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ARCHEAN FORELAND BASIN TECTONICS IN THE WITWATERSRAND, SOUTH AFRICA

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ABSTRACT

The Witwatersrand Basin of ~~of~~ South Africa is the best-known of Archean sedimentary basins and contains some of the largest gold reserves in the world. Sediments in the basin include a lower flysch-type sequence and an upper molassic facies, both of which contain abundant silicic volcanic detritus. The strata are thicker and more proximal on the northwestern side of the basin which is, at least locally, bound by thrust faults. These features indicate that the Witwatersrand strata may have been deposited in a foreland basin and a regional geologic synthesis suggests that this basin developed initially on the cratonward side of an Andean-type arc. Remarkably similar Phanerozoic basins may be found in the southern Andes above zones of shallow subduction. We suggest that the continental collision between the Kaapvaal and Zimbabwe Cratons at about 2.7 Ga caused further subsidence and deposition in the Witwatersrand Basin. Regional uplift during this later phase of development placed the basin on the cratonward edge of a collision-related plateau, now represented by the Limpopo Province. Similarities are seen between this phase of Witwatersrand Basin evolution and that of active basins north of the Tibetan Plateau (eg. the Tarim and Tsaidam Basins). The geologic evidence does not appear so compatible with earlier suggestions that the Witwatersrand strata were deposited in a rift or half-graben.

INTRODUCTION

There are two main varieties of intracontinental sedimentary basins: rifts and foreland basins. Although these two basin types show contrasting tectonic, sedimentary, and thermal histories, they may show confusingly similar subsidence patterns (see for example: Bally, 1982). Both can show rapid initial subsidence for fundamentally different reasons. In rifts (McKenzie, 1978) the first phase of rapid subsidence is produced by isostatic compensation for thinning of the lithosphere, while in foreland basins it is caused by flexure of the lithosphere in response to peripheral loading and compression. In rifts, the second, slower phase of subsidence is produced by thermal relaxation of the lithosphere, while in foreland basins the slowing of subsidence can be attributed to decreasing load as erosion in the peripheral mountains increases and compression declines. It is important to attempt to discriminate between these two types of subsidence in Archean sedimentary basins if the thickness of deposits is to be used to estimate the thickness of the early lithosphere. There is a need for synthesizing all depositional and related geologic data before tectonic, subsidence, and thermal modelling of a basin is attempted.

Stretching or rift models have been applied to many Archean sedimentary basins including the Witwatersrand basin (e.g. McKenzie et al., 1980; Bickle and Eriksson, 1982; Nisbet, 1984). We review the geologic history, including stratigraphic data for the Witwatersrand Basin and suggest that it is not a rift but a foreland trough.

The Witwatersrand basin of South Africa is an elongate structure of Archean age (2.8 - 2.6 Ga) filled predominantly with clastic sediments of the West Rand and Central Rand Groups, together constituting the Witwatersrand Supergroup. These are locally, in the northwestern part of the basin, underlain by the

volcano-sedimentary Dominion Group rocks. The structure lies on the Kaapvaal Craton and trends in a northeasterly direction parallel with, but some distance south of, the high-grade gneissic terrane of the Limpopo Province (Figure 1). The high grade metamorphism, calc-alkaline plutonism, uplift and cooling in the Limpopo are of the same age as Witwatersrand sedimentation and we suggest that they are closely related.

FORELAND BASINS

A foreland basin is an elongate trough adjacent to a mountain belt that should display some, but not necessarily all, of the following characteristics:

(1) Foreland basins have an asymmetric cross-section, with thicker strata and steeper dips on the side next to the mountains

(2) they are likely to be bounded on the side nearest the mountain belt by thrust faults, while the opposite margin may show uplift with or without normal faults throwing down towards the basin

(3) sedimentary facies, as in the Alpine example, may include a lower "flysch" sequence and an upper "molasse" facies

(4) there is unlikely to be much magmatism in the basin

(5) the margin closest to the mountain front may show much erosion or "cannibalization" of sediments, with numerous unconformities in the section

(6) deformation may be syn-sedimentary, younging and becoming weaker away from the mountain front

(7) there may be a progressive migration of (young) normal faults, the depositional axis and thrust faults away from the mountains, especially in collisional-type foreland basins

(8) the basin may shrink in area with time in a way attributable to viscoelastic relaxation of the lithosphere.

With these criteria in mind we here test the hypothesis that the Witwatersrand structure represents a foreland basin.

STRATIGRAPHY OF THE WITWATERSRAND BASIN

The predominantly clastic fill of the Witwatersrand Basin has been divided into the West Rand and the overlying Central Rand Groups which rest conformably on the largely volcanic Dominion Group (Figure 2). The Dominion Group was deposited over approximately 15,000 square kilometers, although its correlation with the Kanye and other similar volcanic suites (DuToit, 1946; Burke et al., 1985b) suggests a larger spatial distribution and temporal variation in Dominion-style volcanism. The overlying West Rand and Central Rand Groups were deposited in a larger basin approximately 80,000 square kilometers in area (Pretorius, 1981). Stratigraphic thicknesses of the West Rand Group generally increase towards the fault-bounded northwestern margin, while those of the Central Rand Group increase towards the center of the preserved portion of the basin. Strata of both groups thin considerably, both in individual and cumulative unit thickness, towards the southeastern basin margin (Figure 3; Jansen, 1969). The northeastern and southwestern margins of the Witwatersrand structure remain ill-defined because Witwatersrand strata become buried under thick sequences of younger cover in these areas (Pretorius, 1981). However, Button's (1977) correlation of the Godwan Formation with the upper Witwatersrand strata (Figure 3) suggests that the depositional basin may be larger and more elongate than previously recognized. Paleocurrent analyses also suggest that the original depositional basin extended considerably past the present southwestern margin (Tankard et al., 1982). Witwatersrand sediments that were originally deposited north of Johannesburg are either deeply buried under younger cover sequences or have been severely modified (and in most places destroyed) by subsequent uplift, tectonism, igneous intrusion and erosion.

Pretorius (1976) has estimated that erosion has removed sediments of the Witwatersrand Supergroup for 50 km northwest of the present basin margin, and for 20 km southeast of the southeastern margin.

DOMINION GROUP

Dominion Group rocks rest non-conformably on 2.9 - 3.0 Ga "granitic" basement and underlie the Witwatersrand Supergroup in many areas. Dominion rocks, which have been divided into the Renosterspruit, Renosterhoek and Syferfontein Formations (Figure 4) attain a maximum thickness of 2250 meters in the Klerksdorp area (S.A.C.S., 1980 and Figure 3).

The Renosterspruit Formation lying at the base of the Dominion Group contains up to 60 meters of sandstone along with minor conglomerate and argillaceous horizons. Its base is marked by a medium to coarse-grained vein quartz pebble conglomerate disposed in sheetlike gravel bars, channels and single pebble layers with local placer mineral concentrations. Paleocurrent analyses and the presence of trough cross beds prompted Tankard et al., (1982) to suggest that the Renosterspruit sediments were deposited by braided streams flowing over subdued topography developed on the granitic basement. Successively higher beds of quartz arenite and pebbly sandstone in the Renosterspruit Formation have been interpreted as a valley fill sequence, deposited under the influence of a gentle southwestward paleoslope (Haughton, 1969). As the upper part of the Renosterspruit Formation grades into the overlying Renosterhoek Formation the sediments become interbedded with silicic tuffs and mafic lava flows and they are extensively intruded by diabase and silicic porphyry sills.

The Renosterhoek Formation contains up to 1100 meters of "basaltic andesites and tuffs" with abundant interflow paleosols and sandstone horizons (Tankard et al., 1982). Tuffaceous breccias commonly mark the tops of major flows.

The Syferfontein Formation is approximately 1550 meters thick and rests conformably on top of the Renosterhoek Formation (S.A.C.S., 1980). It consists of rhyolites, silicic tuffs, volcanic breccias and subordinate andesitic lavas. Whiteside (1970) distinguished three members of the Syferfontein Formation on a textural basis. His lower member contains porphyritic rhyolites, the middle member is composed of alternating rhyolite and andesite flows (up to 810 meters thick) and the upper member contains "cherty" lava with quartz and feldspar phenocrysts, and pink, red, green, dark gray and black tuffs, banded lavas and meta-bentonite beds. Locally the lavas are silicified (Haughton, 1969).

WITWATERSRAND SUPERGROUP

West Rand Group

The West Rand Group consists of southeastward tapering sedimentary wedges of the Hospital Hill, Government, and Jeppestown Subgroups (Figures 5-10). It was deposited conformably on top of and over a larger area than the Dominion Group, coming to overlap granitic basement in many areas. Although rocks of the West Rand Group are observed in surface outcrops over a 42,000 square kilometer oval area, Pretorius (1981) noted that the Bouguer gravity anomalies typically associated with these rocks extend considerably past the northeast and southwest margins of the Witwatersrand structure, supporting Button's (1977) northeastward extension of the basin on stratigraphic evidence (Figure 3). Pretorius (1981) estimated that the West Rand Group was originally deposited over an area of approximately 100,000 square kilometers.

Shale and sandstone in approximately equal proportions characterize the West Rand Group, although a thin horizon of mafic volcanic rocks (the Crown lava) is locally present (Table 1). This volcanic unit thickens to nearly 250 meters near the northern margin of the basin, while being totally absent in the southern part of the structure. The maximum thickness of the West Rand Group (7500 meters) occurs on the northern margin of the basin, northwest of Krugersdorp (Tankard et al., 1982), and the group thins southward and eastward to a minimum preserved thickness of 830 meters at Evander (Figures 3 and 5-10).

The Orange Grove Formation forms the base of the West Rand Group (Figure 2) and is particularly well developed in the Klerksdorp and Heidelberg areas. It contains mature quartzites composed of detrital quartz, minor chert, and less than 1% sericite derived from altered feldspars (Pretorius, 1964). Pale green fuchsite associated with detrital chromite in the Orange Grove Formation has been described from several locations (Pretorius, 1964). Camden-Smith (1980) attributed the maturity of the Orange Grove Formation to tidal and aeolian reworking and, noting the bimodal and bipolar current directions, interpreted its paleoenvironment as an ebb-dominated tidal deposit that was later influenced by beach-swash deposition. Near the top of the Orange Grove Formation is a five meter shale interval interbedded with a few thin sandstone layers that display a hummocky cross-stratification. The sequence from tidal flat to beach and then to (deeper water) shale facies indicates a deepening of the Witwatersrand Basin during deposition of the Orange Grove Formation. Hummocky cross-stratification is usually taken to indicate that sediments were not deposited deeper than the base of storm generated waves.

Sediments composing the numerous formations of the rest of the Hospital Hill, Government, and Jeppestown Subgroup (Figure 2 and Figures 5-10) are all broadly similar, consisting mostly of argillaceous sediments, and are thus treated together here (for detailed descriptions of these rocks see: Pretorius, 1981; Tankard et al., 1982, and references therein). Magnetic shales of this facies have been used to map the subsurface distribution of the Witwatersrand strata (eg. Krahmann, 1936). The shales are laterally persistent, preserve a horizontal sedimentary lamination and are partly chemical in origin, suggesting that they represent a distal shelf facies (Tankard et al., 1982) or an epicontinental sea deposit. Interbedded with these shales of the Government and Jeppestown Groups are graded sand beds (turbidites?) and lensoid lithic arenites which display bipolar paleocurrent directions suggesting some re-working of sediments. These arenaceous deposits (not quartzites, as they have sometimes been called) typically contain 50-60% quartz, 10% feldspar, 5% biotite, 5% rock fragments (including silicic volcanics), 5% calcite and minor sericite, chlorite, pyrite, leucoxene, epidote, apatite, zircon and tourmaline (Pretorius, 1964).

Central Rand Group

The Central Rand Group was deposited conformably on top of the West Rand Group and attains a maximum (preserved) thickness of 2880 meters northwest of both the center of the basin and the Vredefort Structure (Figures 3 and 11-14). The Central Rand Group which is distributed over an area of approximately 9750 square kilometers has been divided into the lower Johannesburg and the upper Turffontein Subgroups (Figure 2).

Sediments of the Central Rand Group consist predominantly of coarse grained sub-graywackes and conglomerates along with subordinate quartz arenite interbedded with local lacustrine or shallow marine shales and siltstones. The

conglomerates are usually poorly sorted with the larger clasts being well rounded while the smaller pebbles are subangular to angular. Paleocurrent indicators show that sediments of the Johannesburg Subgroup prograded into the basin from the northwestern margin (except at Evander; Figure 3) in the form of fan-delta complexes (Figure 15). This is economically important because numerous goldfields in the Central Rand are closely associated with the major entry points into the basin (Borchers, 1961; Brock and Pretorius, 1964). Some transport of sediments along the depositional axis of the basin is indicated by paleocurrent directions from Welkom (Brock and Pretorius, 1964, Tweedie, 1978), Edenville (Hutchison, 1975), and northwest of Vredefort (Brock and Pretorius, 1964).

A few tuffaceous horizons have been described from the Bird Reef of the Johannesburg Subgroup (Pretorius, 1964) and a thin mafic lava flow (Bird amygdaloid, Figure 2) occurs in the northeast of the basin. Like the Crown Lava of the West Rand Group, it is absent in the southeastern part of the basin.

The great dispersion of unimodal paleocurrent directions derived from most of the Central Rand Group indicates that these sediments were deposited in shallow braided streams on coalescing alluvial fans. The paleorelief is estimated at about 6 meters in areas proximal to the source, and less than 1/2 meter in more distal areas (Tankard et al., 1982). Some of the placers of the Central Rand Group have planar upper surfaces, commonly associated with pebbles and heavy placer mineral concentrations, which may be attributed to re-working by tidal currents. Central Rand Group sediments have a complex mineralogy, but in general order of abundance clast lithologies in the conglomerates include vein quartz, quartz-arenite, (volcanic?) chert, banded chert, silicic volcanics, shales and schists, red jasper and a rare tourmalinized rock.

TECTONIC SIGNIFICANCE OF SEDIMENTS AND LAVAS

NATURE OF THE DOMINION GROUP

Several factors suggest to us that the Dominion Group may represent part of an Andean volcanic arc. Andean arcs are built upon continental crust and they typically contain a large proportion of pyroclastic volcanic rocks. They are characteristically deficient in the most mafic members of the calc-alkaline basalt-andesite-rhyolite volcanic arc suite (Miyashiro, personal communication) and this is consistent with the dominantly andesitic-rhyolitic nature of the Dominion lavas. The Dominion lavas were clearly deposited on continental-type crust, and the explosive nature of Dominion volcanism is manifest in the numerous tuffaceous and breccia horizons in the group, although abundant paleosol layers (Button and Tyler, 1981) suggest that significant intervals characterized by subaerial weathering elapsed between individual eruptions. The rocks of the Dominion Group possess strong affinities to Andean arc-type rocks, and they are unlike the volcanic rocks deposited in either island-arc or continental-rift settings (cf. Burke et al., 1985a) which are likely to have a larger to much larger mafic component. Stratigraphic thicknesses in the Dominion Group generally increase towards the north-central portion of its outcrop area (near Klerksdorp, see Figure 4), suggesting that the main eruptive zone was in that direction. Paleocurrent directions derived from sedimentary horizons within the Dominion Group indicate a south to southwestward regional paleoslope (Haughton, 1969), which is consistent with our suggestion that the Dominion lavas were erupted on the landward flank of a continental margin volcanic arc.

Our identification of the Dominion lavas as a product of Andean-type magmatism suggests that there may be contemporary similar volcanic piles along the northern edge of the Kaapvaal Craton. One candidate is the Kanye Volcanic Group, a calc-alkaline series of silicic agglomerates and dacitic to rhyolitic felsites (Tyler, 1979; Key, 1976; Key and Wright, 1982), which DuToit (1954) and Burke et al., (1985b) have correlated with the Dominion lavas because of their similar stratigraphic positions. The Kanye Volcanic Group is intruded by the shallow-level Gaberone Granite Complex (Tyler, 1979; Key, 1976), a relationship which is common in Andean arc tectonic settings.

Identification of other possible volcanic centers of this very old Andean arc is hampered by; (1) very sparse outcrop, (2) by the generally deep erosional level of the area to the north of the Witwatersrand Basin, (3) by extensive covering of older rocks by the circa 2.0 Ga Bushveld intrusion (see Figure 1) and, (4) by the severe deformation and metamorphism suffered by rocks of the northern Kaapvaal Craton. It is perhaps significant that Light (1982) and Fripp (1983) have interpreted some of the gneisses and amphibolites in the Limpopo Province as products of Andean arc magmatism, an hypothesis which is not inconsistent with our interpretation of the Dominion and Kanye Groups.

NATURE OF THE WEST RAND GROUP

The lower West Rand Group records subsidence of the Witwatersrand basin, because the sediments grade vertically from beach deposits (Camden-Smith, 1980) to a distal shallow-water marine facies. This inundation of shallow water sedimentary environments suggests that the subsidence was rapid, but the absence of coarse, immature, fanglomerate-type sediments suggest to us that the subsidence was not accomodated by large displacements on normal faults as would be expected if the sediments were deposited in a rift. Furthermore, the lateral

persistence of most sedimentary units within the West Rand Group is unlike the lower parts of rift-stratigraphy, which typically exhibit rapid lateral variations in thickness and type of sedimentary units (Burke et al., 1985a). We therefore prefer to interpret subsidence in the Witwatersrand Basin as having been flexurally accomodated.

A decreasing rate of subsidence and/or a higher rate of clastic sediment supply is indicated by the progressively shallower-water facies deposits in the upper West Rand Group. Proximal "shelf", beach and fan-delta deposits in the Jeppestown Subgroup reflect both this change and its transition into the molasse-type Central Rand Group.

Numerous silicic volcanic clasts within the West Rand Group indicate that a volcanic arc terrane to the north was contributing detritus to the Witwatersrand Basin. The presence of ilmenite, fuchsite, and particularly chromite, in these sediments demonstrates that an ultramafic source (perhaps parts of an elevated greenstone belt) was also contributing detritus to the West Rand Group (cf. Haggerty, 1976; Stanton, 1972).

NATURE OF THE CENTRAL RAND GROUP

The Central Rand Group contains large amounts of molasse-type sediments disposed as sand and gravel bars in coalesced fluvial/alluvial fans. The West Rand/Central Rand Group division of Witwatersrand stratigraphy into a lower flysch-type sequence and an upper molasse facies is typical of rocks deposited in foreland basins and is unlike the stratigraphy of rift-basins, which commonly grade from a thick sequence of basal coarse clastics, typically of non-marine facies, to an upper, more mature, generally finer-grained, often marine sedimentary cover (cf. Burke et al., 1985a). Extensive mining of gold and uranium bearing placers has enabled the dendritic paleodrainage pattern into the

basin to be accurately mapped and points of entry into the basin have been located (Figure 15). The source of the Central Rand molasse was clearly a mountain range to the northwest because of both facies patterns and the coarsening and thickening of sediments in that direction. This mountain range (here named the Limpopo Mountains) contained a large amount of silicic volcanic deposits and we suggest that it was an Andean-type arc, locally preserved as the Kanye and Dominion silicic volcanic sections.

The growth of folds parallel to the basin margin during sedimentation (Winter, 1964; Toens et al., 1964), and the preferential filling of synforms by the Bird lava flows (Pretorius, 1964; Tankard et al., 1982), indicates that folding was in progress during Central Rand Group sedimentation. Deformation of this kind is in fact characteristic of flexural foreland basins (cf. Dickinson, 1974). Isopach plots reveal that during Witwatersrand sedimentation the depositional axis of the basin migrated considerable distances, generally away from the northwestern basin margin. However, there is some indication of a late northwestward migration (Figure 16), which it is tempting to interpret as a viscoelastic response of the lithosphere to loading, a relationship that has been reported from some foreland basins (Quinlan and Beaumont, 1984). Numerous unconformities developed within the Central Rand Group along the northwestern basin margin may be related to episodic uplift of that margin during deposition (Figure 17) but unconformities near the southeastern basin margin suggest a more complex relationship (Figure 17), including a possible interaction of a southeastward sedimentary progradation and uplift associated with viscoelastic relaxation of the lithosphere.

METAMORPHISM

Metamorphic mineral assemblages (eg. chlorite + muscovite + epidote) in Witwatersrand strata define the metamorphic grade as lower greenschist facies (cf. Miyashiro, 1979) although, locally, higher temperature metamorphic minerals have formed in areas marginal to igneous intrusions. Minerals indicative of exceptionally high pressures (eg., coesite, stishovite, diaplectic silica glass) have formed around the Vredefort structure are generally, though not universally, attributed to a 2.0 Ga impact event (Daly, 1947; Dietz, 1961; Grieve, 1982; Martini, 1978; Manton, 1965).

Hallbauer and Kable (1979) reported geochemical and fluid inclusion studies on Witwatersrand quartzites finding that all fluid inclusions show an homogenization temperature between 120°C and 180°C, suggesting "a common thermal influence" or low-grade metamorphic event. U-Pb isotopic data from the Witwatersrand sediments (Rundle and Snelling, 1977) and evidence from the overlying Ventersdorp succession (Cornell, 1978) suggests that this metamorphism occurred approximately 2.0 Ga ago, and may be related to the Bushveld event.

AGE OF THE WITWATERSRAND STRUCTURE

"Granitic" basement to the Witwatersrand structure has yielded some rather dubious Rb-Sr whole rock isochrons ranging from 2.7 - 3.1 Ga (re-calculated from Allsop *et al.*, 1964). Using leach solutions of sulfide concentrates Van Niekerk and Burger (1969) calculated a Pb-Pb age for the Dominion lavas of 2800+/-60 Ma; this is supported by a U-Pb age (zircon and apatite) of 2.83 Ga for the Dominion Group (Van Niekerk 1969). However, Rundle and Snelling (1977) suggest the adoption of 2740 +/-19 Ma, the "weighted mean age of the Schweizer Reneke granite and the Dominion lavas, using an assumed initial $^{87}Sr/^{86}Sr$ of 0.705, as the maximum age of the Witwatersrand sediments. We prefer not to use this

age, and take the other two remarkably consistent ages as the maximum age for the Witwatersrand sediments but there is clearly a need for further isotopic study. An age of approximately 2.8 Ga is our choice for a maximum age for the Witwatersrand Supergroup, but it should be noted that this implies that the younger Rb-Sr ages calculated for the granitic basement to this group are incorrect. A minimum age for the Witwatersrand Supergroup is given by the age of the Ventersdorp Supergroup which overlies Witwatersrand strata in many areas. Van Niekerk and Burger (1978) calculated a U-Pb age of 2.64+/-.08 Ga for zircons extracted from silicic volcanics of the Ventersdorp Group, supporting their (Van Niekerk and Burger, 1964) earlier age determination (U-Pb on zircons) of 2.63+/-.1 Ga from a correlative unit farther north in the Ventersdorp Supergroup. Deposition of the Witwatersrand Supergroup is therefore constrained between 2.8 and 2.64 Ga ago, over a maximum interval of approximately 150 Ma indicating a duration of deposition not unlike that of some Phanerozoic foreland troughs. Further isotopic work using, for instance, the recently perfected high-precision U-Pb techniques (Krogh, 1982a,b) needs to be done before the age and interval of deposition of the Witwatersrand basin can be better-constrained.

STRUCTURE OF THE WITWATERSRAND BASIN

Inward dipping sedimentary strata dominate the structure of the Witwatersrand basin, with the dips characteristically being much greater on the northwestern margin of the basin than on the southeastern margin (excluding deformation associated with the Vredefort Structure). Where preserved, the northwestern basin margin is a large steep fault zone that brings granitic-gneissic basement rocks (eg., the Johannesburg "dome") on the north into contact with steeply dipping to overturned Witwatersrand strata on the south (Figure 3). The dips of the Witwatersrand beds adjacent to this fault zone decrease from vertical and overturned at the base to only 20^{o} at the top

of the Witwatersrand succession (Pretorius, 1981), demonstrating that this is a thrust fault that was active during deposition of the sediments. Pretorius (1976) has suggested that the (thrust-bounded) northwestern basin margin migrated 60 km to the southeast during deposition of the Witwatersrand strata, with a concomitant 10 km shift of the depositional axis of the basin in the same direction (see also Figure 16). In a pioneer structural analysis of Witwatersrand strata, Fripp and Gay (1972) concluded that folds and cleavage in Hospital Hill series strata near Johannesburg formed under the influence of a north-south compressive stress, while similarly oriented compressive stresses were invoked by Roering (1968) for the formation of cleavage in argillaceous Witwatersrand strata around the Florida-Krugersdorp area.

Faults and folds within the Witwatersrand basin are closely related and some, particularly those disturbing the West Rand Group, demonstrably formed during deposition of the sediments. Cluver (1957) noted that molasse of the Central Rand Group is not disturbed by numerous major faults while sediments of the underlying West Rand Group were considerably more disrupted.

Excluding deformation associated with the Vredefort Structure there are two major fold trends within the Witwatersrand Basin. The earlier of these is oriented roughly northeast-southwest and parallels the curving northwestern margin of the basin. These folds are locally associated with similarly-oriented thrust faults. However, Pretorius (1981) has suggested that these faults represent re-activated normal faults, with the north side being the downthrown block. Early (normal) movement on these faults is believed to have accomodated some of the subsidence of the basin floor. Normal faults of this sort are common in Phanerozoic foreland basins (cf. Chadwick, 1917; Bradley et al., 1985 a,b). Along with the development of new faults many of these normal faults became re-activated as thrusts with a north over south sense of displacement

(Pretorius, 1981).

Several later deformation episodes have affected the Witwatersrand strata. A second fold generation trends northwest and refolds earlier formed structures. Numerous (dominantly) right lateral strike-slip faults have also locally reactivated older faults (Pretorius, 1981). Abundant north to northeast trending normal faults are associated with the circa 2.6 Ga Ventersdorp rifting episode (Burke et al., 1985b).

ANDEAN MARGIN TECTONICS?

The Witwatersrand Basin exhibits many features that are characteristic of foreland basins, including; (1) an asymmetric profile with thicker strata and steeper dips toward the mountainous flank, (2) a basal "flysch" facies overlain by "molasse"-type sediments, (3) one side of the basin is bound by thrust faults marginal to a contemporaneous mountain belt, (4) the basin margin nearest the mountain range displays many unconformities related to the cannibalization of the sediments and, (5) there is a general scarcity of magmatic activity in the basin.

Compressional deformation was clearly in part syn-sedimentary and the associated folds and faults trend parallel to the depositional axis of the basin. Pretorius (1981) has suggested a basinward migration of the margin nearest the orogen (northwestern), and a similar migration of the depositional axis of the basin. The faulting history at most locations in the basin changes from early normal faulting, with the downthrown block towards the mountain range, to later thrust faulting, with the uplifted block toward the mountain range; by analogy with Phanerozoic foreland basins (eg. Chadwick, 1971; Bradley, 1983; Bradley and Kusky, 1985) we suggest that there may also have been a progressive migration of these structural elements across the basin.

Phanerozoic foreland basins have often been quantitatively examined, yielding models for the rheology, thickness, and thermal state of the lithosphere during flexure, as well as information about the loading process, including the shape and size of the peripheral mountain range (eg. Watts et al., 1982; Beaumont, 1981; Jordan, 1981; Porter, 1982; Menke, 1981; Forsyth, 1979). Problems arise, however, in attempting to quantitatively model the flexural history of the Witwatersrand. These problems are primarily the result of the poorly constrained ages of individual stratigraphic horizons; for instance, deposition of the entire Witwatersrand Supergroup is only constrained between about 2.8 and 2.65 Ga ago and it may have taken much less time. It is also disappointing that the south-eastern part of the basin although, from sedimentological evidence, close to the original limit of deposition does not yield as complete a record of forebulge evolution as has been discerned in some younger basins. The number of unknowns among the variables which would be needed to accurately model the flexural behavior of the Witwatersrand Basin precludes any such formal analysis (P. Morgan, personal communication; T. Jordan, personal communication). However, the width of the basin and the shape of the sedimentary wedges in the Witwatersrand (Figures 5-14) fall within the range of variation of Phanerozoic examples (P. Morgan, personal communication), and do not suggest any significant change in the flexural response of the lithosphere to loading between Archean and Phanerozoic times.

Stratigraphic relationships with the underlying Dominion Group, the presence of silicic volcanic clasts throughout the Witwatersrand stratigraphy, and the presence of minor mafic lava flows in the Witwatersrand lead us to suggest that the foreland basin was developed behind a continental margin volcanic arc, and as such it is a "retroarc" basin (Jordan, 1981). Figure 18

shows a map view, and Figure 19 is a schematic cross-section, showing the tectonic scenario which we envisage for the Kaapvaal and Zimbabwe Cratons during early formation of the Witwatersrand Basin. Sediments of the West Rand and, possibly, the Central Rand Group are interpreted as deposited in an actively subsiding foreland trough developed adjacent to an Andean margin and fold-and-thrust belt, represented by the Kanye and Dominion volcanics, and the Thabazimbi Fault and related structures, respectively, Terry Jordan has pointed out to us (personal communication, 1984) that stratigraphic relationships, lithologies, and sedimentary structures in the Witwatersrand Basin are remarkably similar to those in the Sierra de Huaco Basin (cf. Johnsson et al., 1984), which lies above a shallow subduction segment (27^{o}-33^{o}S) in the southern Andean foreland (Eastern Precordillera Province of Ortiz and Zambrano, 1981). Each basin contains a lower flysh sequence overlain by molassic sediments, both of which are rich in silicic volcanic detritus.

The Crown and Bird mafic volcanic flows are important stratigraphic horizons in the Witwatersrand Basin. Mafic lava flows are rare, but not unknown, from recent retroarc basins, and analagous modern tectonic environments may be found, again, in the southern and central Andes (Figure 20). The precise "mechanism" for the eruption of mafic lavas in a foreland basin is not well-defined at present, but is perhaps related to changes in subduction geometry as envisioned by Dewey (1980) or, perhaps, to a steepening of the subduction angle. Retroarc (marginal to active subduction-related arcs) and peripheral (marginal to collisional mountain belts) foreland basins (Jordan, 1981) may show very similar stratigraphic histories, and the presence of lavas in the section may prove to be useful way to discriminate between the two different types of basin.

Unconformities along the northwestern margin of the Witwatersrand Basin indicate episodic uplift of that margin, while overall stratigraphic trends indicate some shrinking in area during the later stages of deposition (Figures 5-14). This is not inconsistent with a viscoelastic relaxation of the lithosphere.

CONTINENTAL COLLISION: THE TERMINATION OF ANDEAN MARGIN EVOLUTION

Several authors have suggested that the high grade metamorphism and deformation in the Limpopo Province (Figure 18) is largely a result of a collision between the Andean margin of the Kaapvaal Craton with a (perhaps passive) continental margin developed on the Zimbabwe Craton (Dewey and Burke, 1973; Light, 1982; Kidd, 1984; Burke et al., 1985b; Eriksson and Kidd, in prep.). This continental collision has been interpreted as having been well-underway by 2.64 Ga ago, when the Ventersdorp rifting event, which succeeds the Witwatersrand, was initiated perhaps in an impactogenal manner (Burke et al., 1985b).

Continental collision zones are typically associated with broad uplifted plateaus (cf. Dewey and Burke, 1973), of which Tibet is the type example. Such collisional plateau uplifts are bound on the cratonward side by foreland basins; examples include the Kopet Dagh foredeep exposed northeast of the Iran/Arabia collision zone (Berberian and Berberian, 1981) and the southern sides of the Tarim and Tsaidam Basins north of the Himalaya (Molnar and Tapponnier, 1975). This class of foreland basin, related to collisional plateau uplift, should not be confused with either peripheral foreland basins, such as the Ganges Basin south of the Himalaya, which formed on the site of the former subduction zone on the opposite side of the orogen, or with retroarc basins on the continental side of Andean arcs. In many old terranes discrimination among

the three types of foreland basin is not possible because of later destruction or burial. The northern margin of the Kaapvaal Craton is, however, exceptional.

The question of whether or not the Zimbabwe/Kaapvaal continental collision was associated with an uplifted collisional plateau has not yet been fully addressed. Metamorphic mineral assemblages from the central part of the Limpopo Province indicate burial to greater than 30 km (Light, 1982), implying double crustal thickness during peak metamorphism at 2.7 - 2.6 Ga (Tankard et al., 1982). The Bulai Gneiss, Razi Granite, Matok Pluton, and numerous other intrusive rocks represent syn-tectonic anatectic granites (Light, 1982; Robertson, 1973a, 1973b) which could be related to continental collision. These granites intrude high grade rocks in the area north of the Witwatersrand Basin and we point out that much of the area between the Limpopo Province and the Witwatersrand Basin, particularly the region north of the Thabazimbi Fault (Figure 18), bears a strong resemblance to a deeply eroded collisional plateau uplift.

By analogy, it is possible that some of the rocks in the Witwatersrand Basin preserve a record of the Kaapvaal/Zimbabwe collisional event and may, in part, represent one of those foreland basins developed on the cratonward edge of the uplifted collisional plateau. Because the isotopic ages of 1) Andean arc magmatism, 2) continental collision and, 3) sedimentation in the Witwatersrand Basin are at present poorly constrained, and because the stratigraphic histories of both types of foreland basins may be confusingly similar, it is hard to say what part of the Witwatersrand Basin is related to Andean convergence, and what part, if any, is related to continental collision. Collisional plateau foreland basins may differ from their Andean margin counterparts in several features, the most notable of which is perhaps their typical exclusion of marine sediments by virtue of elevation above sea level.

A major change in the depositional style in the Witwatersrand Basin occurs essentially at the break between the West Rand and Central Rand Groups (Figure 2). The West Rand Group contains a marine flysch-type sequence, which shows an increase in the amount of terrigenous clastics in its upper portions. In contrast, the Central Rand Group contains large amounts of non-marine molasse rocks whose sedimentary features are similar to those of young collision-related foreland trough molasse at Karamai in the Junggar basin (Chang Chiyi, 1981). It is possible that this break separates the Andean margin (retroarc) and collisional plateau related phases of foreland basin evolution. If future isotopic dating of the Crown and Bird lavas (using, for instance, the Sm-Nd technique) proves them to be contemporaneous with the initiation of continental collision in the Limpopo Province, then the Witwatersrand Basin may be shown to preserve two episodes of foreland basin evolution; a retroarc basin in the West Rand Group (Figure 19), and a collisional plateau-related foreland trough in the Central Rand Group (Figure 20). If, alternatively, the Crown and Bird lavas are shown to be older than the proposed collision, then all of the Witwatersrand Basin is (perhaps) related to subduction driven Andean arc activity. A third and less likely possibility is that all of the Witwatersrand Basin is relatable to the Kaapvaal/Zimbabwe continental collision.

CONCLUSION

Sedimentary rocks of the Witwatersrand Supergroup form an elongate trough of Archean age (2.8 - 2.65 Ga) on the Kaapvaal Craton of southern Africa, and are locally underlain by a volcanosedimentary sequence the Dominion Group. It has been suggested that this sequence resembles a continental rift presumably because it contains both sedimentary and volcanic rocks (see for example; McKenzie et al., 1980; Bickle and Eriksson, 1982; Nisbet, 1984) but the volcanics are not compositionally typical of rift volcanic rocks and the

Witwatersrand sediments do not clearly show the volcanic free, widely distributed thermal relaxation phase of sedimentation typical of rift sequences.

We suggest as an alternative hypothesis that the Witwatersrand Basin is a foreland trough of Archean age which developed in two stages; the first on the cratonward side of a subduction-related Andean arc built on the edge of the Kaapvaal Craton and the second on the cratonward side of an uplifted plateau that resulted from the collision of this Andean arc with the Zimbabwe Craton, approximately 2.7 Ga ago. The Witwatersrand basin is interpreted as analagous in its earlier history to Phanerozoic retroarc foreland basins, such as the Sierra de Huaco Basin in the southern Andes and, in its later stages, to basins such as the Tarim, and the Tsaidam, north of the Tibetan Plateau. Stratigraphically, structurally and tectonically the Witwatersrand Basin is so similar to other, more recent, foreland basins that it appears unneccessary to consider that Archean convergent tectonic processes were much different from those of today. Because the Witwatersrand Basin has been the subject of so much intense geological research over the last century it seems likely that additional ways of testing the idea that the basin is a foreland trough will emerge. For example: it might be worth assessing whether the trace element concentrations of Dominion Group lavas were similar to those of younger Andean lavas.

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FIGURE CAPTIONS

Figure 1. Generalized map of southern Africa showing the location of the Witwatersrand Basin on the Kaapvaal Craton, as well as the distribution of other selected Precambrian geologic elements. Note that the Witwatersrand Basin is elongated parallel to, but located some distance south of the high-grade Limpopo Province (map modified after Tankard et al., 1982).

Figure 2. Generalized stratigraphic column for the Witwatersrand Supergroup. The quoted thicknesses are maximum values which are found near the northwestern basin margin, while the strata thin towards the southeastern basin margin.

Figure 3. Map of the Witwatersrand Basin showing locations referred to in text, outcrop and subsurface distribution of Witwatersrand strata, locations of exposed basement, and the generally fault-bounded northwestern basin margin (map compiled from Pretorius, 1976 and Button, 1977).

Figure 4. Distribution, thickness, and lithologies in the largely-volcanic Dominion Group (after S.A.C.S., 1980 and Tankard et al., 1982). Symbol K shows the location of Klerksdorp.

Figure 5. Isopach map of the Witwatersrand Basin showing the thickness (in meters) and distribution of the Hospital Hill Subgroup. Figures 5 to 14 redrawn after Brock and Pretorius, 1964.

Figure 6. Shape of the Hospital Hill Subgroup sediment wedge determined from isopachs.

Figure 7. Isopach map of the Witwatersrand Basin showing the thickness and distribution of the Government Subgroup.

Figure 8. Restored cross-sections of the Witwatersrand Basin for time immediately after deposition of Government Subgroup.

Figure 9. Isopach map of the Witwatersrand Basin showing thickness and distribution of Jeppestown Subgroup.

Figure 10. Restored cross-sections showing shape of sediment wedge in Witwatersrand Basin immediately after deposition of Jeppestown Subgroup.

Figure 11. Isopach map of the Witwatersrand Basin showing the thickness and distribution of the Johannesburg Subgroup.

Figure 12. Restored cross-sections of the Witwatersrand Basin showing shape of sediment wedge immediately after deposition of the Johannesburg Subgroup.

Figure 13. Isopach map of the Witwatersrand Basin showing the thickness and distribution of the Turffontein Subgroup.

Figure 14. Restored cross-section of the Witwatersrand Basin showing the shape of the sediment wedge immediately after deposition of the Turffontein Subgroup.

Figure 15. Paleocurrent directions derived from upper Witwatersrand strata showing that sediments of the Central Rand Group largely prograded into the basin from the northwestern margin in the form of alluvial and fluvial fans. Some transport of sediments into the basin from its southwestern margin is indicated by paleoccurents from Welkom, and some along axis transport has also been noted from sediments between Kroonstad and Klerksdorp. Paleocurrent directions are from Brock and Pretorius, 1964.

Figure 16. Location of the depositional axis of the Witwatersrand Basin during (1) Hospital Hill, (2) Government, (3) Jeppestown, (4) Johannesburg, and (5) Turffontein times. This axis generally, but not universally, migrated towards the southeast during deposition of the Witwatersrand strata, possibly suggesting an interplay of elastic flexure of the lithosphere, elasticoviscous relaxation, and sedimentary progradation.

Figure 17. Schematic vertical cross-sections of upper Witwatersrand strata from near the northwestern (A) and southeastern basin margins (B). Section B is from Tankard et al., 1982.

Figure 18. Map showing the location of the Witwatersrand Trough in relation to the Limpopo continental collision zone, the Gaberone Granite Complex, and numerous shear zones attributed to uplift and sideways motion of the region to the north of the Witwatersrand Basin. Anatectic granites (in black) are related to the collision between the Kaapvaal and Zimbabwe Cratons. The less intensely ornamented basement marks the future site of Ventersdorp rifting. Map based on data compiled from Tankard et al., (1982); Visser et al., (1976); DuToit et al., (1983); Pretorius, (1976); and Button, (1981).

Figure 19. Tectonic elements suggested for the Kaapvaal and Zimbabwe Cratons during deposition of the lower Witwatersrand Supergroup. The present northern margin of the Kaapvaal Craton includes an Archean Andean margin; the volcanic arc is represented by the Kanye Volcanics, Gaberone Granite Complex, the Dominion Volcanics, and lateral equivalents. Rocks of the Witwatersrand Supergroup (West Rand Group) were deposited in a foreland basin bound on the north by an active fold and thrust belt. All rocks within the "future Limpopo collisional zone" became highly deformed and metamorphosed when the Kaapvaal and Zimbabwe Cratons collided approximately 2.7 Ga ago.

Figure 20. Comparison of areas of mafic lava flows in the Witwatersrand Basin (Crown and Bird lavas) with distribution of mafic lava flows in the Andean foreland basin of the southern Andes (map B modified from Reunion de la Carte geologique de L'Amerique du Sud, 1963).

Figure 21. Cross section through the Kaapvaal and Zimbabwe Cratons across the Limpopo Province, a product of the collision between the two cratons at approximately 2.7 Ga ago. It is suggested in this diagram that the upper Witwatersrand strata were deposited in a foreland basin on the cratonward side of an uplifted plateau (similar to Tibet) related to the Kaapvaal/Zimbabwe collision.

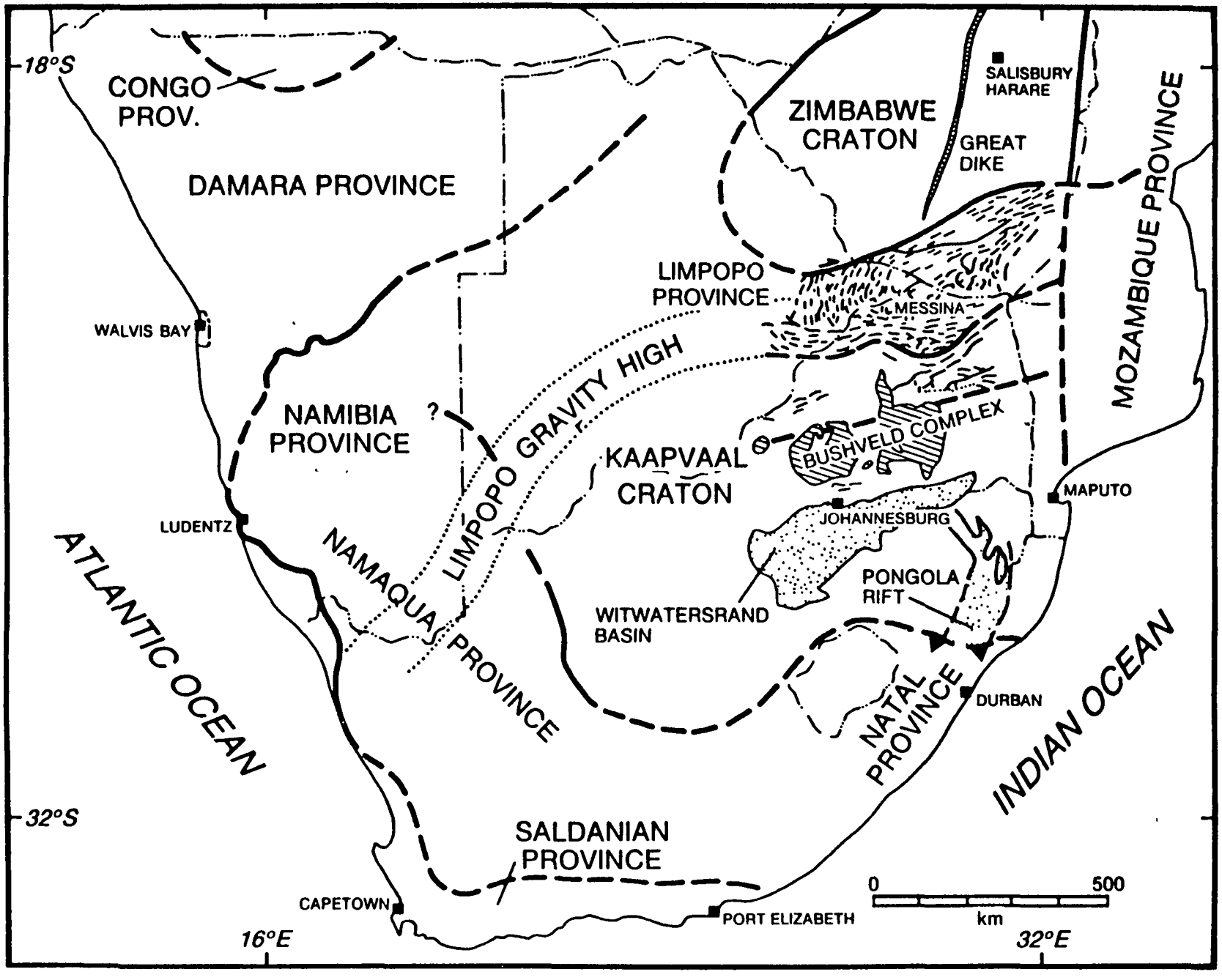

18°S
CONGO PROV.
DAMARA PROVINCE
ZIMBABWE CRATON
SALISBURY HARARE
GREAT DIKE
MOZAMBIQUE PROVINCE
LIMPOPO PROVINCE
MESSINA
WALVIS BAY
NAMIBIA PROVINCE
?
LIMPOPO GRAVITY HIGH
BUSHVELD COMPLEX
KAAPVAAL CRATON
MAPUTO
JOHANNESBURG
LUDENTZ
ATLANTIC OCEAN
NAMAQUA PROVINCE
PONGOLA RIFT
WITWATERSRAND BASIN
NATAL PROVINCE
DURBAN
INDIAN OCEAN
32°S
SALDANIAN PROVINCE
CAPETOWN
PORT ELIZABETH
0
500
km
16°E
32°E

Fig 1

GENERALIZED STRATIGRAPHY

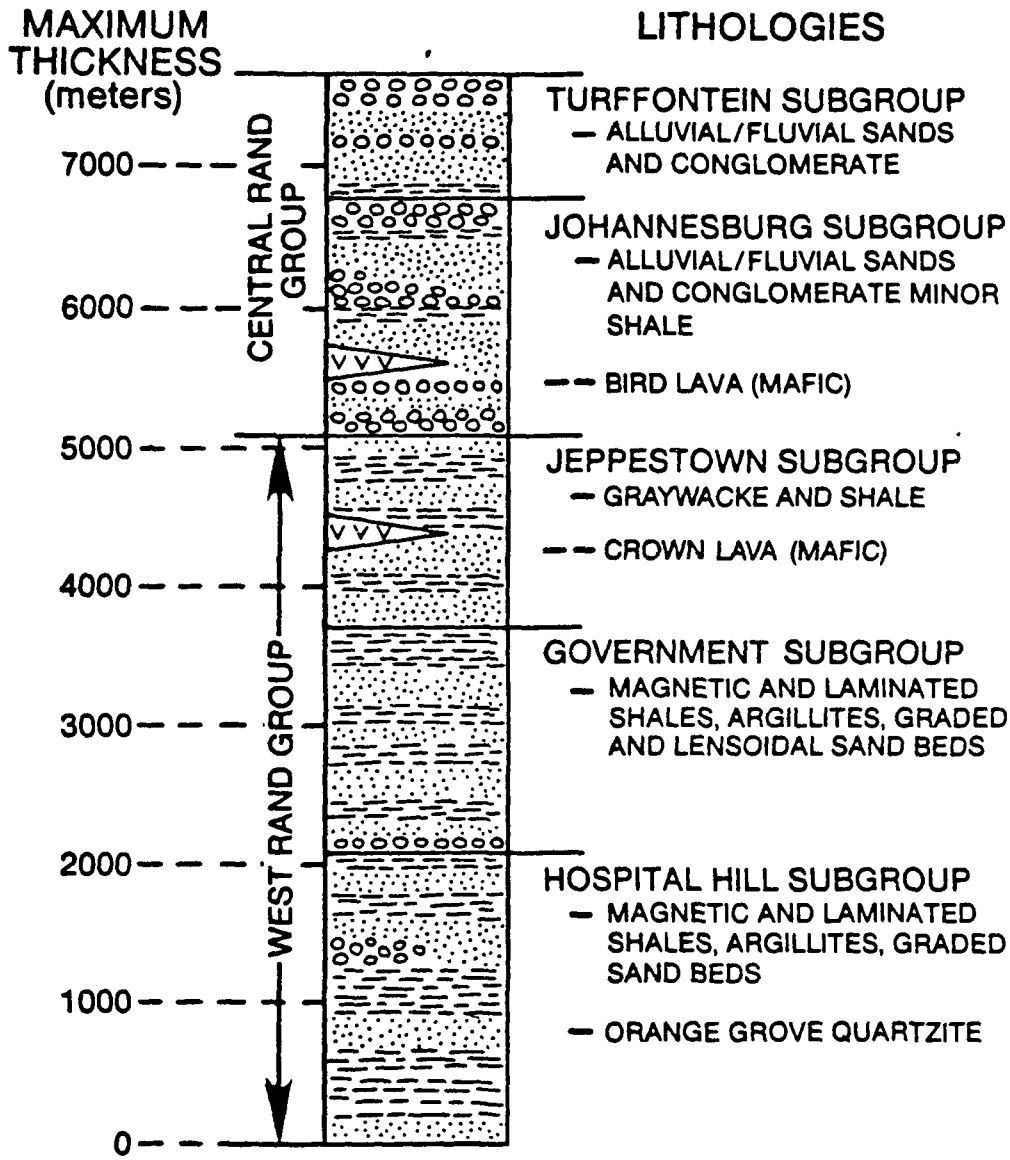


Fig 2

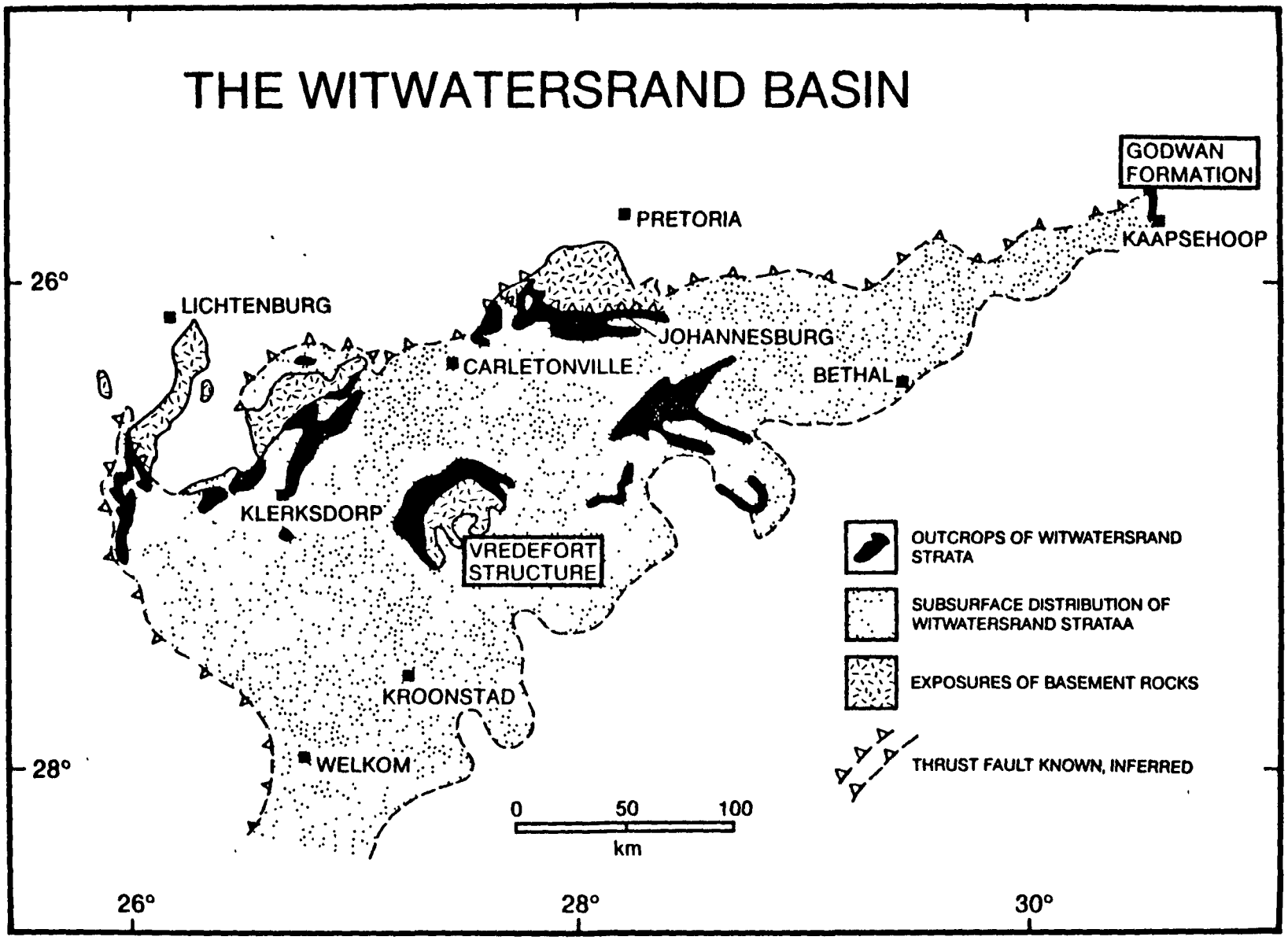

THE WITWATERSRAND BASIN
GODWAN FORMATION
KAAPSEHOOP
PRETORIA
26°
LICHTENBURG
JOHANNESBURG
CARLETONVILLE
BETHAL
KLERKSDORP
VREDEFORT STRUCTURE
KROONSTAD
28°
WELKOM
0 50 100
km
26°
28°
30°
OUTCROPS OF WITWATERSRAND STRATA
SUBSURFACE DISTRIBUTION OF WITWATERSRAND STRATAA
EXPOSURES OF BASEMENT ROCKS
THRUST FAULT KNOWN, INFERRED

Fig 3

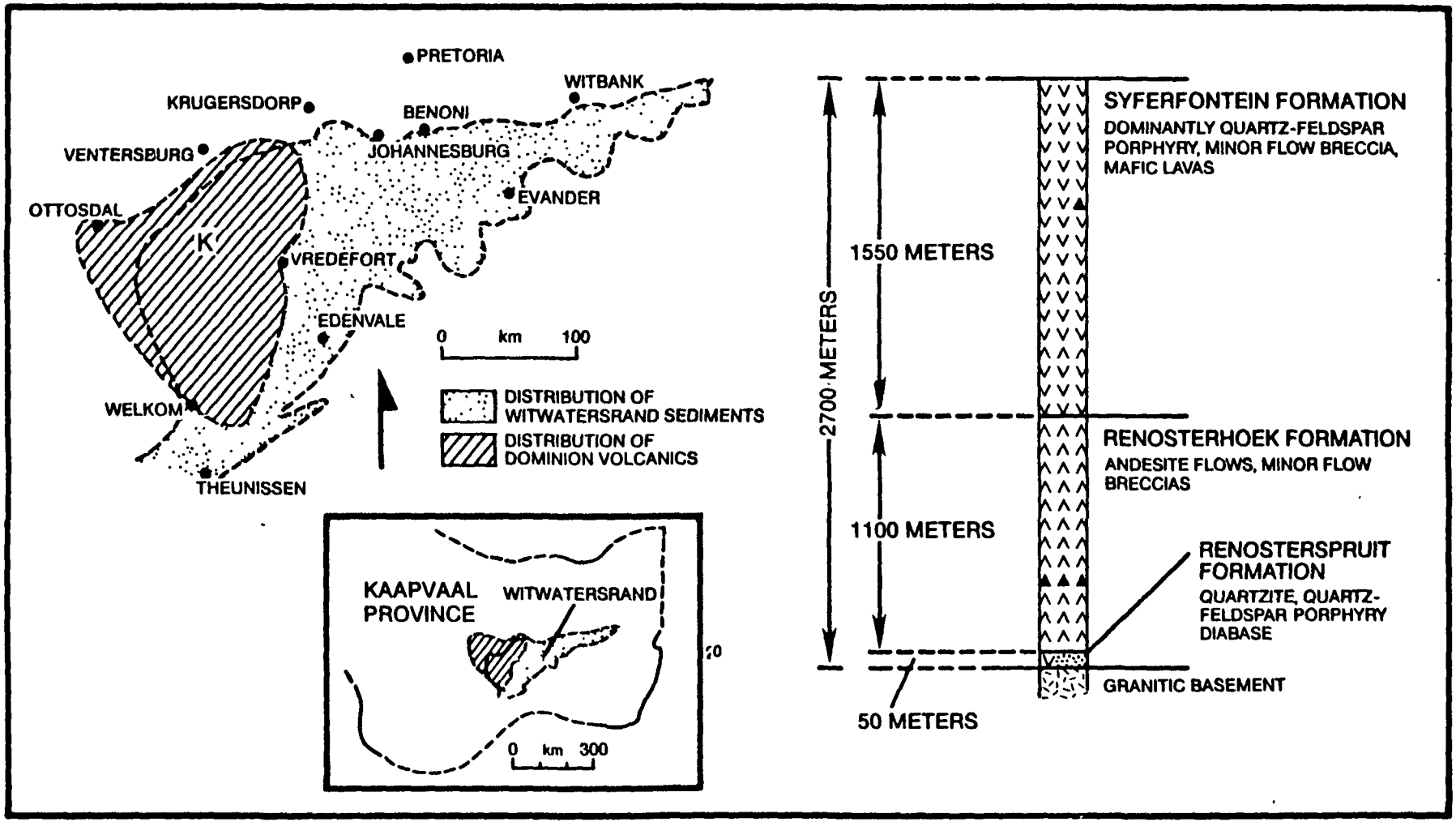

PRETORIA
WITBANK
KRUGERSDORP
BENONI
VENTERSBURG
JOHANNESBURG
EVANDER
OTTOSDAL
K
VREDEFORT
EDENVALE
0 km 100
WELKOM
DISTRIBUTION OF WITWATERSRAND SEDIMENTS
DISTRIBUTION OF DOMINION VOLCANICS
THEUNISSEN
KAAPVAAL PROVINCE
WITWATERSRAND
0 km 300
2700 METERS
1550 METERS
1100 METERS
50 METERS
SYFERFONTEIN FORMATION
DOMINANTLY QUARTZ-FELDSPAR PORPHYRY, MINOR FLOW BRECCIA, MAFIC LAVAS
RENOSTERHOEK FORMATION
ANDESITE FLOWS, MINOR FLOW BRECCIAS
RENOSTERSPRUIT FORMATION
QUARTZITE, QUARTZ-FELDSPAR PORPHYRY DIABASE
GRANITIC BASEMENT

Fig 4

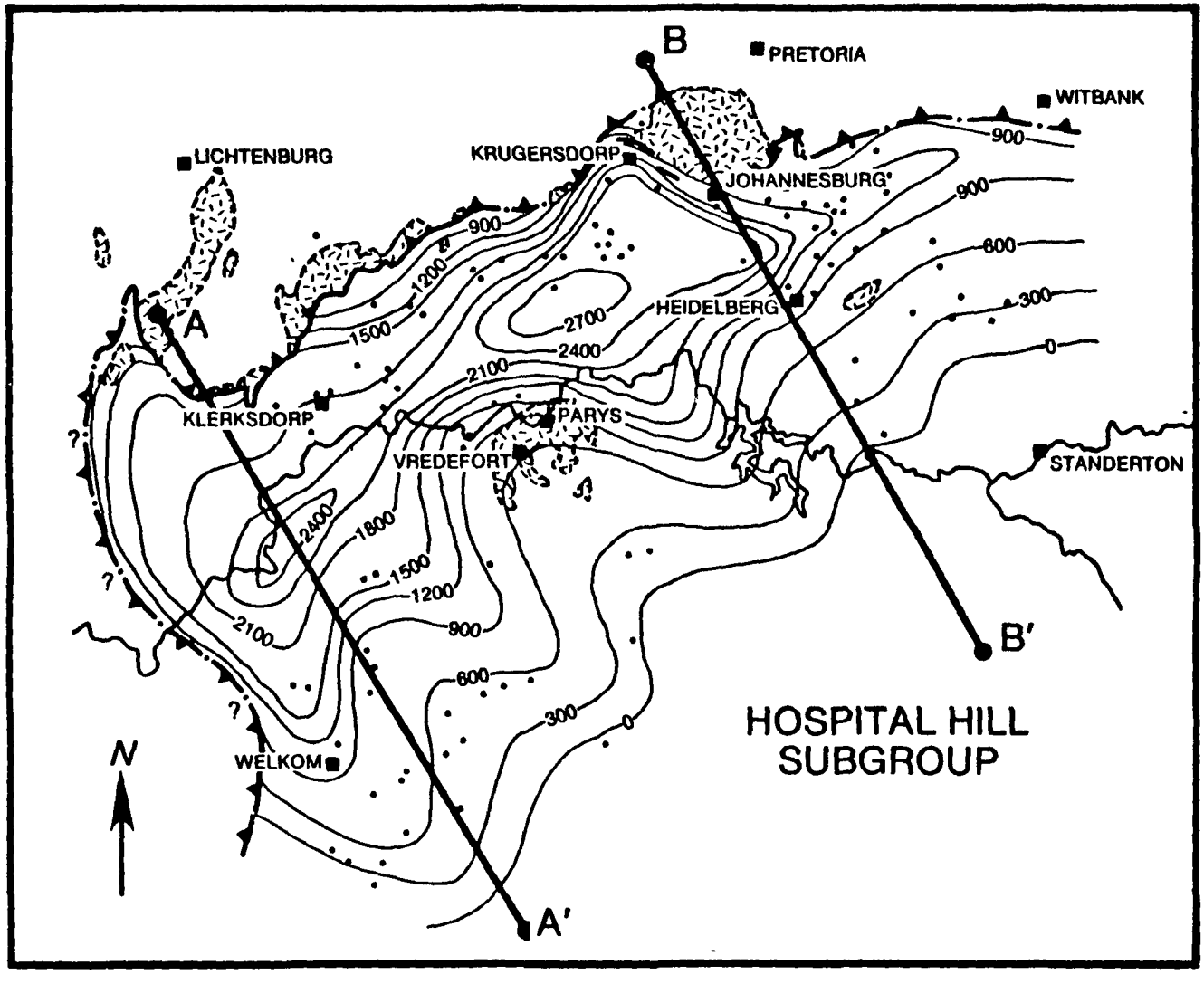

B
PRETORIA
WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
900
900
600
300
0
1200
1500
2700
2400
2100
HEIDELBERG
A
KLERKSDORP
PARYS
VREDEFORT
STANDERTON
2400
1800
1500
1200
900
600
300
0
2100
B'
HOSPITAL HILL
SUBGROUP
N
WELKOM
A'

Fig 5

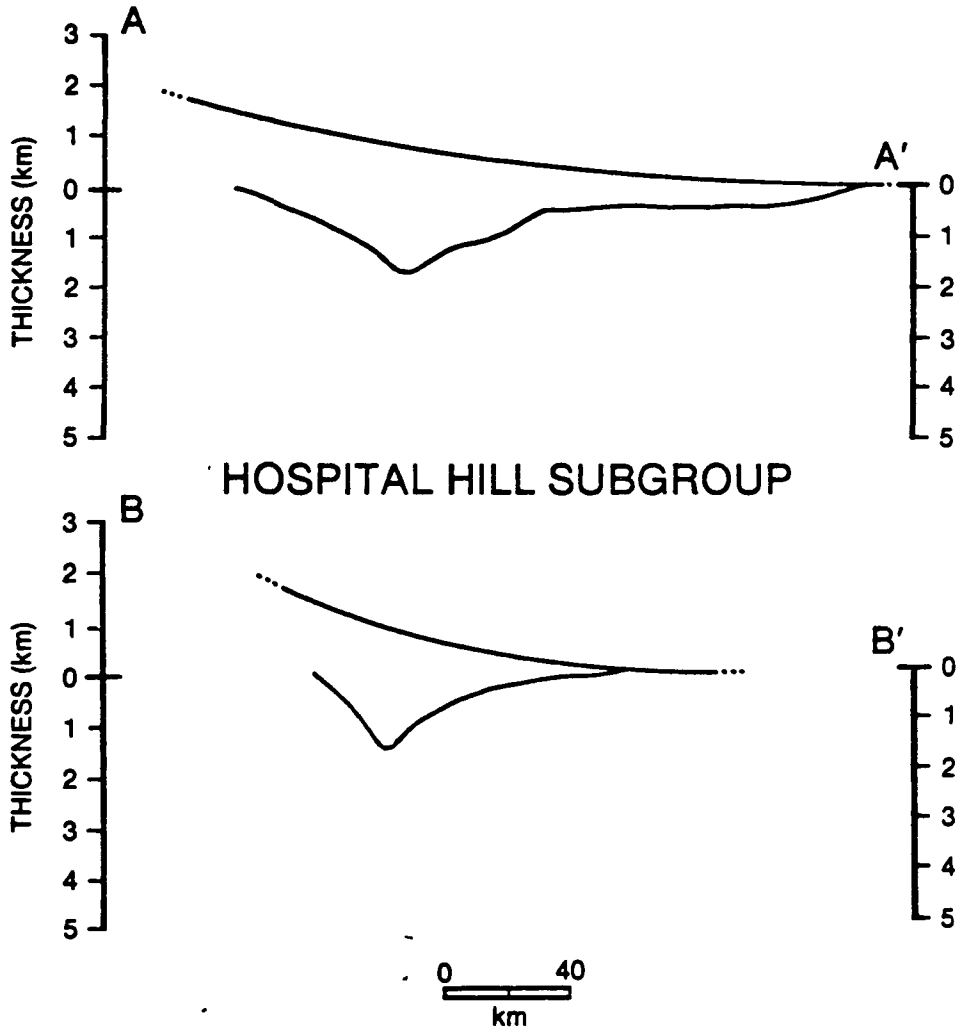

A
A'
THICKNESS (km)
3
2
1
0
1
2
3
4
5
HOSPITAL HILL SUBGROUP
B
B'
THICKNESS (km)
0
40
km

Fig 6

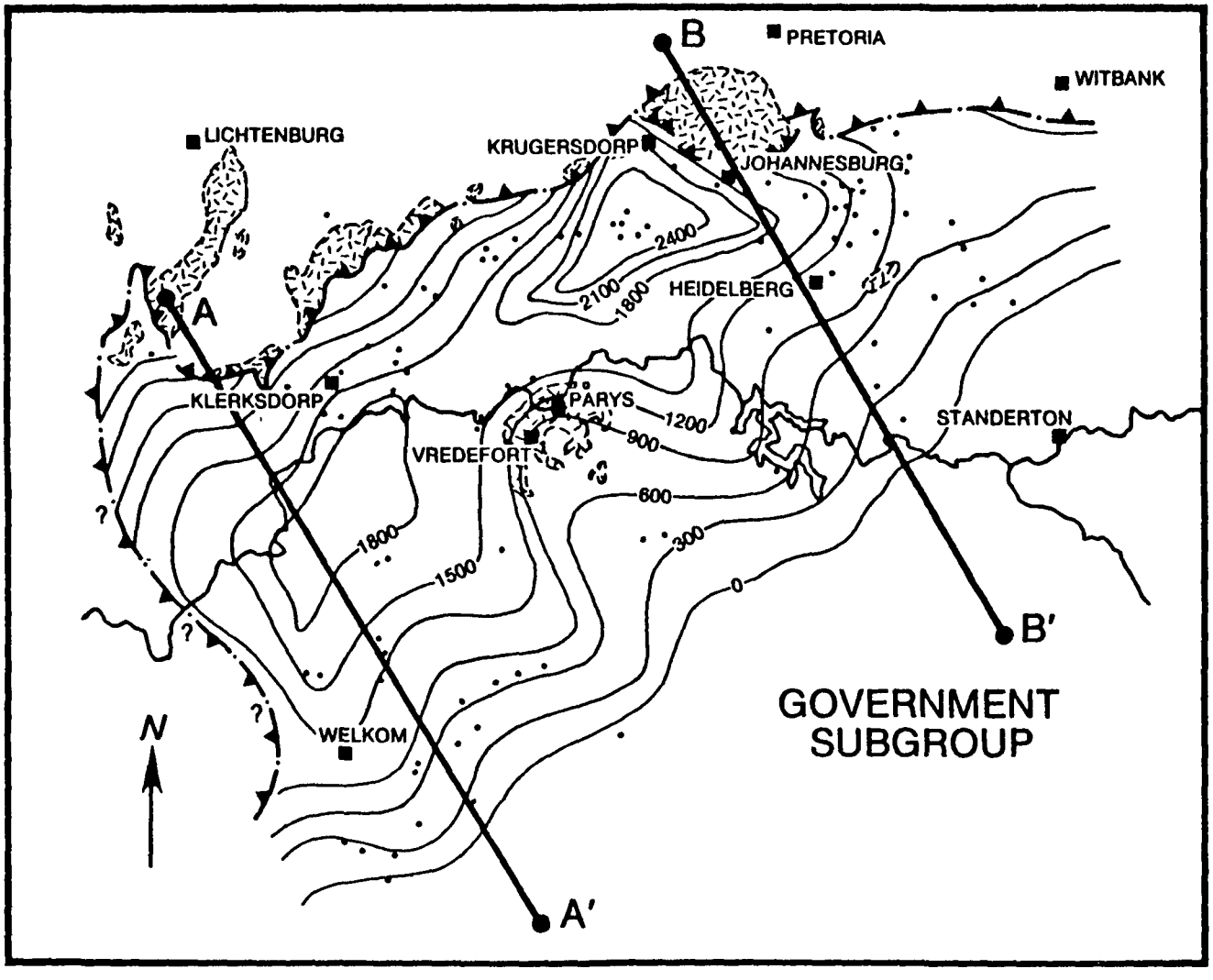

B
PRETORIA
WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
2400
2100
1800
HEIDELBERG
A
KLERKSDORP
PARYS
1200
900
STANDERTON
VREDEFORT
600
1800
300
1500
0
B'
GOVERNMENT
SUBGROUP
N
WELKOM
A'

Fig 7

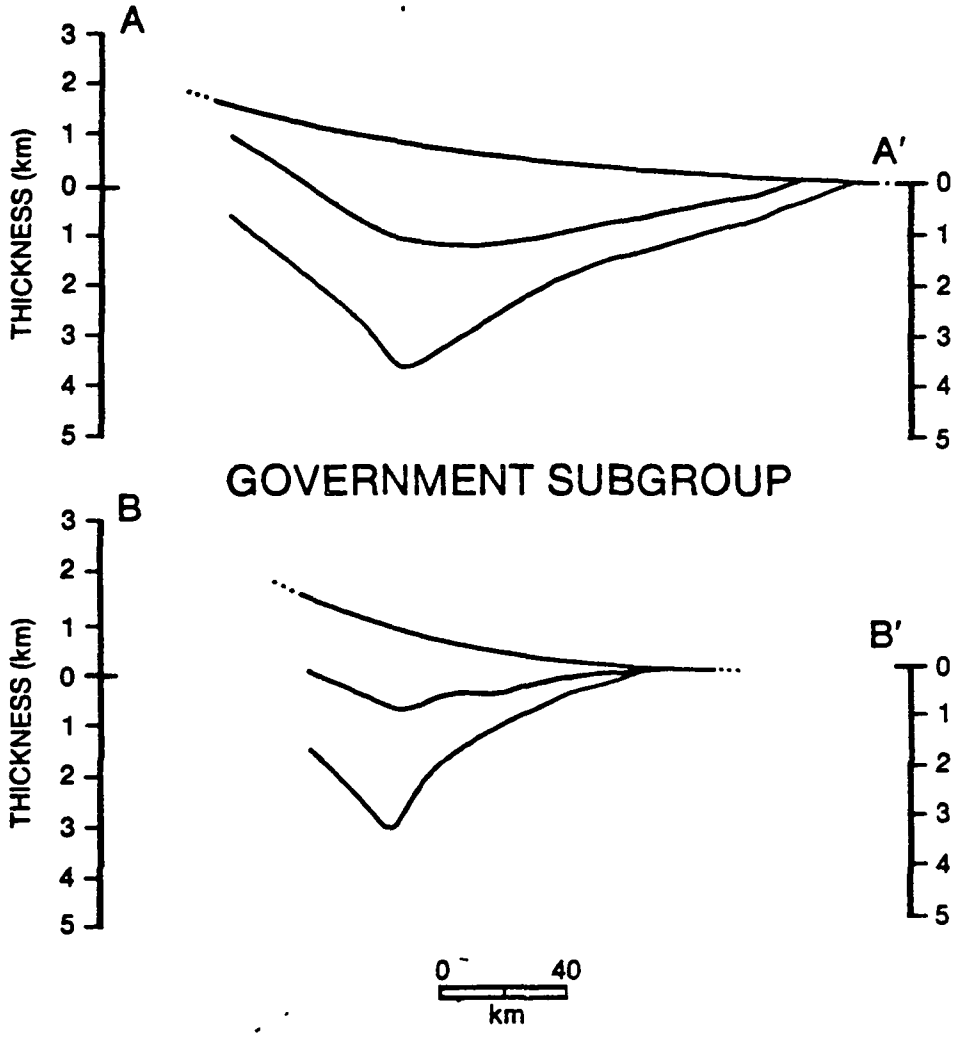


Fig 8

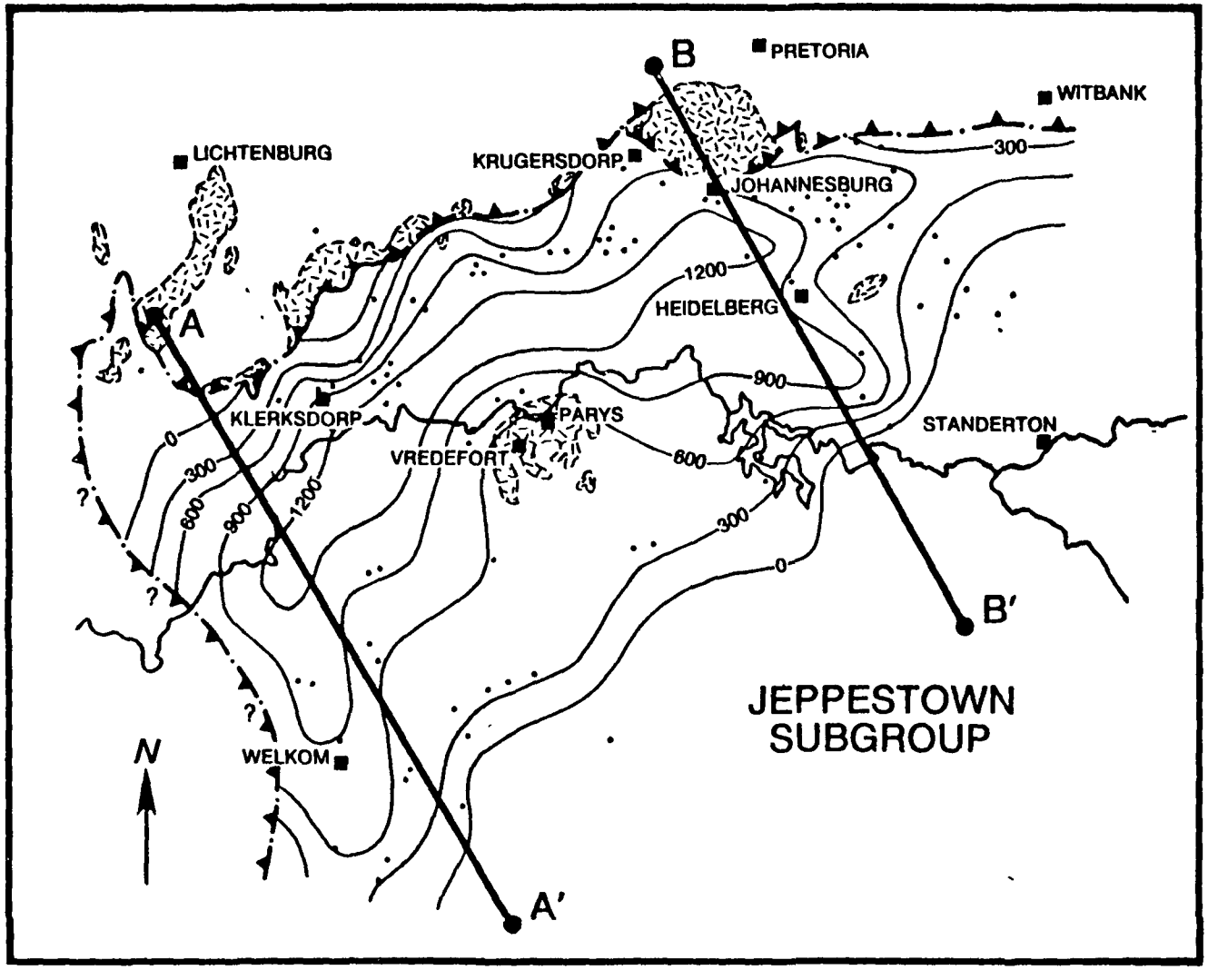

B
PRETORIA
WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
300
1200
HEIDELBERG
A
900
KLERKSDORP
PARYS
STANDERTON
0
300
600
VREDEFORT
600
900
1200
300
0
B'
JEPPESTOWN
SUBGROUP
N
WELKOM
A'

Fig 9

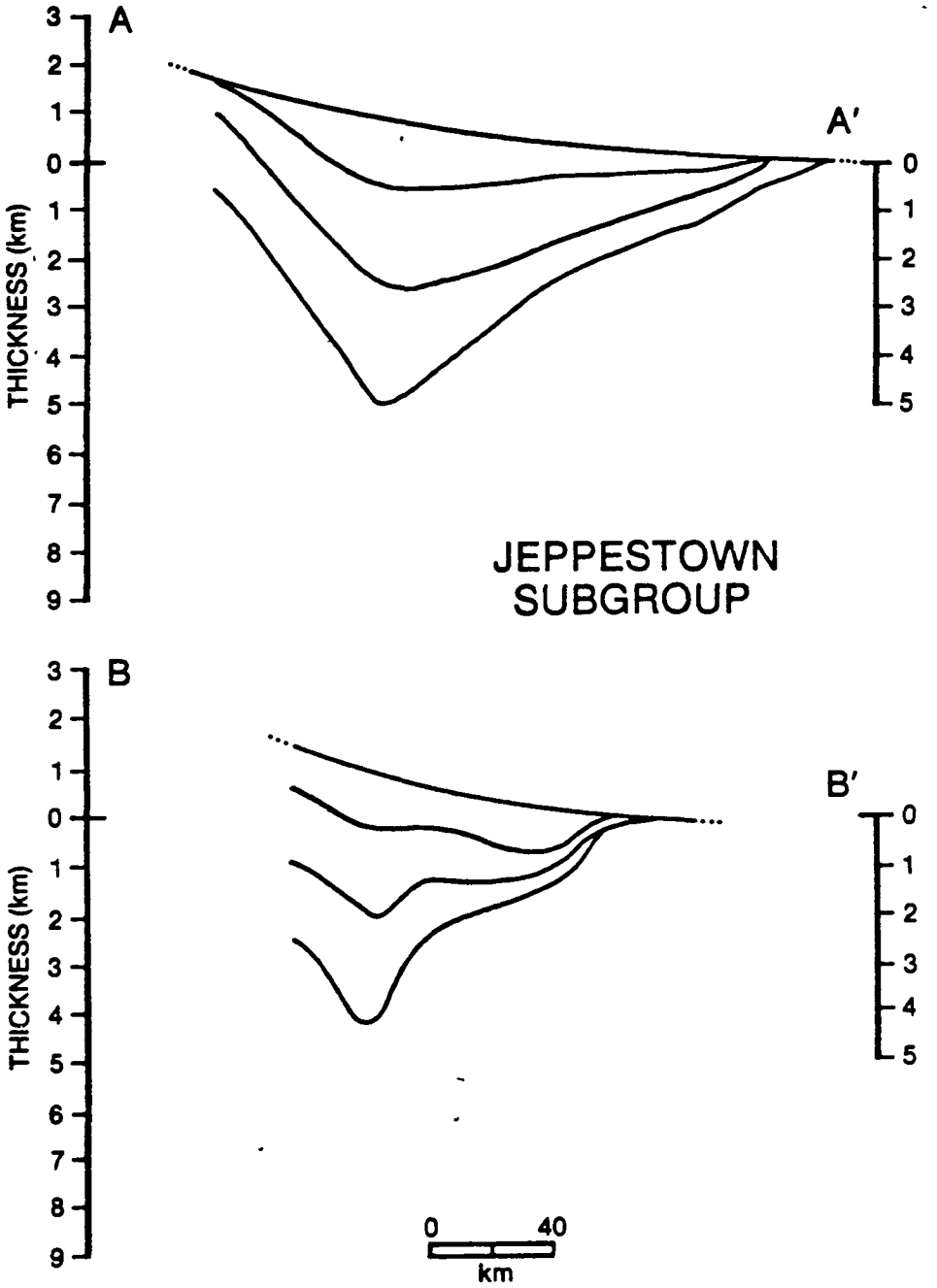


Fig 10

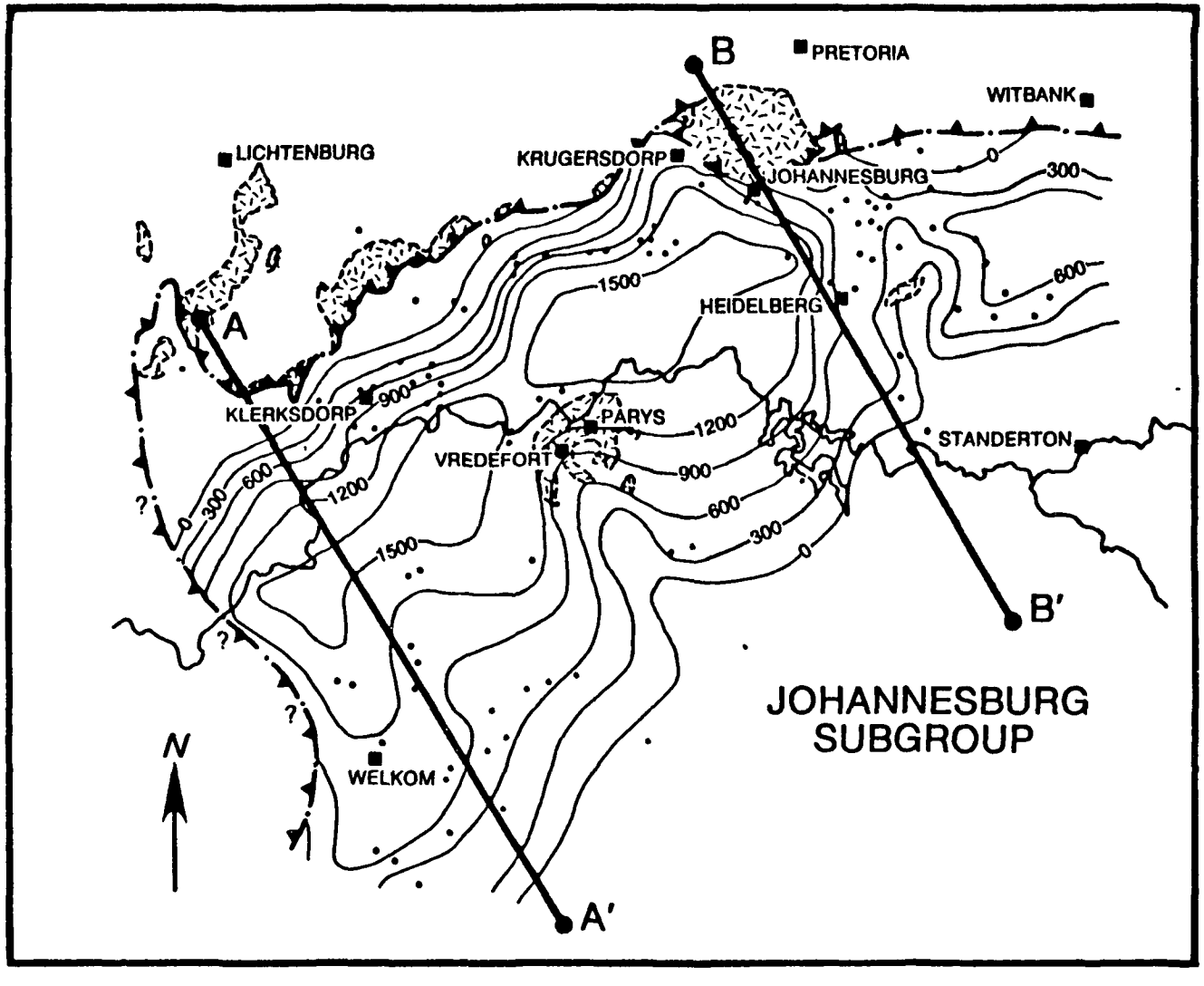

B
PRETORIA
WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
0
300
600
1500
HEIDELBERG
A
KLERKSDORP
900
PARYS
1200
STANDERTON
VREDEFORT
900
600
300
0
1200
600
300
0
1500
?
?
?
B'
JOHANNESBURG
SUBGROUP
N
WELKOM
A'

Fig 11

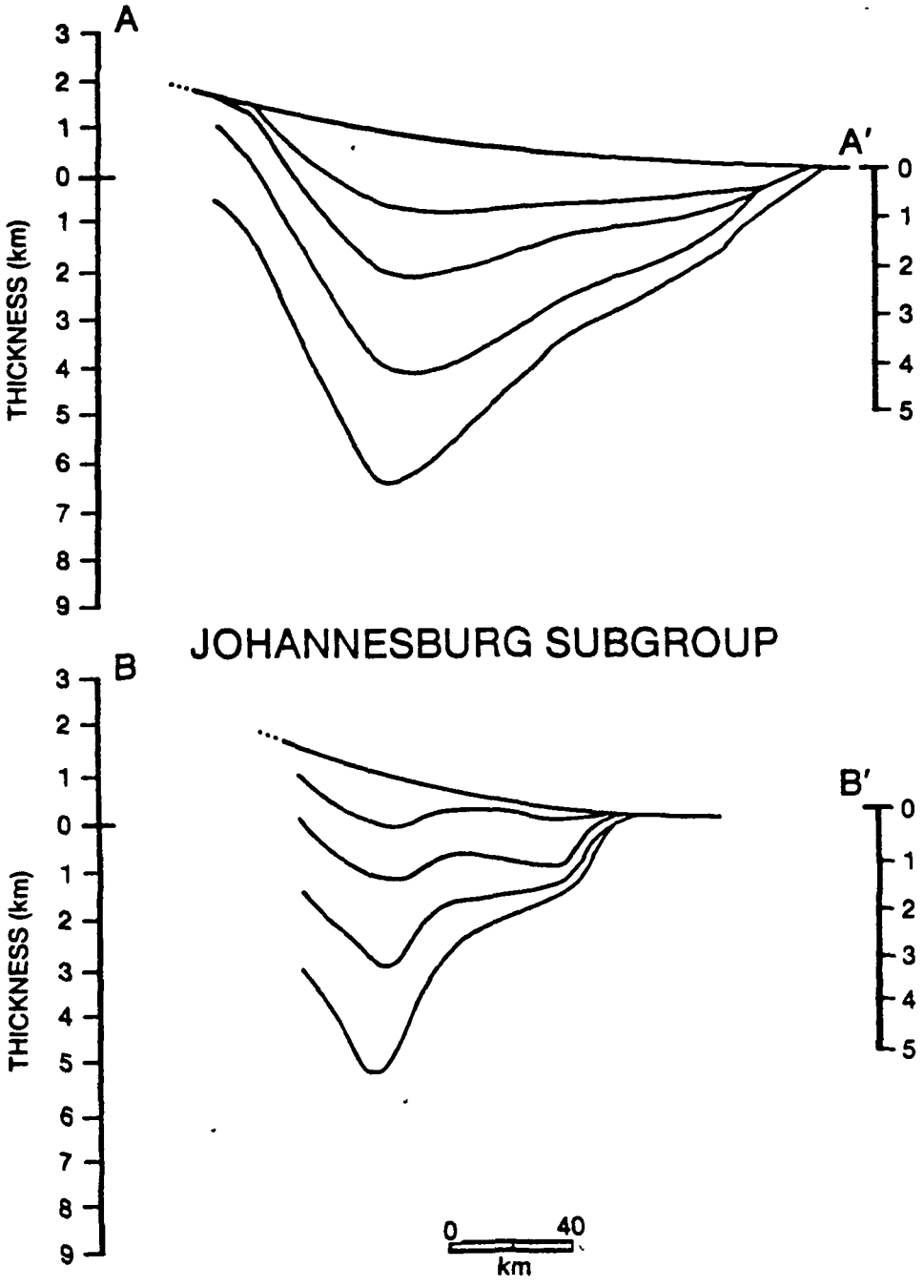


Fig 12

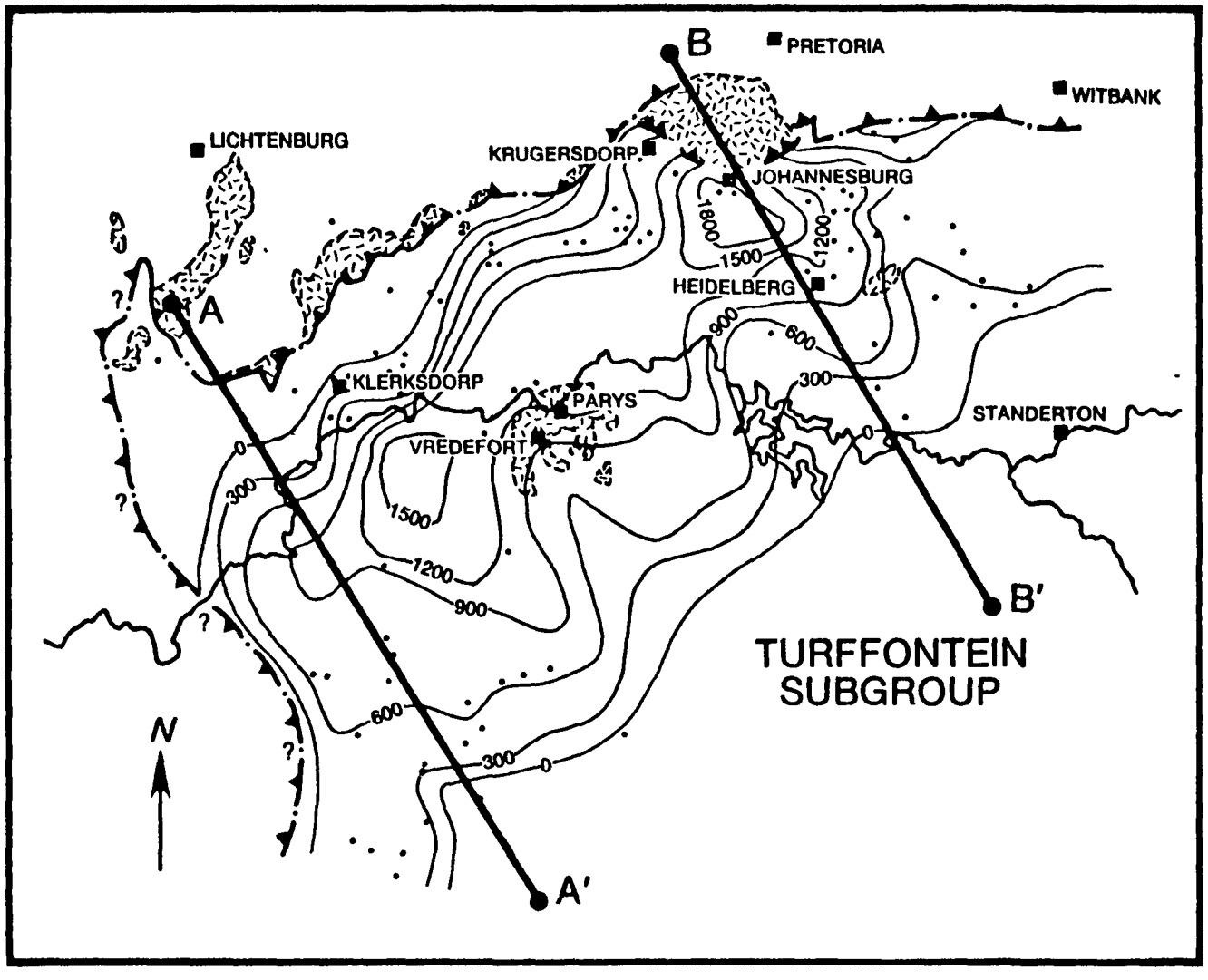

B
PRETORIA
WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
1800
1500
1200
HEIDELBERG
900
600
300
0
A
KLERKSDORP
PARYS
STANDERTON
VREDEFORT
1500
1200
900
600
300
0
B'
TURFFONTEIN
SUBGROUP
N
A'

Fig 13

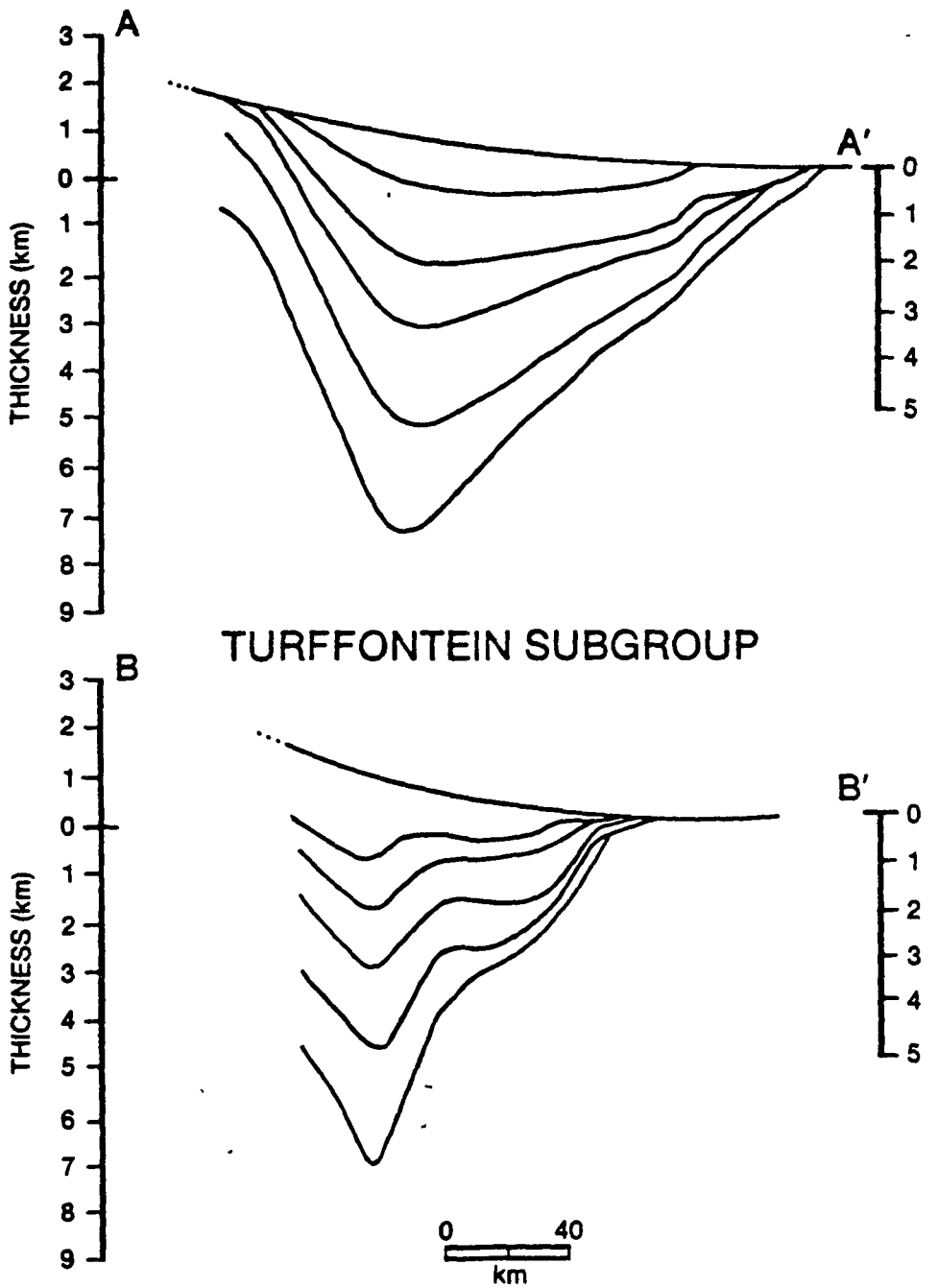

A
A'
B
B'
THICKNESS (km)
3
2
1
0
1
2
3
4
5
6
7
8
9
0
1
2
3
4
5
TURFFONTEIN SUBGROUP
0
40
km

Fig 14

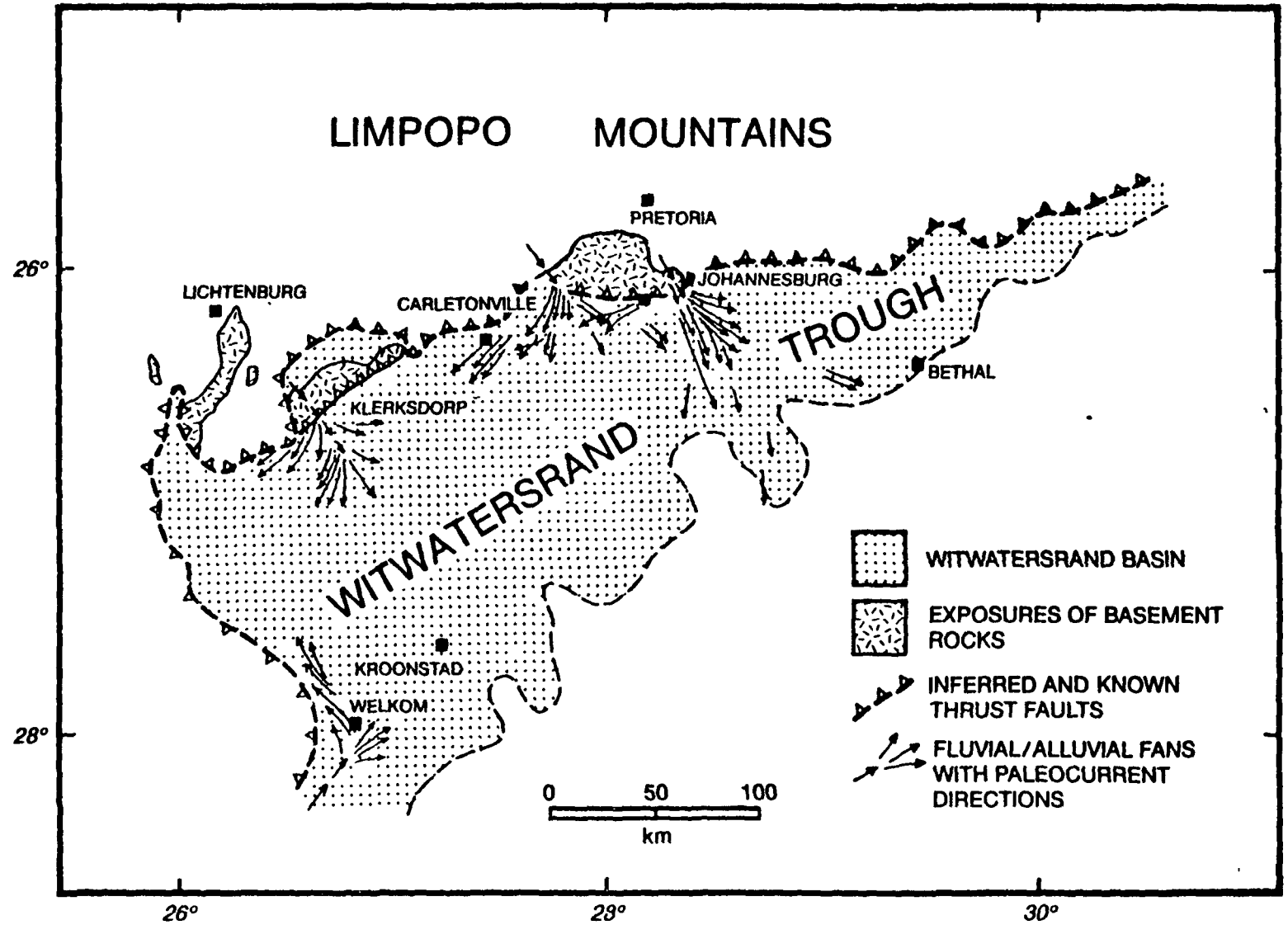

LIMPOPO MOUNTAINS
PRETORIA
JOHANNESBURG
LICHTENBURG
CARLETONVILLE
KLERKSDORP
KROONSTAD
WELKOM
BETHAL
WITWATERSRAND
TROUGH
26°
28°
26°
28°
30°
0
50
100
km
WITWATERSRAND BASIN
EXPOSURES OF BASEMENT ROCKS
INFERRED AND KNOWN THRUST FAULTS
FLUVIAL/ALLUVIAL FANS WITH PALEOCURRENT DIRECTIONS

Fig 15

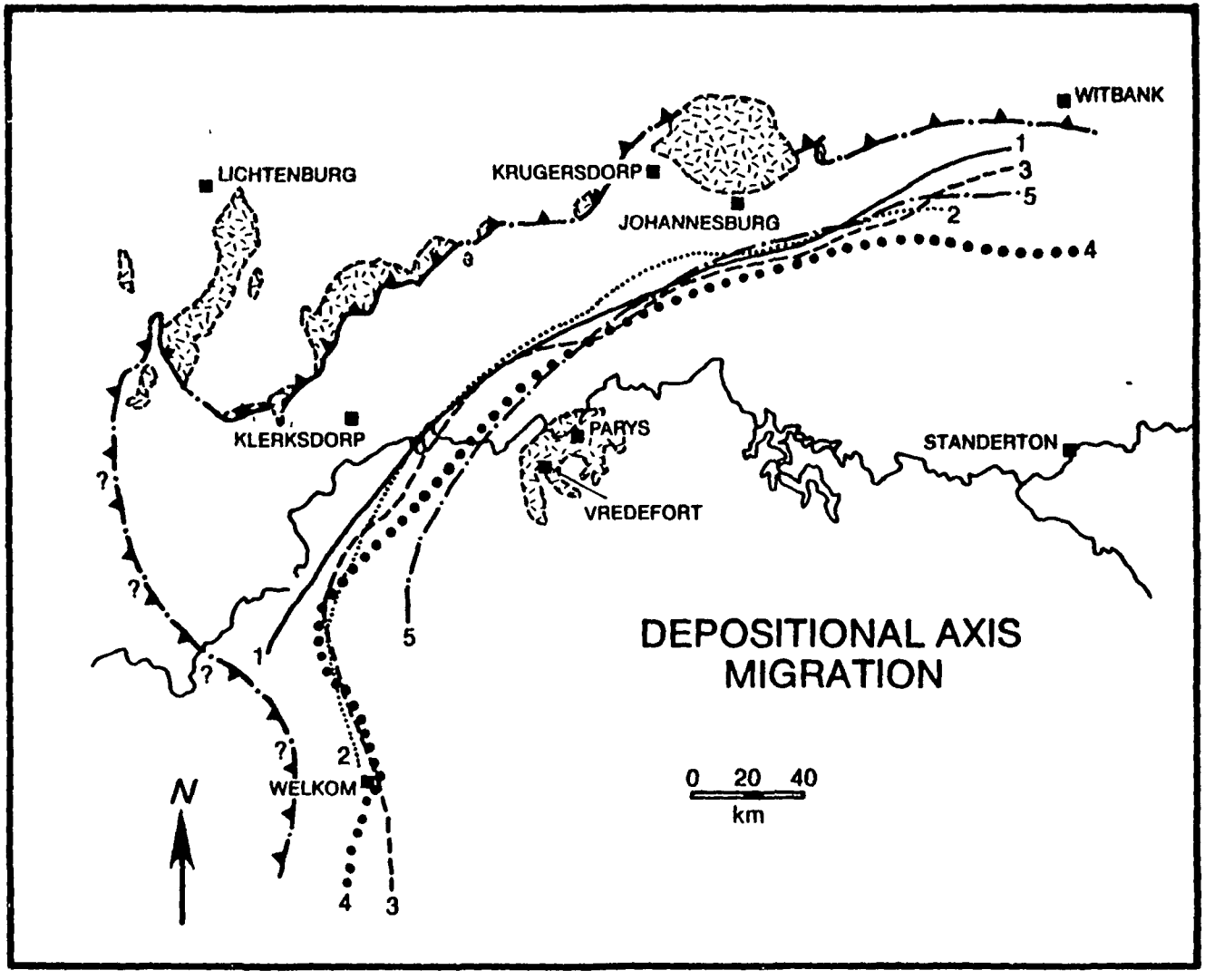
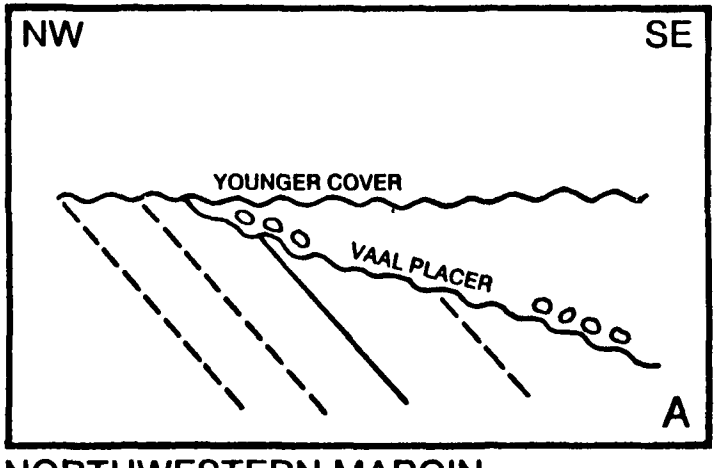
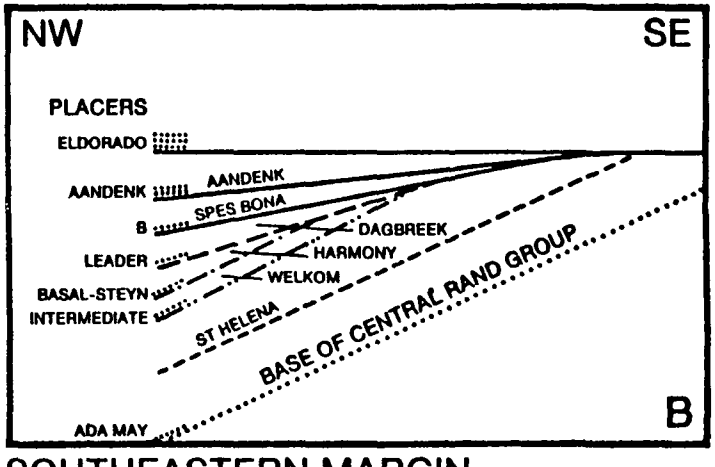

WITBANK
LICHTENBURG
KRUGERSDORP
JOHANNESBURG
KLERKSDORP
PARYS
STANDERTON
VREDEFORT
WELKOM
N
DEPOSITIONAL AXIS
MIGRATION
0 20 40
km
1
2
3
4
5

Fig 16

SCHEMATIC VERTICAL CROSS-SECTIONS



NORTHWESTERN MARGIN



SOUTHEASTERN MARGIN

Fig 17

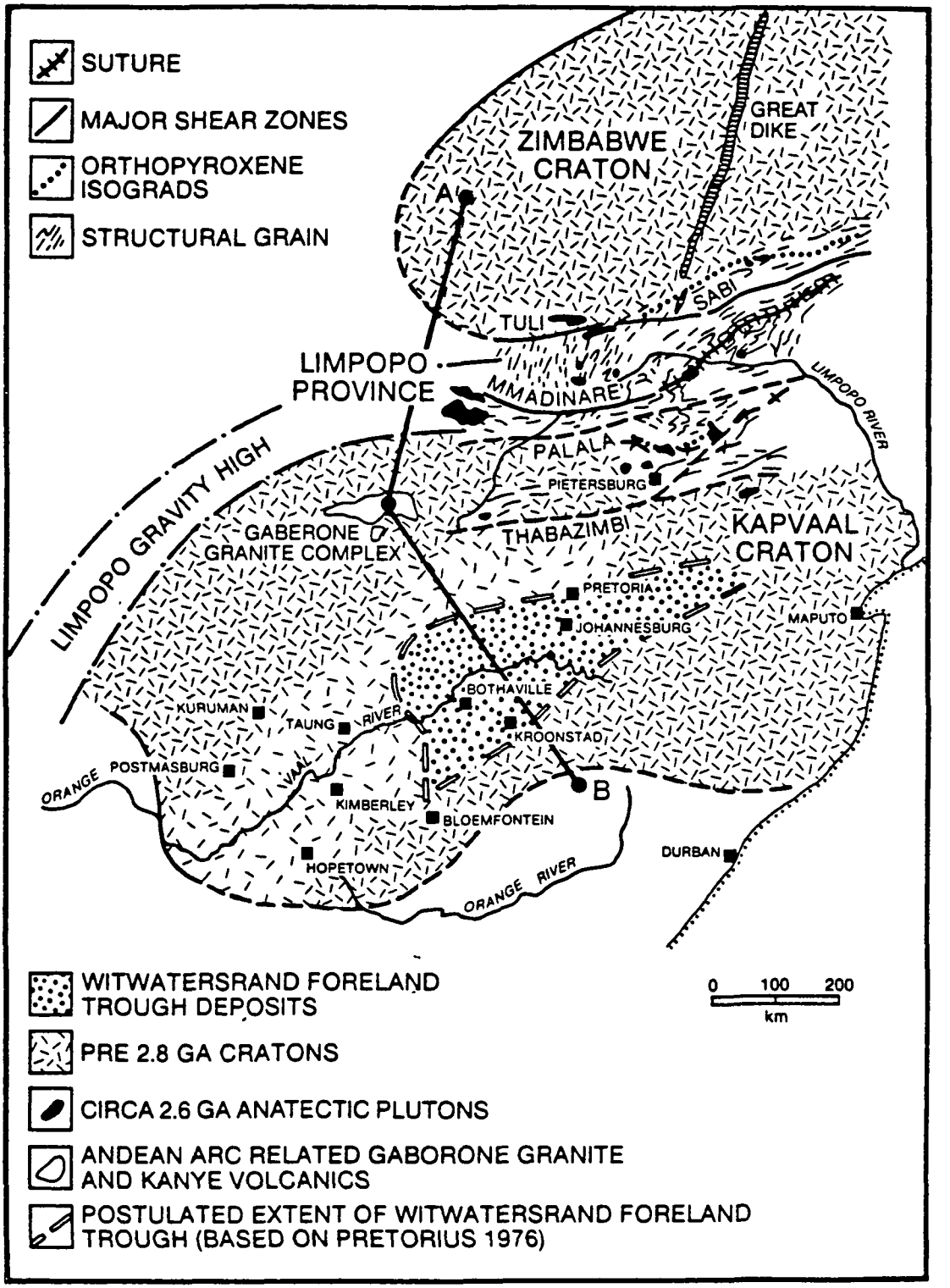

SUTURE
MAJOR SHEAR ZONES
ORTHOPYROXENE ISOGRADS
STRUCTURAL GRAIN
ZIMBABWE CRATON
GREAT DIKE
A
SABI
TULI
LIMPOPO PROVINCE
MMADINARE
LIMPOPO RIVER
PALALA
PIETERSBURG
LIMPOPO GRAVITY HIGH
GABERONE GRANITE COMPLEX
THABAZIMBI
KAPVAAL CRATON
PRETORIA
JOHANNESBURG
MAPUTO
BOTHAVILLE
KURUMAN
TAUNG
RIVER
KROONSTAD
POSTMASBURG
VAAL
ORANGE
KIMBERLEY
B
BLOEMFONTEIN
DURBAN
HOPETOWN
ORANGE RIVER
0 100 200
km
WITWATERSRAND FORELAND TROUGH DEPOSITS
PRE 2.8 GA CRATONS
CIRCA 2.6 GA ANATECTIC PLUTONS
ANDEAN ARC RELATED GABORONE GRANITE AND KANYE VOLCANICS
POSTULATED EXTENT OF WITWATERSRAND FORELAND TROUGH (BASED ON PRETORIUS 1976)

Fig 18

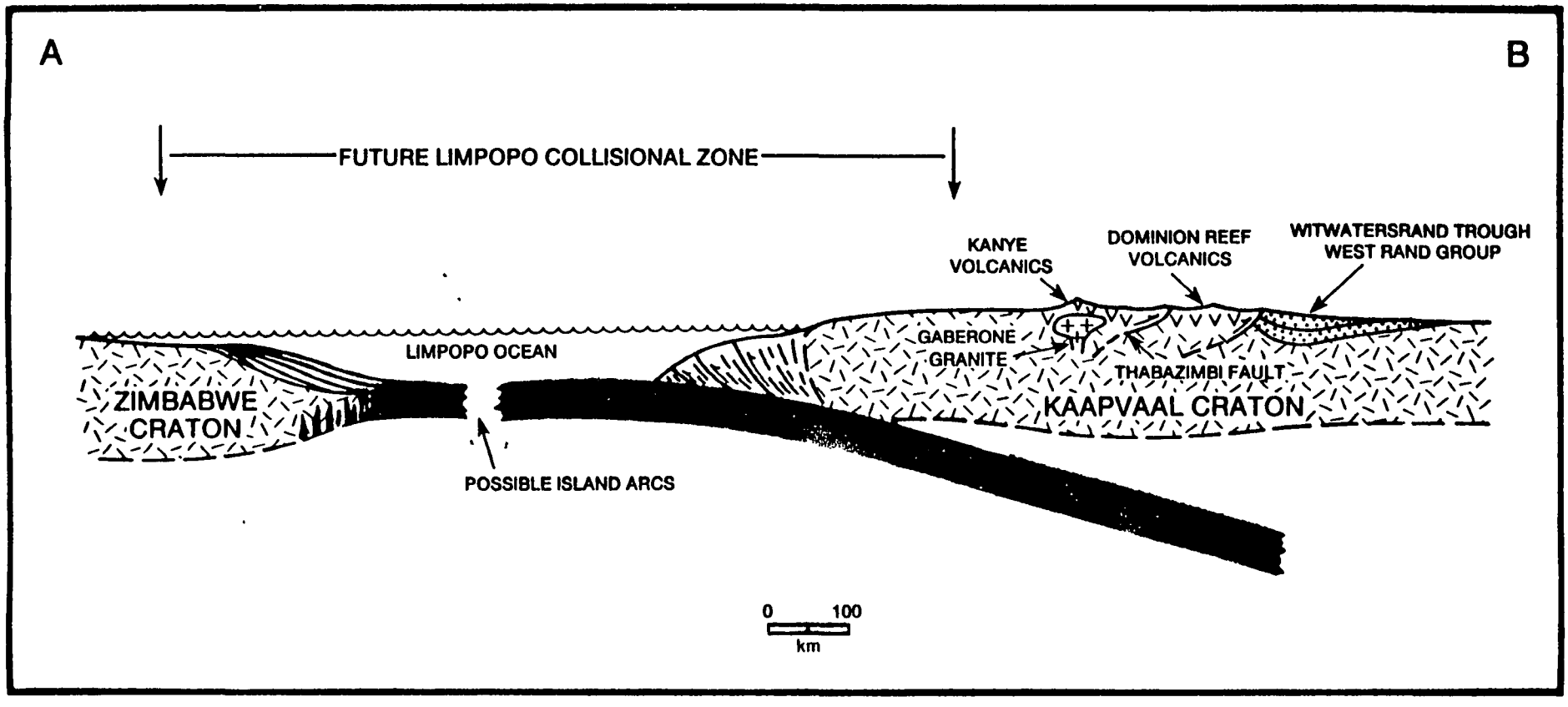

A
B
FUTURE LIMPOPO COLLISIONAL ZONE
KANYE VOLCANICS
DOMINION REEF VOLCANICS
WITWATERSRAND TROUGH
WEST RAND GROUP
LIMPOPO OCEAN
GABERONE GRANITE
THABAZIMBI FAULT
ZIMBABWE CRATON
KAAPVAAL CRATON
POSSIBLE ISLAND ARCS
0
100
km

Fig 19

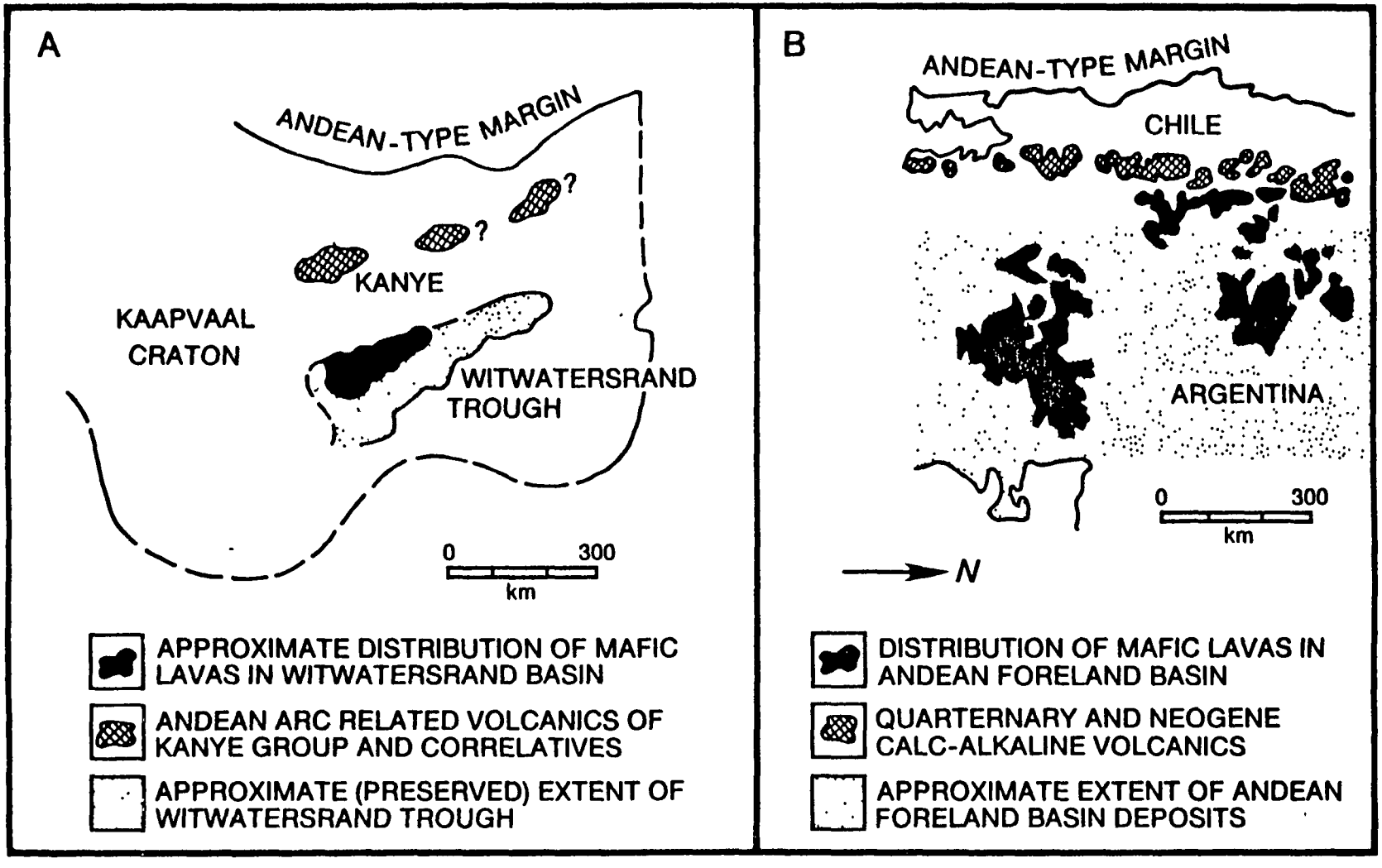

A
ANDEAN-TYPE MARGIN
?
?
KANYE
KAAPVAAL
CRATON
WITWATERSRAND
TROUGH
0
300
km
APPROXIMATE DISTRIBUTION OF MAFIC LAVAS IN WITWATERSRAND BASIN
ANDEAN ARC RELATED VOLCANICS OF KANYE GROUP AND CORRELATIVES
APPROXIMATE (PRESERVED) EXTENT OF WITWATERSRAND TROUGH
B
ANDEAN-TYPE MARGIN
CHILE
ARGENTINA
0
300
km
N
DISTRIBUTION OF MAFIC LAVAS IN ANDEAN FORELAND BASIN
QUARTERNARY AND NEOGENE CALC-ALKALINE VOLCANICS
APPROXIMATE EXTENT OF ANDEAN FORELAND BASIN DEPOSITS

Fig 20

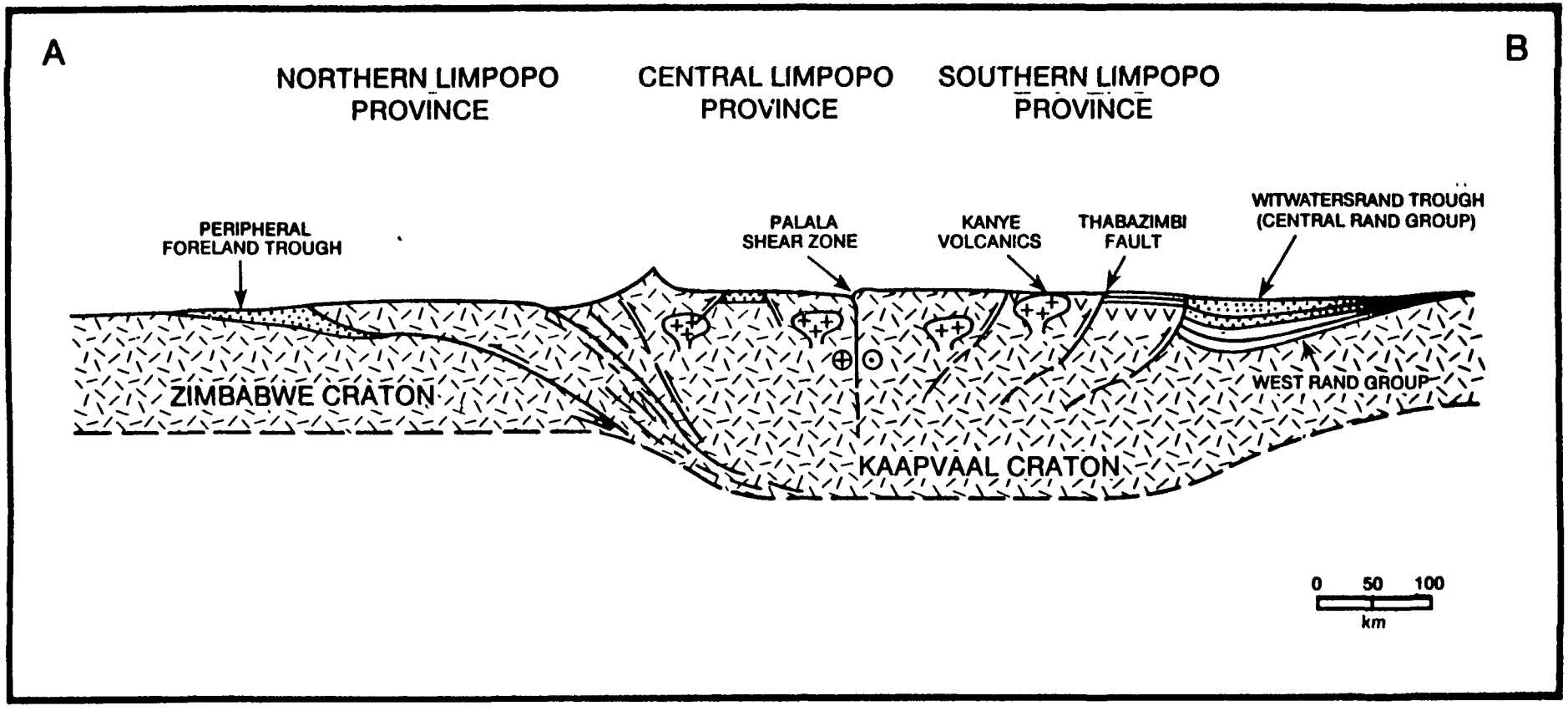

A
B
NORTHERN LIMPOPO PROVINCE
CENTRAL LIMPOPO PROVINCE
SOUTHERN LIMPOPO PROVINCE
PERIPHERAL FORELAND TROUGH
PALALA SHEAR ZONE
KANYE VOLCANICS
THABAZIMBI FAULT
WITWATERSRAND TROUGH (CENTRAL RAND GROUP)
ZIMBABWE CRATON
KAAPVAAL CRATON
WEST RAND GROUP
0 50 100
km

Fig 21