individual flows show flattened pumiceous lava tops and contain red chakedony nodules. The proportion of sediments to lavas increases stratigraphically upward toward the overlying Bothaville Formation, which is entirely sedimentary. Sediments of the Rietgat Formation are similiar to those in the Kameeldoorns and Makwassie Formations, except that the Rietgat Formation does not contain great thicknesses of coarse clastic debris (Winter, 1976). Thin breccia, talus scree and debris flow deposits commonly flank horst blocks and contain identifiable well rounded cobbles of Withwatersrand strata (Buck, 1980). Where the horsts are granitic the sediments are arkosic (Visser et al., 1976). These deposits grade laterally into graywackes, sandstones and impure lacustrine limestones that locally contain small stromatolites (Buck, 1980). Episodes of non-deposition, minor erosion and oxidation are recorded in a strong alteration of many lava flow tops, especially stratigraphically high in the formation (Winter, 1976).

In summary, sporadic ephemeral fluvial activity rapidly transported and deposited Rietgat sediments into coalescing alluvial fans that filled the basins and valleys developed by normal faulting of older strata. The formation progressively onlapped these horst blocks as they became deeply eroded (Buck, 1980), and the topographic difference between them and the grabens decreased (Figure 5), implying the cessation or waning of fault activity during deposition of the Rietgat Formation.

Bothaville Formation

The Bothaville Formation is approximately 430 meters thick and rests conformably on the Rietgat Formation in its type area (Winter, 1976; S.A.C.S., 1980). However, in numerous other localities sediments of the Bothaville Formation unconformably onlap older Ventersdorp strata, Witwatersrand rocks or the Archean basement. The Bothaville Formation contains the most mature sediments of the Ventersdorp succession and differs markedly from rocks of the underlying formations as it is not characterized by the rapid facies changes typical of the Platberg Group and occurs in a fairly constant thickness over a relatively large area. It is interpreted as having been deposited on gently undulating plains.

A poorly sorted conglomerate containing well rounded pebbles of jasper, chert and (Makwassie) quartz porphyry lies at the base of the Bothaville Formation. The conglomerate is interbedded with quartzites, shales and stromatolitic limestone which give way to a upward fining sequence containing graded beds of siltstone, mudstone and sandstone of subarkose to subgraywacke composition (Winter, 1976). Sand bodies throughout the formation are lenticular in form, displaying shale drapes, shale intraclasts and low angle cross-stratification indicating unimodal to trimodal current directions (Visser et al., 1976). The shales are usually planar bedded, often displaying dessication cracks and local stromatolites (Buck, 1980). This sequence contains another conglomeratic horizon near the top of the Formation.

The Bothaville Formation is interpreted to have been deposited by a southward flowing (present coordinates) braided and meandering stream network on an alluvial plain experiencing a marine transgression from the south and southeast (Buck, 1980; Tyler, 1979; Visser et al., 1976). The Bothaville sands appear to have been deposited and reworked in beach and shallow neritic environments, accumulating by the landward migration of sand bars. Variations in the amount and maturity of clastic debris shed into the Bothaville Basin could reflect shifting fan loci and variable river discharge related to climatic variation (Buck, 1980; Tankard et al., 1982).

Allanridge Formation

Extrusion of the Allanridge lavas abruptly terminated deposition of the Bothaville sediments. The Allanridge Formation which is up to 740 meters thick (S.A.C.S., 1980) was laid down conformably on top of and overlaps the outcrop limits of the Bothaville Formation. It consists of a monotonous sequence of lavas ranging from tholeiitic flood basalts to andesitic flows (Table 1). The flows typically have a dark green-gray color, numerous amygdules, and are very fine grained. Porphyritic zones with acicular feldspar phenocrysts alternate with non-porphyritic zones (Winter, 1976) and the presence of pillow lavas in the lower part of the Formation demonstrate the subaqueous nature of these flows (Visser et al., 1976). Although a few early eruptions deposited relatively minor volcanic breccias and tuffs most of the volcanism was probably of a rapid and quiet effusive nature, as suggested by the lack of interbedded sedimentary horizons, the thickness (750 meters) and uniformity of flows over large areas and the scarcity of pyroclastic deposits. The flows were most likely fissure fed (Vissure et al., 1976) and are quite similar to the flood basalts of the

Kliprivlersberg Group.

TECTONIC SIGNIFICANCE OF VENTERSDORP VOLCANICS AND SEDIMENTS

Nature of the Klipriviersberg Group

Klipriviersberg lavas demonstrate remarkable chemical homogeneity, both laterally and vertically. Their substantial thickness (1830 meters), wide distribution and bulk composition all strongly resemble more recent continental flood basalts. The Klipriviersberg lavas bear many similarities to alkalic magmas erupted in continental areas that have experienced lithospheric extension. Examples include the East African Rifts, Rhine Graben of western Germany, the Oslo Graben of Norway and the Cenozoic Basin and Range Province of the western United States (Basaltic Volcanism Study Project 1981). Continental basalts associated with rifting often have a quartz-tholeiite composition, which is consistent with the chemistry of the Klipriviersberg lavas (Table 1).

The komatiitic nature of the lower Meredale Member suggests that these lavas were derived directly from large amounts of partial melting in the mantle. The vast thickness and scarcity of sedimentary horizons in the Klipriviersberg Group is suggestive of considerable extension. Evidence was cited above demonstrating that extensional faulting did accompany extrusion of the Klipriviersberg flood basalts and it is therefore suggested that they represent a mantle-derived early rifting phase of volcanism in the Ventersdorp Province. The upper Klipriviersberg tholeiites could have formed by fractionation of liquids having a composition similar to the basal Meredale Member, as indicated by A-F-M plots of Ventersdorp volcanics and variation diagrams in McIver et al., (1981).

Nature of the Platberg Group

The Platberg Group consists essentially of silicic pyroclastic and volcanic rocks interdigitated with very immature locally derived sediments. These were rapidly deposited in linear fault troughs that evidently became active near the end of the extrusion of the Klipriviersberg flood basalts. Although some of these basin margin faults may have been active throughout the extrusion of the Klipriviersberg lavas, the enormous thickness and abundance of coarse talus deposits in the Platberg Group suggests increased faulting during this time. The faulting appears to have virtually ceased by the end of Platberg deposition, as the elevation of the horst blocks was diminished by erosion, and the coarse locally derived sediments and pyroclastics of the Platberg Group gradually graded up into more mature recycled sediments, ultimately forming the entirely sedimentary Bothaville Formation. Sediments of the Platberg Group as a whole fine upward, also suggesting that most of the faulting activity occurred early. Sedimentary facies similar to those of the Platberg Group, typically grading from fanglomerates to fine lacustrine deposits over short distances, are characteristic of many rift valley provinces of today (Kinsman, 1975).

We suggest that the Platberg Group was deposited in grabens during the rapid initial subsidence phase of the Ventersdorp Rift System. The nature of volcanism in the Ventersdorp Rift Province changed from quartz-tholeiitic flood lava extrusion to silicic quartz porphyry and ignimbrite eruption (Makwassie

Formation) at a time corresponding to an episode of increased faulting. A similar change in the style of volcanism with the onset of graben formation (tholeiitic basalt to quartz porphyry rhyolites) in the Oslo Graben of Norway was noted by Williams (1982).

Nature of the Bothaville Formation

In comparison with other formations of the Ventersdorp Supergroup, sediments of the Bothaville Formation are highly mature and were deposited over relatively subdued topography. The Bothaville Formation lacks volcanics, drapes over the outcrop limits of older formations and its deposition was not influenced by faulting. The geometry and nature of the Bothaville Formation therefore conforms to the upper part of the "Steer's Head Condition" (Burke, 1979) characteristic of the thermal of subsidence in rifts (Figure 6). Sediments formed during this phase are a response to the recovery and sinking of mantle isotherms, with a concomitant conductive heat loss and increase in density of the lithosphere. Subsidence (and hence, deposition) during this phase is likely to be slower than during the rapid initial subsidence phase (McKenzie, 1978), favoring sediments reworking, and therefore, increased maturity compared to sediments deposited during the initial or stretching phase of subsidence. Sediments of the Rietgat Formation become interbedded with fewer volcanics and become more mature upward as they grade into the overlying Bothaville Formation, suggesting that the transition from initial to thermal subsidence was gradual. It also appears that the deposition of these thermal subsidence phase sediments was abruptly terminated by the extrusion of the Allanridge lavas.

Nature of the Allanridge Formation

Lavas of the Allanridge Formation were quietly extruded in both subaerial and subaquatic environments. Their chemistry, morphology and wide distribution suggest that they are continental flood lavas. There is a striking similarity between the Klipriviersberg and Allanridge lavas, with the exception that the Allanridge basalts usually contain more feldspar phenocrysts of andesine composition (Visser et al., 1976; Winter, 1976). The Allanridge Formation appears to represents a renewed rifting event in the Ventersdorp Rift Province which disrupted the deposition of sediments of a wide spread thermal subsidence phase.

Age of the Ventersdorp Structure

Isotopic age determinations on rocks of the Ventersdorp Supergroup have yielded an array of ages ranging from 1920 ± 100 Ma (Rb-Sr whole rock; Cornell, 1978) to greater than 2.6 Ga, with most ages clustering around 2.3 and 2.6 Ga. This large spread of dates is attributed to the homogenization of the strontium isotopic system at 1970 ± 100 Ma (Cornell, 1978), which occurred during intense metasomatism related to a thermal pulse associated with the Bushveld intrusion and/or burial metamorphism under the thick Transvaal sequence (see below). Strontium isotope homogenization appears to have occurred on a scale of tens of meters (Cornell, 1978), rendering the interpretation of Rb-Sr age determinations difficult at best. However, a reliable age for the Ventersdorp Group has been obtained by Van Niekerk and Burger (1978), using the U-Pb isotopic system, the analytical techniques of Krogh (1973), and decay constants of Steiger and Jager (1977). These authors made measurements on zircon phenocrysts extracted from rocks of the Makwassie Formation at widely separated (330 km) sample localities. An age of 2643 $\pm$ 80 Ma (quoted 95% confidence level) is indicated by data that do not show recent bulk lead loss, although Van Niekerk and Burger's (1978) quoted uncertainty is certainly smaller than their data justifies. This age is indistinguishable from an earlier Rb-Sr whole rock date (recalculated) of 2638 $\pm$ 125 Ma reported by Crockett (1971). A date of approximately 2.64 Ga is therefore accepted as the age of eruption and deposition of the Ventersdorp Supergroup.

Metamorphism

The Ventersdorp Supergroup has been metamorphosed to the lower greenschist facies although it lacks a penetrative deformation fabric (Tankard et al., 1982). The greenschist facies is characterized by the ubiquitous mineral assemblage albite-chlorite-actinolite-zoisite (Cornell, 1978; Winkler, 1979). In metabasites the mineral assemblage is commonly albite-epidote- quartz-chlorite-biotite-sphene-muscovite-actinolite, also characteristic of lower greenschist facies metamorphism. Based on oxygen isotope geothermometry Cornell (1978) estimated that the temperature of metamorphism was 200-400°C, and because of the lack of oriented fabrics and penetrative deformation he suggested the alteration was caused by burial metamorphism (c.f. Miyashiro, 1973). The pressure required by the mineral parageneses suggest burial to a depth of six km (Cornell, 1978). The burial metamorphism was associated with severe metasomatic alteration of the Ventersdorp rocks with a corresponding large scale (up to tens of meters) migration of the more mobile elements, and a complete resetting of the strontium isotopic system. A Rb-Sr whole rock isochron obtained from Ventersdorp rocks (Cornell, 1978) defines the time of this event as 1920 ± 100 Ma, roughly corresponding to the age of the Bushveld event, which occurred while the Ventersdorp rocks were buried under the thick Transvaal sequence. The metasomatic alteration is also suggested to have caused the anomalously low sodium content of some of the Ventersdorp rocks (Cornell, 1978).

KANYE VOLCANIC GROUP AND THE GABERONE GRANITE COMPLEX

The Gaberone Granite Complex and the Kanye Volcanic Group occur in the northern outcrop areas of the Ventersdorp Supergroup (Figure 7), although the poor exposure has generally obscured their contact relationships with the Ventersdorp rocks. Tyler (1979) has suggested that the Ventersdorp rocks rest non-conformably on both the Kanye volcanics and the Gaborone Granite, while Key and Wright (1982) consider the Gaborone complex to be intrusive into the Ventersdorp rocks and possibly even the much younger (circa 2.3 Ga) Transvaal Supergroup.

The Kanye volcanics are calc-alkaline and consist of silicic agglomerates, massive red and blue felsites which contain occasional varioles, and are dacitic to rhyolitic in composition (Tyler, 1979). The wide range of reported Rb-Sr ages for the Kanye volcanics (1.8-3.4 Ga, Key, 1976; Tyler, 1979) emphasizes the confusion regarding their "true age"; however, a sericitic paleosol horizon occurs between rocks of the Kanye volcanic Group and the Ventersdorp Supergroup, suggesting subaerial weathering between eruption of the Kanye and Ventersdorp volcanics (Tyler, 1979).

The Gaberone Granite contains a core of rapikivi granite that is surrounded by leucocratic granite and porphyritic microgranite, and intrudes the Kanye Volcanic Group (Boocock, 1959; Crockett, 1971; Poldervaart, 1952; Key and Wright, 1982). It has yielded a Rb-Sr whole rock isochron of 2394 $\pm$ 26 Ma with an R_i of 0.74991 $\pm$ 0.00643 (Key and Wright, 1982), but dating of the Gaberone Granite Complex has proved enigmatic in the past, with reported Rb-Sr isochrons ranging from 2328 Ma* (recalculated using decay constants of Steiger and Jager, 1977. Harding et al., 1974) to 2690 $\pm$ 75 Ma (Key, 1976).

Following DuToit (1946) we suggest a possible relationship between the Dominion Reef volcanics and Kanye volcanism. We interpret both as products of Andean-type arc magmatism. Key and Wright (1982) and Tyler (1979) have identified the Kanye magmatism with that of the Gaberone Complex, and this seems compatible with our Andean interpretation. Rb/Sr whole rock ages were considered by Key and Wright to yield a younger age of about 2.4 Ga. If this result is confirmed by, for example, modern zircon ages then our preferred hypothesis is invalid and the Gaberone Granite and Kanye volcanism are unrelated to the closure of the Limpopo Ocean (> 2.6 Ga and before the intrusion of the Great Dike).

THE LIMPOPO PROVINCE AS A CONTINENTAL COLLISION ZONE AND ITS RELATION TO THE VENTERSDORP RIFTING

The Limpopo Province contains some of the oldest rocks so far recognized on earth and has evidently experienced a complex and varied history prior to involvement in the suturing of the Kaapvaal and Zimbabwe Cratons. The collision appears to have been between an Atlantic-type margin developed on the southern (present coordinates) margin of the Zimbabwe Craton and an Andean-type margin developed on the Kaapvaal Craton side of the closing Limpopo Ocean. Following Dewey and Burke (1973) Light (1982) has identified the suture between the two cratons as a distinctive east-northeast trending line of sheared amphibolitic and peridotitic gneisses associated with schistose fuchsitic quartzites, aplitic riebeckite gneisses and granulites (Figures 7 and 8). North of the line of the proposed suture is a belt of thrusts and nappes which emplaced greenstones of oceanic affinity over carbonate shelf facies rocks of the Beitbridge Group resting on the Craton (Figure 8; Light, 1982; Kidd, 1984; Eriksson and Kidd, in prep.; Stowe, 1974). The Limpopo Province contains many major subvertical horizontally lineated shear zones, including the Tuli-Sabi, Mmadinare, Palala and Thabazimbi faults (Figure 7) (Kidd, 1984). Collision related sinistral motion accommodating hundreds of kilometers of displacement has been suggested for some of these faults (Tankard et al., 1982). However, some possible dextral motion (Kidd, 1984) may involve an episode of strike-slip faulting different from that concurrent with the formation of the Ventersdorp rifts. the orientation of Ventersdorp rifts indicates sinistral motion on associated northeast trending shear zones (Figure 7).

Metamorphic mineral assemblages from the central Limpopo Province indicated burial to greater than 30 km (Light, 1982), implying double crustal thickness at peak metamorphism (Kidd, 1984) consistent with the continental collision model set forth by Dewey and Burke (1973). The peak metamorphic/deformational event (2.7-2.6 Ga, Tankard et al., 1982) was associated with the intrusion of a large amount of syn-tectonic anatectic granites, including the Bulai Gneiss, Razi Granites, the Matok Pluton and numerous small bodies intruding the Baviaanskloof Gneiss (Light, 1982; Robertson, 1973a, 1973b). DuToit et al., (1983) have shown that the Matok pluton is cut by sinistral strike-slip faults.

Discussion

The Ventersdorp Supergroup was deposited in a complex rift system approximately 2.64 Ga ago. The development of this rift system closely resembled that of rifts which have formed more recently in the earth's history. Because rifts form in many tectonic environments in which the lithosphere is put into extension it is important to attempt to characterize the environment in which individual rifts form more fully. Collisional rifts are most numerous in the Phanerozoic and we here show that the Ventersdorp rifts were also associated with collision.

We have shown elsewhere (Burke et al., in prep) that the Witwatersrand Supergroup was deposited in a foreland basin (c.f. Jordan, 1981) developed on the continental side of a complex Andean arc. This Andean margin of the Kaapvaal Craton collided with the Atlantic-type margin of the Zimbabwe Craton approximately 2.64 Ga ago, forming the Tibetan style Limpopo Province (Burke and

Dewey, 1972; Burke, 1973; Dewey and Burke, 1973; Burke et al., 1976; Van Biljon, 1977; Light, 1982). The timing of the Kaapvaal/Zimbabwe continental collision is constrained between the age of the 2.7 Ga old "Bulawayan" greenstones and metasediments involved in the collision, and the 2.6 Ga old post-tectonic Chibi granite suite (Tankard et al., 1982). Barton et al., (1979) have reported ages of 2649 +82 -88 Ma (Pb-Pb whole rock) and 2693 ± 45 Ma (Rb-Sr whole rock) from the syn-tectonic (their D2-D3) Bulai gneiss from the central zone of the Limpopo Province. Furthermore, the suturing occurred before 2.5 Ga, the age of the Great Dike of Zimbabwe (Nisbet, 1982), which intrudes the Limpopo Province (Figure 7).

DuToit et al, (1983) and Tankard et al., (1982) have emphasized that the Limpopo zone is associated with a large amount of left lateral strike-slip faulting and anatectic granites (Figure 7). We link these to sideways motion on major strike-slip faults and anatexis related to crustal thickening following collision. The large amount of strike-slip faulting associated with this collision (Figure 7) leads us to suggest that the contemporary Ventersdorp rifting occurred in a zone of "imperfect strike-slip faulting" associated with the collision, similar in a way to formation of the Basin and Range Province of the western United States (Sengor and Burke, 1978).

<u>In Summary</u>:

We suggest that the Ventersdorp Structure is impactogenal (cf. Sengor et al., 1978) because the Ventersdorp Rifts strike at a high angle into the Limpopo continental collision zone and because the age of the Ventersdorp rifting is indistinguishable from that of the suturing of the Kaapvaal and Zimbabwe Cratons. Analagous structures are forming at the present time in China and elsewhere in Asia (for example; the Shansi Grabens and the Baikal Rift System) in response to the contemporary India/Asia continental collision (Molnar and Tapponnier, 1975). The deep erosional level (> 30 km) offered by the Limpopo collision zone could provide some useful insights about the processes currently occurring at inaccessible depths below the surface of Tibet and through much of Asia.

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Figure Captions

Figure 1. Location, outcrop distribution and inferred sub-surface distribution of the Ventersdorp Supergroup. Much of the area to the northeast of the Ventersdorp structure has been intruded by the Bushveld complex (not shown), or is obscured by younger rocks. The crescent shape of Ventersdorp outcrops southwest of Pretoria is related to overturning of strata associated with the Vredefort impact. The high grade Limpopo Province is inferred to extend under younger cover along the 'Limpopo Gravity High' (map based on S. A. C. S., 1980; Tankard et al., 1982 and Pretorius, 1976).

Figure 2. Stratigraphic column of the Ventersdorp Supergroup from its type area near Bothaville showing many features which are typical of rift stratigraphy, including: immature basement derived sediments, rapid lateral facies variations and volcanic rocks that are bimodal in silica content (based on Winter, 1976 and S. A. C. S., 1980).

Figure 3. Schematic cross section of the Ventersdorp Supergroup across the De Bron Horst. Note the rapid lateral facies variations in the Platberg Group, the rotated attitude of the Klipriviersberg lavas with respect to overlying formations and the blanketing nature of the Pniel Sequence (Bothaville and Allanridge formations). After Buck (1980) and Tankard et al., (1982).

Figure 4. Facies distribution for the lower Platberg Group in the vicinity of the De Bron Horst (from Buck, 1980). Note the fanglomerate deposits bounding major faults and the rapid transition into lacustrine deposits in basin centers. (a) is for the Ararat Basin, and (b) is for the Virginia Basin. Inset in Figure 3 for location.

Figure 5. Schematic paleoenvironmental reconstructions of the lower Platberg Group (a), upper Platberg Group (b), and Bothaville Formation (c). Note the gradual erosion of horst blocks and filling of grabens suggesting decreasing activity on faults during deposition of the upper Platberg Group. (From Buck, 1980.)

Figure 6. A) Schematic cross section of the Ventersdorp structure between Sodium and Taung (from Visser et al., 1976). Note the draping of the Bothaville Formation, forming the "steer's head" geometry which is typical of the thermal subsidence phase of rift development (Burke, 1979). B) Shows the location of the cross section, as well as basement horsts in the Ventersdorp Rift System.

Figure 7. The Ventersdorp Rift System of the Kaapvaal Craton showing basement horsts, which along with the rift margins, strike at a high angle into the Limpopo continental collisional zone. Because the collision between the Kaapvaal and Zimbabwe Cratons and the formation of the Ventersdorp Rifts occurred synchronously (at approximately 2.64 Ga), we suggest that the Ventersdorp structures are impactogens. Numerous anatectic plutons of a similar age occur throughout the continental collision zone and are perhaps related to crustal thickening following collision. Limpopo style rocks do not outcrop along the northwestern part of the Kaapvaal Craton but are obscured by younger sequences. We interpret the Limpopo Gravity High (Pretorius, 1976), as indicating that they extend into this area. This map is based on data compiled from Tankard et al., (1982); Visser et al., (1976); DuToit et al., (1983); Pretorius, (1976); Button, (1981).

Figure 8. Cross section through the Zimbabwe and Kaapvaal Cratons across the Limpopo Province, which is interpreted as the product of continental collision.

The circa 2.64 Ga collision was between an Atlantic-type margin on the Rhodesian Craton (the Beitbridge Group represents the highly deformed remnants of the carbonate shelf, Kidd, 1984), and an Andean-type margin formed on the Kaapvaal Craton. The Witwatersrand Supergroup was deposited in a foreland trough behind this Andean arc, which is represented by the Kanye volcanics and the Gaborone Granite, as well as by the Dominion Reef lavas, which we correlate with the Kanye volcanics. The Ventersdorp Rifts strike at high angles into the Limpopo collisional zone (see Figure 7), and are here interpreted as impactogens accommodating extension in a zone of imperfect left-lateral strike-slip faulting induced by the collision.

Table 1. Chemical analyses of the volcanic rocks from the Ventersdorp Supergroup. Note that these rocks are bimodal in silica content, which is typical of volcanic rocks deposited in rifts. Analyses 1-7 from McIver et al., (1982), #4 = mean of eight analyses, #5 = mean of 3 analyses, #6 = mean of 12 analyses, #7 = mean of 14 analyses. Analyses 8, 9 and 10 from Tyler (1979) (his #92A, #93A and #94A). Analysis #12 from Tankard et al., (1982).

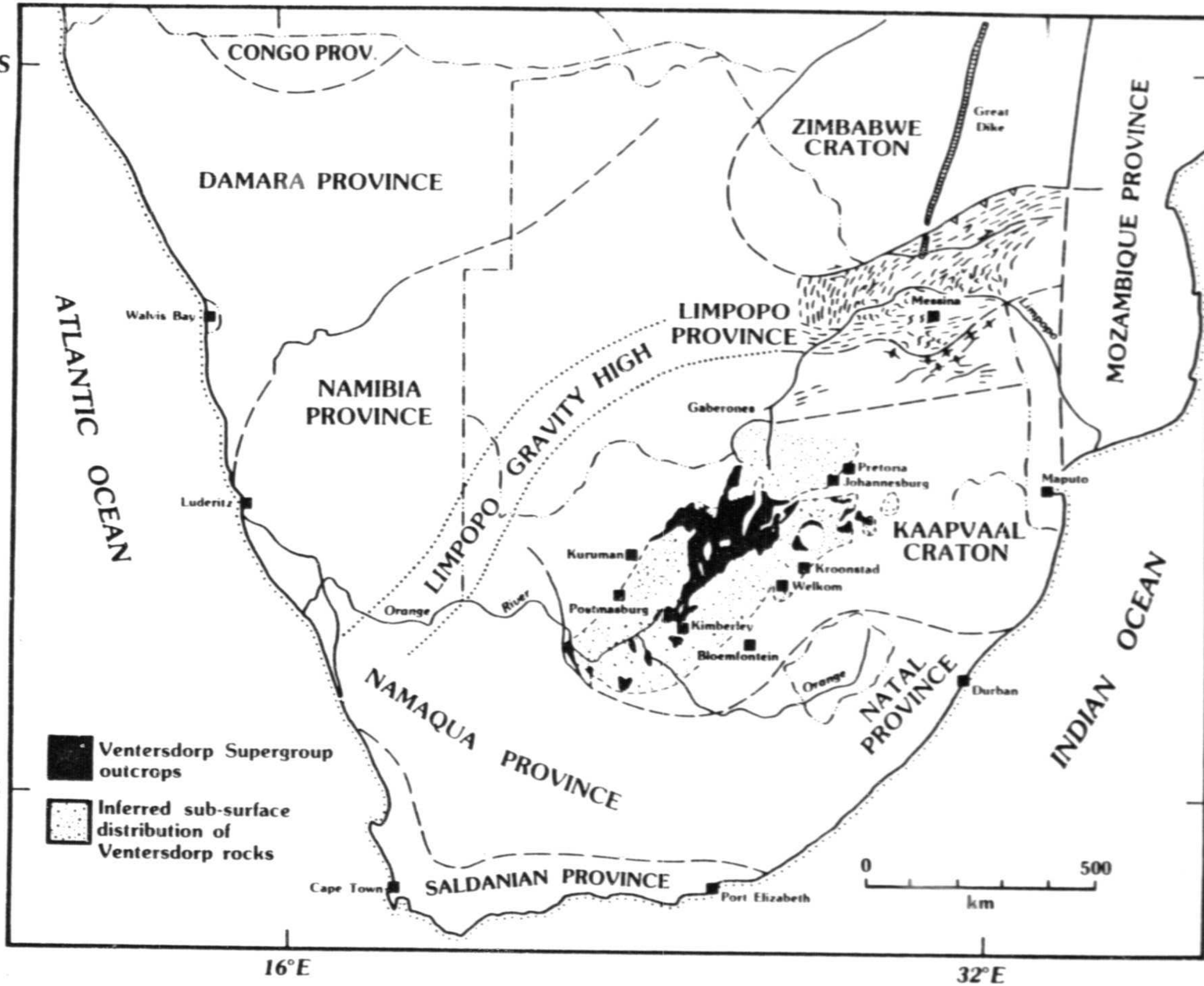

18°S
32°S
16°E
32°E
CONGO PROV.
DAMARA PROVINCE
ZIMBABWE CRATON
Great Dike
MOZAMBIQUE PROVINCE
ATLANTIC OCEAN
Walvis Bay
Luderitz
NAMIBIA PROVINCE
LIMPOPO PROVINCE
LIMPOPO GRAVITY HIGH
Messina
Limpopo
Gaberones
Pretoria
Johannesburg
Maputo
KAAPVAAL CRATON
Kuruman
Kroonstad
Welkom
Postmasburg
Kimberley
Bloemfontein
Orange
River
Orange
NATAL PROVINCE
Durban
INDIAN OCEAN
NAMAQUA PROVINCE
Ventersdorp Supergroup outcrops
Inferred sub-surface distribution of Ventersdorp rocks
SALDANIAN PROVINCE
Cape Town
Port Elizabeth
0
500
km

Fig 1

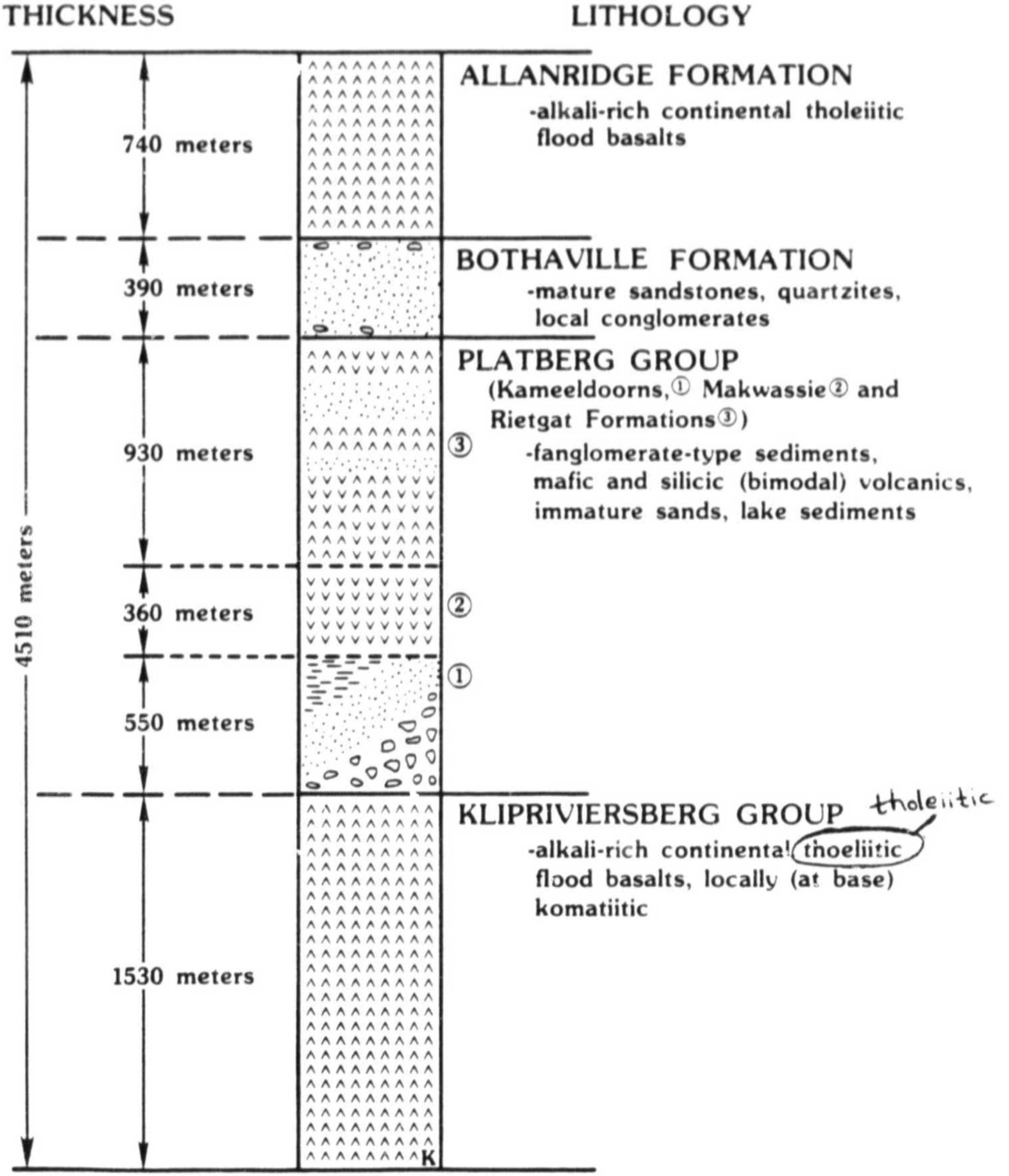
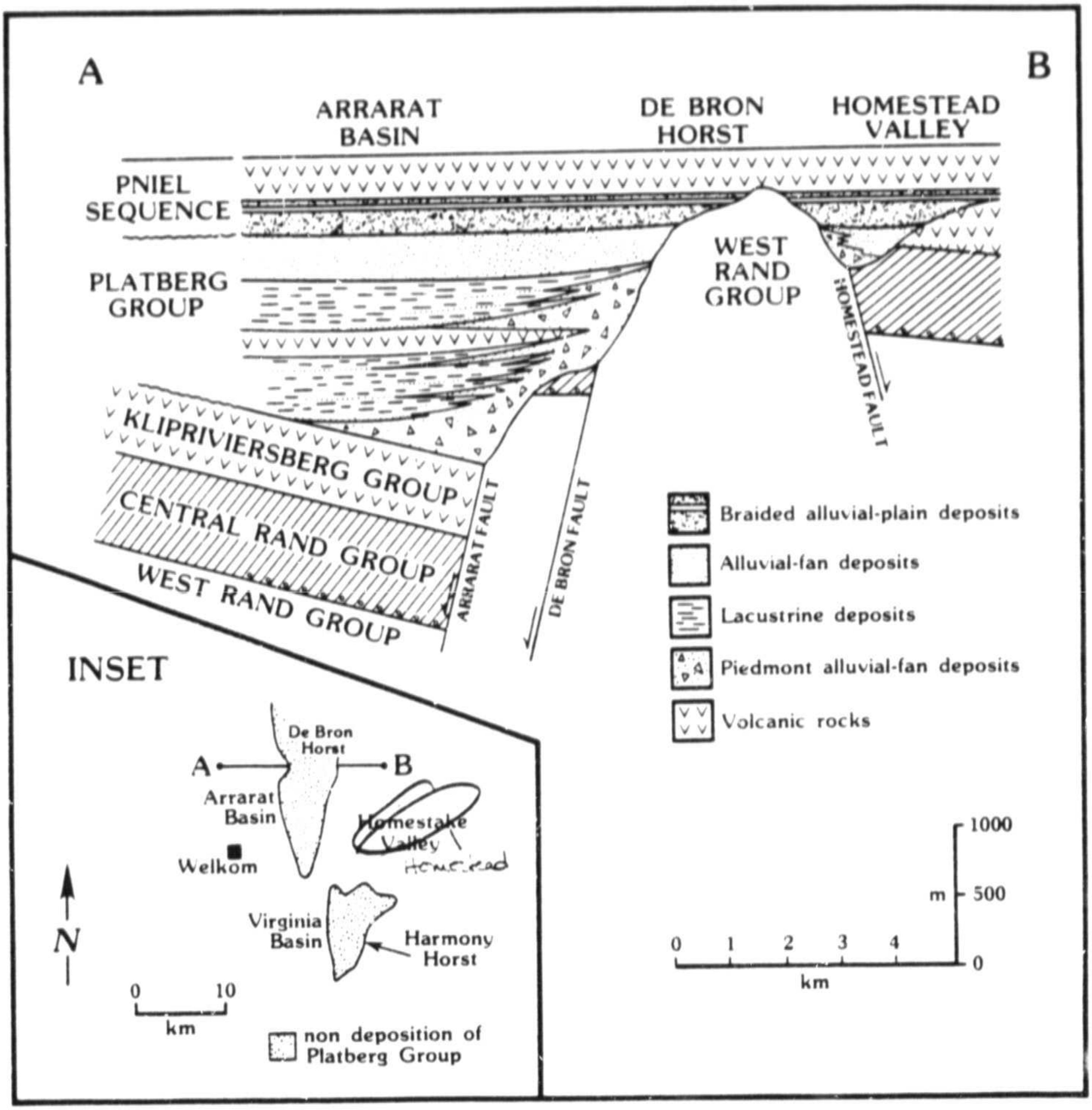

THICKNESS
LITHOLOGY
740 meters
ALLANRIDGE FORMATION
-alkali-rich continental tholeiitic flood basalts
390 meters
BOTHAVILLE FORMATION
-mature sandstones, quartzites, local conglomerates
930 meters
PLATBERG GROUP
(Kameeldoorns,① Makwassie② and Rietgat Formations③)
-fanglomerate-type sediments, mafic and silicic (bimodal) volcanics, immature sands, lake sediments
③
360 meters
②
550 meters
①
4510 meters
1530 meters
KLIPRIVIERSBERG GROUP
tholeiitic
-alkali-rich continental thoeliitic flood basalts, locally (at base) komatiitic
K

Fig 2



A
B
ARRARAT BASIN
DE BRON HORST
HOMESTEAD VALLEY
PNIEL SEQUENCE
PLATBERG GROUP
WEST RAND GROUP
HOMESTEAD FAULT
KLIPRIVIERSBERG GROUP
CENTRAL RAND GROUP
WEST RAND GROUP
ARRARAT FAULT
DE BRON FAULT
Braided alluvial-plain deposits
Alluvial-fan deposits
Lacustrine deposits
Piedmont alluvial-fan deposits
Volcanic rocks
INSET
De Bron Horst
A
B
Arrarat Basin
Homestake Valley
Homestead
Welkom
N
Virginia Basin
Harmony Horst
0
10
km
non deposition of Platberg Group
1000
m
500
0
0
1
2
3
4
km

Fig 3

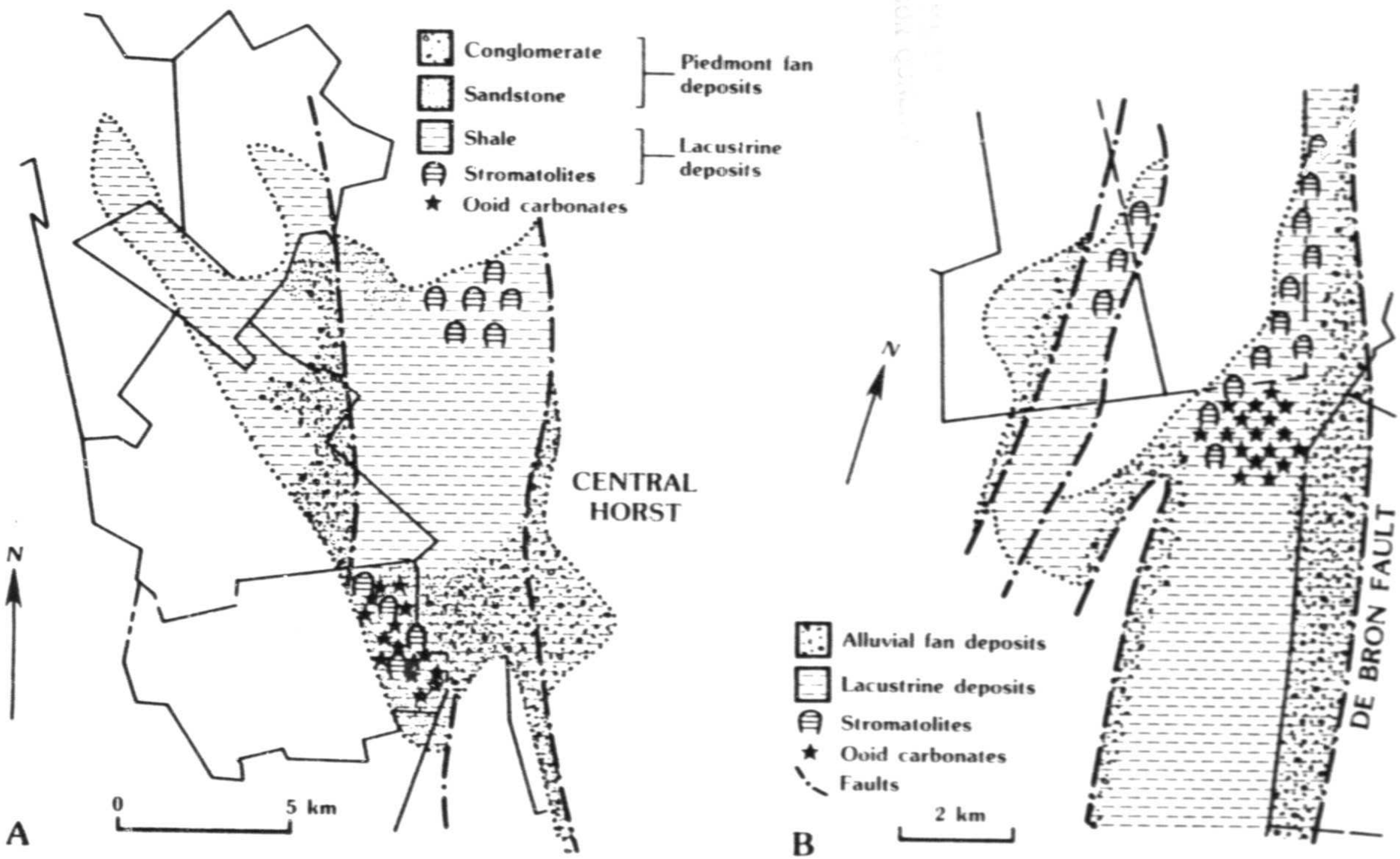

Conglomerate
Sandstone
Piedmont fan deposits
Shale
Stromatolites
Lacustrine deposits
Ooid carbonates
CENTRAL HORST
N
0
5 km
A
N
Alluvial fan deposits
Lacustrine deposits
Stromatolites
Ooid carbonates
Faults
2 km
DE BRON FAULT
B

Fig 4

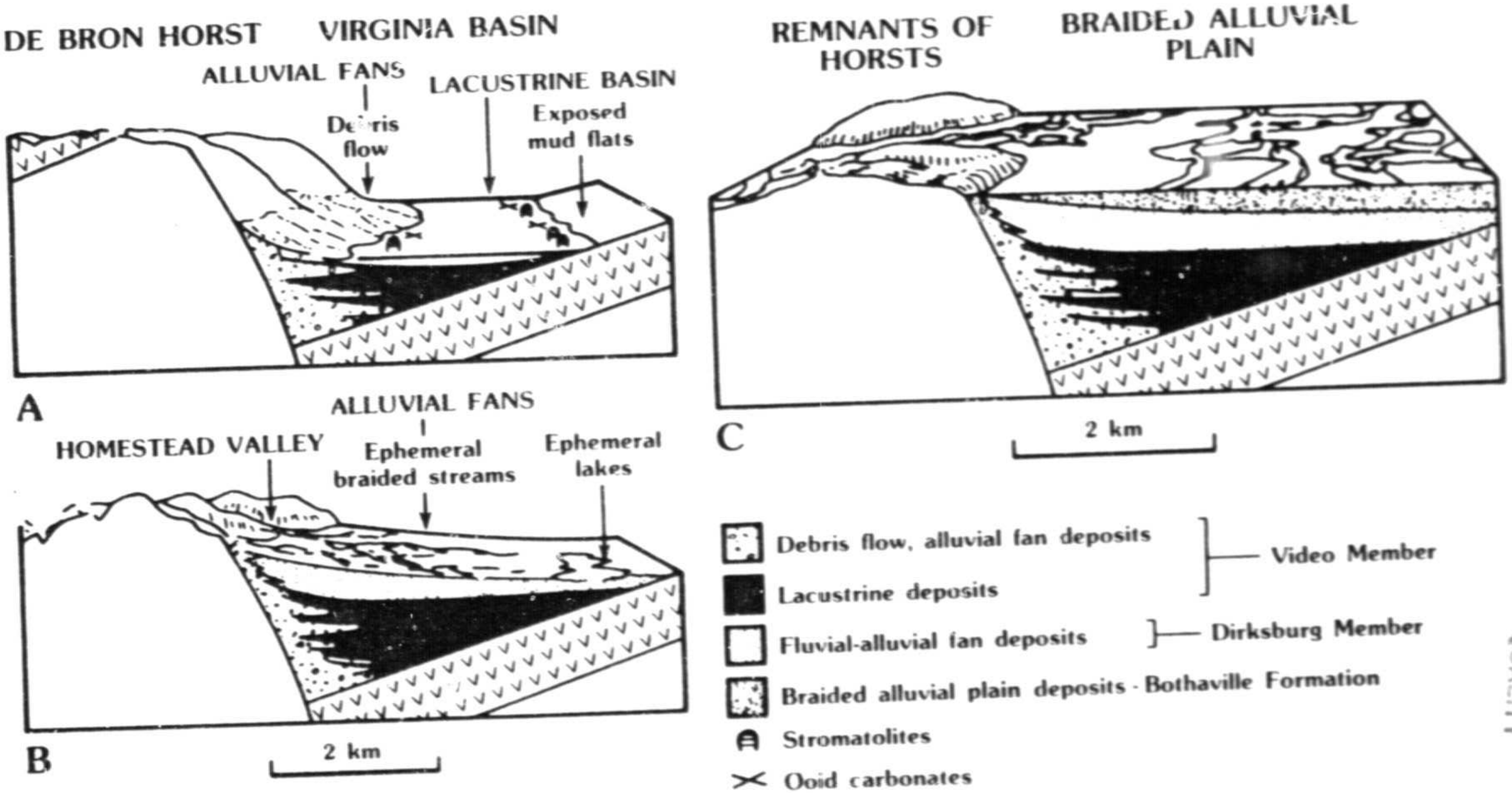

DE BRON HORST
VIRGINIA BASIN
ALLUVIAL FANS
LACUSTRINE BASIN
Debris flow
Exposed mud flats
A
ALLUVIAL FANS
HOMESTEAD VALLEY
Ephemeral braided streams
Ephemeral lakes
B
2 km
REMNANTS OF HORSTS
BRAIDED ALLUVIAL PLAIN
C
2 km
Debris flow, alluvial fan deposits
Lacustrine deposits
Video Member
Fluvial-alluvial fan deposits
Dirksburg Member
Braided alluvial plain deposits - Bothaville Formation
Stromatolites
Ooid carbonates

Fig 5

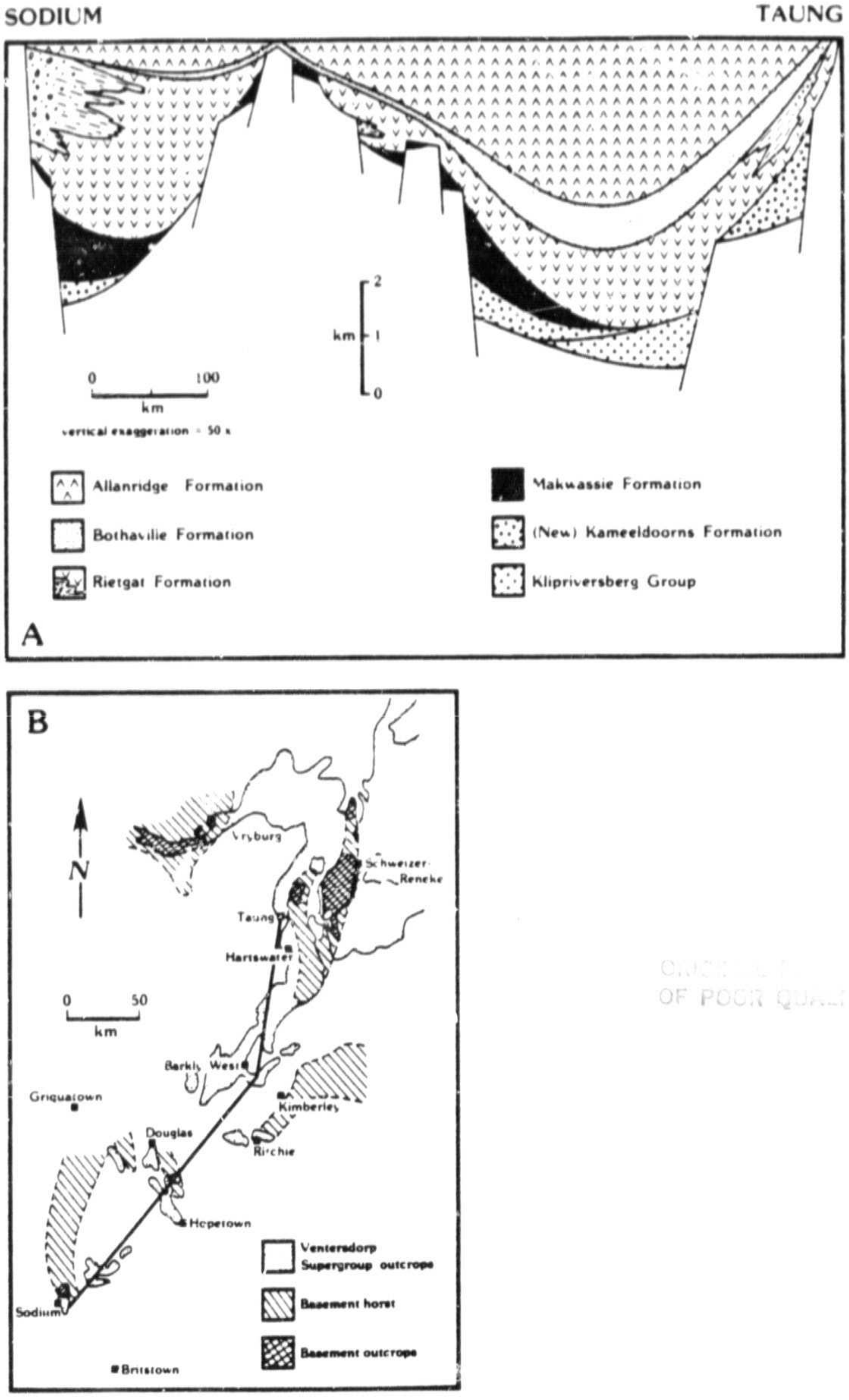

SODIUM
TAUNG
km
2
1
0
0
100
km
vertical exaggeration = 50 x
Allanridge Formation
Bothaville Formation
Rietgat Formation
Makwassie Formation
(New) Kameeldoorns Formation
Klipriversberg Group
A
B
N
Vryburg
Schweizer-Reneke
Taung
Hartswater
0
50
km
Barkly West
Griquatown
Kimberley
Douglas
Ritchie
Hopetown
Sodium
Britstown
Ventersdorp Supergroup outcrops
Basement horst
Basement outcrops

Fig 6

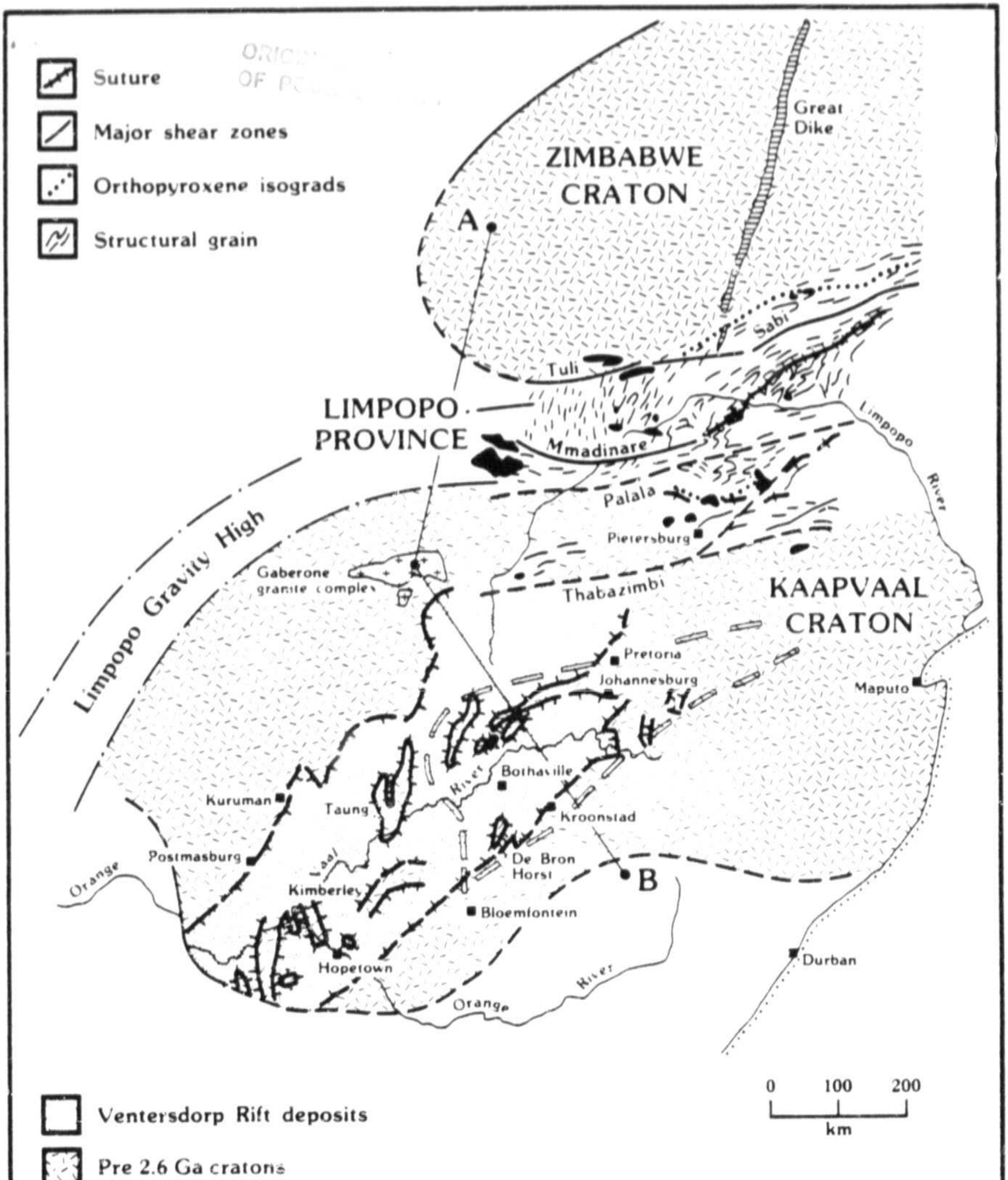
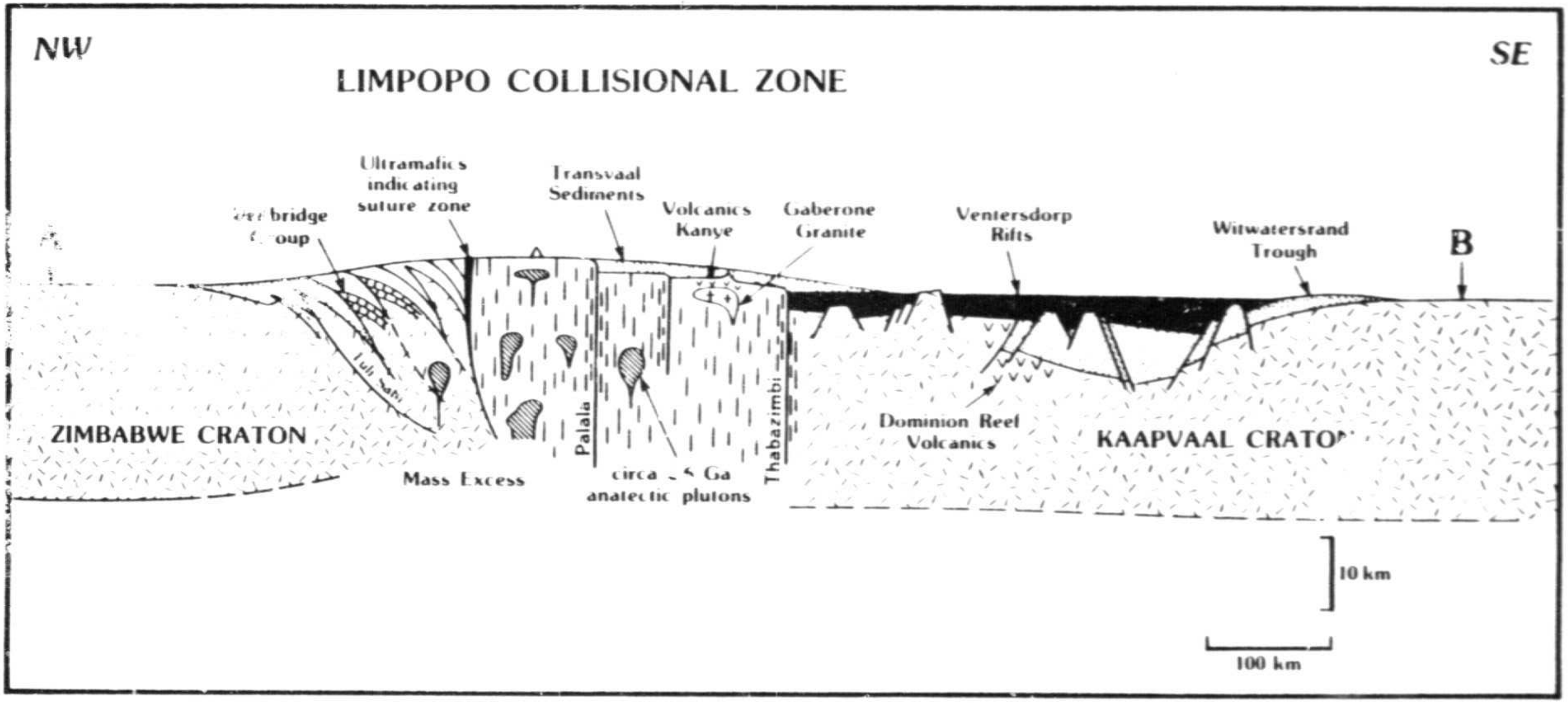

Suture
Major shear zones
Orthopyroxene isograds
Structural grain
Great Dike
ZIMBABWE CRATON
A
Sabi
Tuli
LIMPOPO PROVINCE
Mmadinare
Limpopo River
Palala
Pietersburg
Limpopo Gravity High
Gaberone granite complex
Thabazimbi
KAAPVAAL CRATON
Pretoria
Johannesburg
Maputo
Bothaville
River
Kuruman
Taung
Kroonstad
Postmasburg
Vaal
De Bron Horst
B
Orange
Kimberley
Bloemfontein
Hopetown
Durban
Orange River
0 100 200
km
Ventersdorp Rift deposits
Pre 2.6 Ga cratons
Circa 2.6 Ga anatectic plutons
Horsts, normal faults
Andean arc related Gaborone Granite and Kanye volcanics
Postulated extent of Witwatersrand foreland trough (based on Pretorius 1976)

Fig 7



NW
SE
LIMPOPO COLLISIONAL ZONE
Ultramafics indicating suture zone
Transvaal Sediments
Volcanics Kanye
Gaberone Granite
Ventersdorp Rifts
Witwatersrand Trough
B
Tuli Sashi
Palala
Thabazimbi
ZIMBABWE CRATON
Mass Excess
anatectic plutons
Dominion Reef Volcanics
KAAPVAAL CRATON
10 km
100 km

Fig 8