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**GEOLOGY OF NHLANGANO-MAHAMBAMBA AREA, SHISELWENI
REGION, SOUTHWESTERN SWAZILAND**

by

LEO L. KANYA, B.Sc.(Hons).

A thesis submitted to

the Faculty of Graduate Studies and Research

in partial fulfilment of

the requirements for the degree of

Master of Science

Department of Earth Sciences

**Carleton University
Ottawa, Ontario**

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REGION, SOUTHWESTERN SWAZILAND**

submitted by

Leo Linda Kanya, B.Sc.(Hons)

in partial fulfilment of the requirements
for the degree of Master of Science

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**GEOLOGY OF NHLANGANO - MAHAMBA AREA, SHISELWENI
REGION, SOUTHWESTERN SWAZILAND**

ABSTRACT

The Mahamba-Nhlangano area is underlain by deformed and metamorphosed rocks of Early to Late Archean age, Late Archean intrusive rocks, and unmetamorphosed Palaeozoic strata of the Karoo Supergroup.

The main Archean unit, the Pongola Supergroup, comprises the predominantly volcanic Nsuze Group and the predominantly sedimentary Mozaan Group. The Nsuze Group (not exposed within the map area) consists mostly of andesitic lavas which crop out in the western part of the area. The younger Mozaan Group, which is exposed, contains abundant quartz arenites, some of which have been completely recrystallized to quartzites. Minor arkosic, phyllitic and schistose layers are interbedded with the quartzites.

The Archean sedimentary strata are adjacent to the Nhlangano Gneiss, which is tonalitic in composition and has a radiometric age of 2929 Ma. This suggests that it may be basement that was partially reactivated during deposition of the uppermost beds of the Upper Mozaan Group of the Pongola Supergroup as the latter was deposited during the time interval from 3000 to 2900 Ma.

A melanocratic garnetiferous gneiss adjacent to the Nhlangano gneiss,

now named the Mahamba Gneiss, was earlier said to belong to the Bimodal Suite (Ngwane Gneiss); it is very susceptible to weathering, and therefore most of it is covered by overburden. Its U-Pb age (on zircons) ranges from 3297 to 3553 Ma (Kroner & Compston, 1988), suggesting that the Mahamba Gneiss was part of the basement on which the Pongola Supergroup was deposited.

A dark coloured rock of amphibolite grade, that has intruded the Nhlanguano gneiss, forms a border between the Mahamba Gneiss and the younger Karoo sediments.

Younger felsic intrusions of the Hlatikulu Granite have extensively intruded both the Pongola Supergroup and the earlier-formed granite gneiss. These granites are post-tectonic and have an isotropic fabric.

The youngest sedimentary rocks of the area, Permo-Triassic strata of the Karoo Supergroup, comprise units of sandstone, arkose, and carbonaceous shale. The oldest of these units, the Dwyka Formation (tillite), unconformably overlies the Pongola Supergroup.

Two types of dolerite and gabbro dykes are observed in the area: pre-Karoo dykes which trend northerly, parallel to the trends of the major formations; and post-Karoo dolerite dykes, whose trend is variable. Provenance of the Pongola Supergroup sediments has been difficult to establish because of lack of primary structures and heavy minerals. The Pongola sediments were extensively altered during metamorphism

of the Mahamba Gneiss and Nhlangano Gneiss. Metamorphism took place before intrusion of the Hlatikulu Granite, which still retains its isotropic fabric.

The younger sediments of the Karoo Supergroup show a strong resemblance in composition to the Hlatikulu Granite. SEM analytical work on the biotites of rocks of all the different units points to the Hlatikulu Granite as the probable source for most of the Karoo sediments.

Geothermometry of biotites and garnets of the Mahamba Gneiss has led to the conclusion that the temperature of metamorphism was 600 degrees. Petrographic work on the metasediments shows a change in metamorphic grade from low (andalusite) in the southwest to medium (sillimanite) in the northeast

CHAPTER 1

INTRODUCTION

1.1 Purpose of Study

The Shiselweni region in southwestern Swaziland has in the last 20 years been overlooked when projects to map the country were initiated. Thus, most previous work pertaining to the area is now outdated, and there was need to map the area anew. Much progress has been made through recent studies of rocks of the Pongola Supergroup in the Republic of South Africa, but the same cannot be said of the Swaziland segment, and therefore this study of the Pongola was initiated to rectify this disparity.

1.2 Study Area

The study area is just west of Nhlangano Township (fig.1), in the Mahamba area of the Shiselweni region, southwestern Swaziland. This area of approximately 55 km² contains a wide variety of rock units. Two months in 1993 were spent mapping the area. The Mkhondvo river, which runs through the area in a northeasterly direction, has a number of tributary streams, some of which originate in areas underlain by Pongola sediments. The eastern and western parts of the area are at higher altitudes because they are underlain by Pongola quartzites which are resistant to erosion. Between these highlands is a low-lying, gently rolling area underlain by the

Nhlangano and Mahamba Gneiss suites, which have been intruded by younger granites and two generations of dykes, one pre- and the other post-Karoo. Southern parts of the study area towards the western boundary are underlain by the Karoo Supergroup, which contains the youngest consolidated sediments in the area. These crop out in river and stream beds.

Figure 1. Index map showing location of the study area in Nhlangano/Mahamba, Shiselweni region.

1.3 Previous Work

Humphrey (1912) initially included the Pongola strata in the Swaziland Supergroup. He referred to the Nsuze and Mozaan Groups as Lower Pongola and Upper Pongola series respectively, but later (Humphrey, 1931) he reassigned the Pongola to a separate series. Hamilton (1938) mapped the Mozaan of Kubuta, noting similarities to the lower parts of the Witwatersrand Supergroup. Truter (1949) agreed with Hamilton's interpretation and correlated the Nsuze Group with the Dominion Supergroup. Mehliiss (1961) correlated the Mozaan Group with the Witwatersrand Supergroup.

Hunter (1953) mapped both the Kubuta area and unmapped parts of the Mahlangatsha area in. These were re-mapped by Winter (1962). Both Hunter and Winter drew attention to the intrusive nature of the granitoid rocks of these areas, and because of uncertainty that surrounded the correlation of the lithological units, preferred to refer to the Swaziland section of these beds as the Nsuze and Mozaan Series.

Wilson (1980) mapped the area during a programme that covered the entire country. In 1988, Wilson and Jackson discussed the mantled gneiss domes of the area and the concept of 'stable' Pongola cratonic cover, concluding that conditions prevailing during Pongola sedimentation were actually metastable and temporary, being

supplanted by a regime of thermal reactivation which manifested itself in the form of dome-and-basin folding and high-grade metamorphism of the Pongola cover, gneiss-dome formation, and intrusion of a sheet-like granite batholith and sharply discordant granite plutons.

Hegner & Kroner (1987) concluded that the Pongola Supergroup represents 'the oldest cratonic cover sequence.' They noted that the supergroup is intruded by the predominantly gabbroic Usushwana Suite at or near the base of its lower volcanic pile, whereas differentiated mafic to ultramafic sills occur at various levels in the upper sedimentary unit.

Matthews (1987), who studied the Pongola Supergroup of southeastern Transvaal and Natal for many years, has divided it into two structural domains, the Northern and Southern Domains.

(1) The Northern Structural Domain comprises a thick (~13km) conformable assemblage of strata of low to very low metamorphic grade, consisting of a predominantly volcanic lower sequence (Nsuzi Group) and a mainly sedimentary upper sequence (Mozaan Group). Both are asymmetrically distributed in east-west sections across the basin. The structure of this basin is also asymmetric, with a relatively undisturbed western margin, but extensively deformed eastern and southern margins; disrupted segments extend between several large, intrusive post-Pongola granitoid plutons.

(2) The Southern Structural Domain is composed of contrasting major structural units which occur in ascending order from east to west, due to a westward regional structural plunge. The lower unit is an extensive segment of the pre-Pongola granite-greenstone basement, containing two widely spaced (20km) east- trending structural wedges of folded and downfaulted Nsuze strata, although the northernmost wedge also contains an unconformable, folded Mozaan sequence that oversteps Nsuze formations to the south.

CHAPTER 2

2.1 GENERAL GEOLOGY OF THE STUDY AREA

The Nhlanguano-Mahamba area contains a variety of igneous lithological units (fig. 2, and pocket at back) which are difficult to distinguish in the field, although they can be clearly distinguished on the basis of petrographic analysis. Along the eastern boundary of the area, the upper part of the Pongola Supergroup, the Mozaan Group, is made up of conglomerates and sillimanite-bearing metaquartzites. Wilson and Jackson (1988) named these rocks the Mkhondo Valley Metamorphic Suite. Within the Mozaan beds are conformable lenses of pelitic schists and banded iron-formations, suggesting that intervals of low-energy sedimentation interrupted high-energy conditions that prevailed during deposition of the arenites and conglomerates.

Immediately west of the Mozaan Group, three different rock types can be identified: (1) the quartzofeldspathic Nhlanguano Gneiss, which is the predominant rock in the area; (2) a sill-like unit of amphibolite, which probably intruded the gneisses; (3) a younger granitoid rock, the Hlatikulu Granite, that has intruded all other major units, and appears to underlie most parts of the area.

Two types of diabase dyke have been identified: a northerly-trending pre-Karoo diabase/gabbro dyke in the central part of the map area, and throughout the area, numerous post-Karoo dolerite dykes that do no obvious preferred trend. In addition, on the north bank of the Mkhondo River, adjacent to the bridge, a single outcrop of diorite is

exposed.

Toward the west, adjacent to the Nhlanguano Gneiss, is the oldest rock of the area, the Mahamba Gneiss, a melanocratic garnetiferous gneiss which, because it is very susceptible to weathering, is extensively covered with overburden. This unit, distinguished by its dark grey and white bands, is composed of quartz, feldspar, biotite and garnet. Some samples contain minor amounts of sericite, chlorite, amphibole, zoisite, muscovite and epidote.

West of the Mahamba Gneiss is the youngest sedimentary unit in the area, the Karoo Supergroup, which crops out at the Mkhondo River. In other areas it is covered with overburden. The base of the Karoo is the glaciogenic Dwyka Tillite, composed mainly of matrix-supported conglomerates. The Dwyka, deposited during the early Permian, is conformably overlain by the Eccia Group, which was deposited during Late Permian to Early Triassic time. It is made up of feldspathic sandstones, arkosic wackes, and carbonaceous shales. A borehole was drilled near the western edge of the Karoo Supergroup.

The western part of the study area comprises the Mozaan Group quartzites, which trend roughly N-S and dip $\sim 40^\circ$ to the east. These quartzites are less deformed than those of the Mkhondo Valley Metamorphic Suite, and the widespread occurrence of andalusite suggests that these rocks formed as a result of low-pressure, low-temperature metamorphism. Beds of banded iron-formation appear to be conformably interlayered with the Mozaan quartzites.

Generalized Stratigraphic Column of Mahamba - Nhlangano
and adjacent areas

Figure 2. Generalized stratigraphic column of Mahamba - Nhlangano area, with ages of the different units

2.2 MAHAMBBA GNEISS

The outcrop of the Mahamba Gneiss forms a peripheral arc along the western margin of the Nhlangano Gneiss, which occurs as a central domal structure. Samples of the Mahamba Gneiss are fine- to coarse-grained; they contain garnet, but no amphibole, except in samples from two localities (Appendix 1, p.98 close to the contact with amphibolite gneiss. The main minerals of the Mahamba Gneiss are quartz, plagioclase, biotite, garnet, and chlorite, with or without muscovite. K-feldspar and sericite are minor constituents. The foliation ranges from distinctly tabular (fig.3) to lenticular (Fig.4). Garnets tend to be abundant in felsic clots (Fig. 5).

The Mahamba Gneiss has a zircon Pb-Pb evaporation age ranging from 3297Ma to 3553Ma +/- 6 (Kroner & Compston, 1988). The Mahamba Gneiss, therefore, is not coeval with part of the Pongola Supergroup as suggested by Wilson and Jackson (1988). Instead, it represents the basement on which all other units accumulated (Maphalala & Kroner, 1993).

Rocks of the Mahamba Gneiss Suite appear to represent the late stages of formation of the Ngwane Gneiss (formally called the Bimodal Suite), which is part of the Ancient Gneiss Complex. A traverse through the area, where the Mahamba Gneiss crops out, revealed that the contact with the Nhlangano Gneiss is gradational.

Figure 3. A typical example of the Mahamba Gneiss suite, showing distinct gneissic layering (Diameter of coin is 2 cm).

Figure 4. A photograph of the Mahamba Gneiss showing the banded fabric and numerous garnet porphyroblasts (Diameter of coin is 2 cm).

3

4

The Mahamba Gneiss has a distinctive grey and white banding. In the Mkhondo River, it forms a flat pavement along the river course. The gneiss contains two obvious generations of intrusions: the first is an amphibole-rich veining that cuts through the original gneissic fabric; the second is a felsic migmatite component that cuts across these veins. All samples belonging to the Mahamba Gneiss suite virtually have the same mineralogy, with only minor differences here and there. All samples contain quartz, plagioclase, orthoclase, microcline, chlorite, biotite, garnet and opaques. Only two contain amphibole (samples 35 and 40). In addition, some samples contain muscovite, titanite, tourmaline and zircon. The samples are generally fine- to medium- grained, but one (sample 35) is coarse- grained, and contains zoned microcline. All of the minerals are subhedral- to anhedral. An amphibolitic gneiss with quartz lenses crops out on the Vonde River. The surrounding higher area is covered with overburden. The amphibolite contains ptigmatic folds and displays low-angle thrusts (fig. 6). The folds are concurrent with the first phase of deformation, but some of the thrusts cut through the foliation fabric, thus making the faults second generation. Some quartz lenses are boudinaged, and commonly show a sinistral direction of displacement. Sample 24 (from the amphibolite) shows distinct mineral alignment. Ilmenite and magnetite are the metallic minerals in the rock. Samples 35 and 50, from the same locality, display a distinct layered fabric. In sample 50, some of the biotite is green and some is brown, probably reflecting differences in the relative

abundance of Ti, Fe^{2+} and Fe^{3+} . The garnet is rounded and there is a distinct layered fabric due to the alignment and segregation of minerals.

Figure 5. Typical Ma'namba Gneiss showing abundant garnets in felsic clots. Coin is 2cm in diameter.

Figure 6. A photograph of an amphibolitic gneiss from the Vonde River, showing ptygmatic folds and low-angle thrusts (Diameter of coin is 2 cm).

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2.3 NHLANGANO GNEISS

The Nhlangano Gneiss, whose Pb-Pb zircon evaporation age is 2929 $\pm$ 5 Ma (Maphalala and Kroner, 1993), comprises a sub-circular dome, which is centred on the township of Nhlangano, and is bordered to the west and south by the Mahamba Gneiss, Pongola and Karoo strata, and to the north and east by the Hlatikulu Granite. The dome is a periclinal anticlinorium containing three gneiss-cored antiforms separated by two narrow synforms in which metasedimentary cover is preserved.

The core of the Nhlangano Gneiss dome comprises a biotite-microcline-quartz granite gneiss. Compositional variations are subtle, but the appearance of the rock varies widely in correspondence to changes in the tectonite fabric. In the centre of the dome, the rock is structurally isotropic and has a hypidiomorphic texture that is typical of granite. The intensity of fabric development increases outwards, although this trend is not consistent because of second-order folds and shear zones within the dome (fig.7).

Medium- to coarse- grained quartzofeldspathic rocks of the Nhlangano Gneiss, which crop out throughout almost the entire field area, are easily recognizable because of their pinkish-red weathered colour. The gneiss, which is locally migmatitic, can also be distinguished from other gneisses by its high content of quartz and feldspars.

Figure 7. An example of the Nhlanguano Granite Gneiss, showing folded foliation near the contact with the Mahamba Gneiss (Coin is 2 cm in diameter).

Figure 8. A photograph of the Nhlanguano Gneiss illustrating its banded fabric, with quartz porphyroblasts (diameter of coin is 2 cm).

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The Nhlangano Gneiss is distinguished from other gneisses by migmatitic fabric and a high content of quartz and feldspars, together with varying amounts of amphibole. Other common minerals in this gneiss are sericite, chlorite, biotite and minor titanite, tourmaline, zircon, and opaques (Appendix 1, p.99). Calcite in one sample appears in a late-phase vein (fig.9). A representative sample of the Nhlangano Gneiss (sample 20), is a medium-grained rock with a faint foliation, and shows both perthitic and antiperthitic textures. A sample taken from an outcrop adjacent to a post-Karoo diabase (sample 39) is fine-grained, anhedral, and almost equigranular, but displays only rare perthitic textures.

Sample 43 is from another exposure of the Nhlangano Gneiss that crops out within a cluster of roughly north-trending pre-Karoo dolerite dykes, a few hundred metres south of the locality of sample 39. This medium- to coarse-grained sample shows anhedral to subhedral textures, and contains both perthite and antiperthite. The microcline is zoned, and the quartz shows undulose extinction. Banding is clearly visible.

Sample 48, next to the road from Mahamba to Nhlangano, contains lenses of quartz and calcite, but without any cross-cutting features. It is fine to very coarse-grained, and most of the minerals are anhedral. The sample shows a distinctly layered fabric which is accentuated by alignment and layer-segregation of biotite (fig. 9). Perthitic and granophyric textures occur locally.

Sample 57, collected south of the study area, also shows perthitic texture.

Figure 9. Typical Nhlanguano Gneiss with a late phase calcite vein cutting through the migmatitic fabric. (Diameter of coin is 2 cm).

Figure 10. A photograph of a porphyritic Hlatikulu Granite containing phenocrysts of microcline. (Diameter of coin is 2 cm).

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2.4 HLATIKULU GRANITES

Younger granitic intrusions, with no signs of deformation, crop out at various localities throughout the study area. These are late-stage Archean granite intrusions, belonging to the 2722 ± 7 Ma Hlatikulu Batholith (Maphalala & Kroner, 1993, Pb-Pb, zircon evaporation). The Hlatikulu Granite was earlier dated by Barton et.al. (1983), to be 2417 ± 148 Ma (Rb-Sr, whole rock), the age being due to significant scatter in analytical data, reflecting an open system behaviour.

Most outcrops of the Hlatikulu Granite are fresh, medium- to coarse-grained and porphyritic, with abundant phenocrysts of euhedral potassium feldspar (fig. 10). This unit has intruded quartzite and banded iron-formation of the Mkhondo Valley Metamorphic suite of the Mozaan Group. One outcrop is near the contact of the Hlatikulu Granite and Nhlanguano Gneiss, but the contact cannot be clearly defined. A sample from this outcrop contains anhedral quartz, and subhedral to euhedral feldspars. Perthite and antiperthite textures are common in the feldspars. Not all outcrops of the Hlatikulu Granite are porphyritic; for instance, a fine-grained dyke near and almost parallel to the Mkhondo River does not have any phenocrysts. This dyke, however, contains a xenolith of the Mahamba gneiss. The mineral compositions of the different outcrops emphasize the notion that all were derived from the same protolith, or from protoliths of the same composition. One outcrop, however, contains abundant epidote which is interpreted to have formed as a result of hydrothermal alteration.

A spherulitic felsite dyke (fig.11) is inferred to belong to a swarm of granite dykes, including the one described above. The spherulitic dyke, which intrudes an earlier diabase, is very fine-grained and contains quartz, orthoclase, opaques, biotite and the sodic amphibole, arfvedsonite. It contains abundant spherulites of quartz and feldspar up to 5 mm in diameter.

The Hlatikulu Granite is typically grey, with some buff-coloured local variations. It is medium to coarse grained, and is locally porphyritic. The dyke described above that is intrusive into the Mahamba Gneiss is medium- to fine-grained. The granite of the study area is not banded, and where apophyses of the granite embay the sediments, they do not have a pegmatitic character, as is the case in other areas. The medium-grained granite has a spotted appearance because of the light coloured minerals that tend to collect in clusters, with the mafic minerals arranged between these clusters. One granite near the Mozaan contact contains small crystals of magnetite which are surrounded by felsic rims. The magnetite grains are ~3mm in diameter. The Hlatikulu Granite contains quartz, microcline-microperthite, plagioclase and biotite (Appendix 1, p.99). Orthoclase is subordinate to microcline, which occurs as irregular grains of variable size, depending on the texture of the granite in which it occurs. The microcline is fresh, with only traces of alteration. It has embayed and replaced earlier-formed plagioclase; relics of plagioclase are common in the larger microcline grains. Delicate fibre perthite and coarser vein perthite are common in most grains of microcline.

Plagioclase forms between 5 and 10% of the modal mineral compositions in the granite. In the fine-grained samples the grains are euhedral and similar in size to the rest of the feldspars and quartz. In the coarser granites the plagioclase has irregular outlines, commonly showing extensive replacement by potash-feldspar. The plagioclase of medium-grained rocks is usually fresh, but where the grains have been corroded and mantled by potash-feldspar, the plagioclase is commonly turbid. In many cases the altered cores are surrounded by clear rims. The plagioclase is normally oligoclase, but some andesine is present. Most of the plagioclase is zoned, with a progressive increase in the albite component towards the margins. All grains of plagioclase display multiple twinning, with the albite type predominant, generally in combination with Carlsbad twinning. Biotite is the most abundant of the mafic minerals, and occurs as ragged flakes. In places it shows abundant pleochroic haloes surrounding fine-grained zircon and/or allanite inclusions. Muscovite and amphibole are absent from the Hlatikulu Granite; accessory minerals (apatite, zircon, allanite, epidote and magnetite) are generally more plentiful in the darker granites. Allanite is common in granites that have intruded the Mozaan sediments. All samples from the Hlatikulu Granite are very similar, except for grain size (which varies from fine to coarse) and the degree of sericitization. Orthoclase content ranges from 5 to 10%, microcline from 10 to 20%, and plagioclase from 5 to 10%. All samples contain biotite and chlorite, but one (34) contains 5% epidote, which may be a result of hydrothermal alteration.

Figure 11. Spherulitic felsite dyke belonging to the Hlatikulu Granite. The spherulites are made up of quartz and feldspar.(Diameter of coin is 2 cm).

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The minerals range in grain size from fine- to very coarse- grained. Sample 13 in hand specimen is medium- to coarse- grained and porphyritic, with phenocrysts of euhedral potassium feldspar (fig. 12). It intrudes the Mozaan Quartzite and associated BIF, and contains perthitic textures in some of the microcline and orthoclase.

Sample 22 is almost in contact with the Nhlanguano Gneiss, but, as can be seen in the Fig.13, the contact cannot be clearly defined. Perthite and exsolution textures are common in the feldspars.

Sample 33, collected 200m south of Mkhondo River, is from a granite dyke that belongs to a swarm of dykes that intrude the Nhlanguano gneiss and dolerite. Saussuritization and sericitization are at an advanced stage. Perthitic and antiperthitic textures are common.

Sample 34 is from the same vicinity, except that it is adjacent to a pelitic schist/quartzite outcrop. Epidote probably formed as a result of hydrothermal alteration.

A granite dyke (sample 36) near, and almost parallel to the Mkhondo River, contains a xenolith of the Mahamba Gneiss, and is made up of quartz, microcline, plagioclase, chlorite, sericite and opaques. It is medium- to coarse- grained, and the grains are mostly anhedral (fig.14).

Figure 12. Photomicrograph of a porphyritic Hlatikulu Granite showing phenocrysts of microcline (crossed polars). (Field of view is 4 mm)

Figure 13. Contact of the Hlatikulu Granite and Nhlengano Gneiss. (Lens cap on foreground is 5 cm in diameter).

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Sample 32, which intrudes the gabbro (Sample 31), is composed of quartz, orthoclase, opaques, biotite, and the sodic amphibole, arfvedsonite (fig.15).

Figure 14. Hlatikulu Granite near the Mkhondo River containing a xenolith of Mahamba Gneiss.(Diameter of coin is 2 cm).

Figure 15. Photomicrograph of spherulitic felsite that intrudes gabbro. The spherulites are made up of quartz and contain minor sodic amphibole (crossed polars). Field of view is 4 mm.

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2.5 SEDIMENTS AND METASEDIMENTS OF THE PONGOLA SUPERGROUP

The Pongola Supergroup is a supracrustal succession that extends north from the Republic of South Africa into Swaziland as two synclinoria. It is made up of lower Nsuzi and upper Mozaan Groups, and is intruded by the Usushwana intrusive suite in the west. The Nsuzi Group in Swaziland rests on a paleoregolith that was produced by weathering of basement rocks during the Archean (Wilson and Jackson, 1988). There is no sharp contact between the Nsuzi and the Mozaan. Only the Mozaan crops out in the study area (fig.16).

The Mozaan Group is mainly sedimentary, as opposed to the volcanoclastic Nsuzi Group. Because most of the sediments have undergone high-grade metamorphism, Wilson and Jackson (1988) assigned them the name Mkhondo Valley Metamorphic Suite. These metasediments are exposed in the eastern part of the map area, and have a southeasterly trend. The main component of the sediments is quartzite, interrupted by minor lenses of pelitic schist and banded iron-formation. These sediments are also intruded by the Hlatikulu Granite.

Figure 16. The distribution of the Pongola in Swaziland and South Africa (after Matthews 1967).

THE NSUZE GROUP: Mehlis (1947), who named this the Lower Pongola series, observed that it underlies the Mozaan Group (Upper Pongola Series) with a marked angular unconformity. Because Nsuze rocks are not exposed in the study area, this unconformity was not observed.

The Nsuze group consists predominantly of volcanic flows with patches of quartzite and phyllite (Mehlis, 1947). Its maximum thickness (3000m), is reached in the southwestern part of the basin in the Natal area, and thins progressively towards the north. The Nsuze crops out in two different localities in Swaziland. The larger area of outcrop enters from the Transvaal north of Mahamba, passes east of Sicunusa, and terminates near the Ngwempisi River. The second area lies to the east of Hluti and extends northward to the Sigwe Hills near Maloma, interposed between the granite and the base of the Karoo Supergroup.

Volcanic rocks of the Sicunusa-Mahamba area are predominantly andesites, mostly fine grained and light grey, bluish grey and greenish grey. Rare amygdules, up to 1cm in diameter, contain calcite, quartz, zoisite, epidote, and chlorite. Felsic volcanic rocks occur immediately east of Sicunusa, and extend from beneath the overlapping Mozaan quartzites in the Mozaan valley and pinch out near the Ndlotane River. They are dark and very fine grained, with a bluish black colour. The andesites have undergone considerable alteration.

The plagioclase, which was apparently the main constituent, is very

turbid due to saussuritization; the little that remains appears to have a composition in the range of oligoclase to andesine. The mafic minerals are green hornblende in ragged laths; chlorite, epidote, zoisite and clinozoisite are common products of alteration of the ferromagnesian minerals.

Felsic flows display a uniformly fine-grained groundmass of quartz and feldspar, with minor amounts of biotite and chlorite. The feldspar is commonly saussuritized, making identification difficult, but both albite and orthoclase appear to be present. Oxides and secondary leucoxene are common accessory minerals.

Associated sedimentary units are rare, and the only one of importance is a phyllitic layer that overlies the felsic flows. It extends from the Nhlangano-Sicunusa road and reaches a maximum thickness of 90m before thinning and disappearing beneath the overlap by the Mozaan Group. The phyllites are generally brownish yellow, finely laminated rocks which break into characteristic blocks due to well-developed cleavages. Slightly ferruginous types give rise to a wide range of colours, depending on the amount of iron in the phyllite. The phyllites are uniformly fine grained, and rarely show wide variations in grain size, although quartzites and quartz schist occur along the strike of the phyllite. The phyllite has been deeply weathered at one place where it is siliceous, producing a deposit of quartzose sand. In zones of shearing, augen of diaspore and andalusite up to 60 cm in diameter

are sheathed by phyllitic grey schists that consist of quartz, pyrophyllite, and sericite. A spotted grey phyllite contains very fine-grained aggregates of mica, probably pseudomorphous after andalusite. It appears that the metamorphism was mainly dynamic, because thermal metamorphism would have resulted in a greater interaction between the diaspore and quartz, which occur in close proximity.

The Nsuzi Group thins northwards, finally terminating near the Ngwempisi River. The base of the group is intruded by aplite dykes, several of which may be seen near the Nhlanguano-Sicurus road. The basal contact, coupled with overstep of the Mozaan Group, restricts the thickness of the Nsuzi to a maximum of about 300m. Attitudes could be determined only in phyllites of the Nsuzi Group. The phyllites dip gently eastward at between 30° and 50°.

THE MOZAAN GROUP

Alternating beds of quartzites and shales of the Mozaan Group overlie the Nsuzi volcanics. At Mahamba, an outcrop of quartzites and shales gradually broadens northwards and then northeastwards to the Ngwempisi River. The second occurrence of Mozaan rocks is east of Hlatikulu, and extends eastwards as far as Maloma. Pongola rocks are not exposed north of the Usuthu River.

The Mozaan Group forms rugged topography, with quartzite ridges

standing out from the shales. The basal quartzite at Sithobela in Central Swaziland forms a prominent scarp, succeeded southwards by a repetition of scarps and dip-slopes. The hills overlooking the Ngwavuma valley east of Hlatikulu are distinctly stepped, again due to the presence of interbedded hard quartzites and softer shales. East and northeast of Sicunusa, the Mozaan sediments are predominantly quartzitic, producing a plateau-like topography. North of Mahamba, the shales are thin, and generally fail to influence the landscape.

The two main areas of Mozaan Group sediments represent an overlap of sediments towards the east, with the succession in the Mahamba-Mahlangatsha area representing the lowermost half of the Group. The top quartzite of this area forms the base of the Maloma succession.

The conglomerates in the Mozaan Group are best developed near Kubuta, and to a lesser extent north of Mahamba. The mean pebble diameter is ~ 3 cm, with some clasts as large as 7 cm. They are closely packed, with the interstices filled by a fine quartzitic matrix containing pyrite and sericite. The pebbles are almost entirely vein quartz, but a few are dark, vitreous and chert-like. The pebbles are rounded to oval and are poorly sorted.

The quartzites throughout the succession are light coloured and give rise to prominent outcrops. The grain size of the quartzites ranges from fine-grained to gritty, with the latter type grading into

conglomerates. Impurities are rare, except in the gritty beds where cubes of pyrite give rise to a reddish discoloration due to oxidation. The quartzites are well jointed. Cross-bedding, although characteristic of the quartzites south of Swaziland, was observed in only one outcrop in the study area. The grit beds, which are more abundant in the lower quartzites of the Mahamba and Kubuta area, contain abundant subangular grains. Towards the intrusive granite, the quartzites show evidence of progressive recrystallization, and have locally developed a slight greenish tinge due to the presence of sericite.

Knobby-weathered schistose layers occur within the quartzites, 5 km north of Nhlangano (fig.17). One such layer, approximately 3 m thick, occurs towards the base of the quartzite. Reflecting an increase in silica content, the schist grades along strike from a greenish or purplish tinted rock with a distinctive flaky appearance to a rock closely resembling a sheared quartzite. The flaky appearance results from an abundance of sericite. These schists attracted attention during radiometric traverses, because they gave readings of the order of four times background. The actual radioactive zones in the schist are normally very narrow and range from 3 to 10 cm. They are distinctive on account of their darker colour and the presence of red-brown streaks that follow cross-bedding. Small-pebble conglomerates up to 10 cm thick occur in association with these radioactive zones. The zones merely differ from the normal schistose sequence by virtue of a greater abundance of heavy minerals, among which are monazite,

xenotime, zircon, rutile, and chromite-magnetite.

Above a higher quartzite unit, pelitic beds predominate. In the lower half of the Mozaan Group, such beds are generally thin. The basal pelitic units are represented by grey, phyllitic schist containing knots of andalusite. Gritty and conglomeratic lenses are common. Yet another higher pelitic layer is the first major phyllitic unit, ranging in thickness from 45m to 90m.

Andalusite schists are prominent north of the study area near Gege (Fig. 1), but grade northwards into sandy phyllitic schists with abundant sericite. Above this level, the pelitic layers are uniformly represented by pinkish weathering sandy phyllites in which more arenaceous layers and lenses occur. Grey meta-mudstones are present in the upper pelitic layer; pink ferruginous layers are common in the middle of the group.

The Mozaan Group, which is the one that crops out in the study area, forms the eastern and western flanks of high ground in the field area. Primary structures of these rocks have been almost everywhere completely obliterated. The main minerals of the rocks of the Mozaan Group are quartz, muscovite and opaques (Appendix 1, p. 101). The eastern flank contains sillimanite, and the western flank, andalusite. Heavy minerals like zircon and tourmaline are rare. Grain size ranges from medium to coarse, and the grains show evidence of remobilization.

**Figure 17. Photograph of pelitic schist (Station 18) of the Mozaan Group.
View looking north.**

Figure 18. A coarse-grained quartz arenite from the Mozaan Group (station 52). This is the only outcrop that shows distinct cross-bedding. Younging in this arenite is to the west. Diameter of coin is 2 cm.

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Representative samples of the Mozaan Group, (eg. sample 11) are equigranular, coarse-grained recrystallized quartzites consisting almost entirely of angular quartz grains. In one locality, distinct cross-bedding (fig. 18), shows younging towards the west. Sample 15 is fine-grained and contains iron-rich layers that trend at 345° and dip 77° west. In thin section this sample contains zircon and iron oxide, in addition to those minerals mentioned above. It is fine to coarse-grained, and the framework grains are subrounded to angular, and in most instances equant, although some are elongate. Sorting is poor, and some grain boundaries are coated with sericite and Fe-oxide.

Between the outcrops of Samples 15 and 16, the strike of the quartzite swings to 280 degrees and the dip changes to 64° north. Sedimentary structures other than bedding are lacking (fig. 19). It becomes conglomeratic, as can be seen in Sample 16. This sample is at an advanced stage of weathering and contains specks of muscovite. An interval within the conglomeratic quartzite contains finer grained arkosic sandstone. It is very coarse-grained, subangular and inequigranular. The muscovite is concentrated along grain boundaries.

Sample 17 contains sillimanite and opaques. It is very fine to very coarse grained, subangular, and the minerals are aligned to form a distinct foliation. The quartzite grades into a schistose outcrop 30m wide trending 345° , with a dip of 69° west. In places the contact with

the conglomerate is sharp. Sample 18 was collected from this outcrop (fig. 20). It consists of sericite, sillimanite, quartz, chlorite and opaques. It is a fine-grained foliated pelitic schist (fig.21).

Sample 25 is a quartzite from the massive Vondekop outcrop which forms a plateau. The trend of this quartzite is 055° and dips 82° southeast. It is a coarse-grained quartzite and is not aligned to the above-mentioned outcrops. On the west side of the plateau, the beds strike 015° and dip 45° east. It is coarse grained. Sample 25 also contains zircon and opaques. It is angular. Sample 30, a pelitic schist, was collected at the bend of the Mkhondo River. Similar to Sample 18, it consists of sericite, sillimanite, talc, chlorite, and minor opaques. The sample is very- fine grained, and foliated with mineral alignment. Sample 51 belongs to a 10m thick banded iron-formation lens which is sandwiched between low-grade Mozaan quartzite at Mahamba mountain (fig.22). The iron-formation has a strike of 350° and dips at 37° E, parallel to the general trend and dip of the surrounding quartzites. It is composed of magnetite that poikilitically encloses ilmenite. Quartz, some of which is rounded, is common at the base of each cycle. The sample is fine grained, and some of the ilmenite shows magnetite exsolution textures.

Sample 53, collected from the Mahamba Gap, is composed of quartz, sericite, muscovite and zircon. It is medium to coarse grained and the grains are angular and poorly sorted, but there is some semblance of grain alignment. Sample 54, taken farther away, south of the Mahamba Gap, closely resembles Sample 52.

Figure 19. Weathered micaceous quartz arenite. There are no primary structures other than bedding in this outcrop. (Coin is 2 cm in diameter).

Figure 20. A phyllite lens (knobby-weathered surface) within Mozaan quartzite. (Coin is 2 cm in diameter).

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Figure 21. Photomicrograph of a foliated pelitic schist. (crossed polars). The coarse-grained cluster of quartz and chlorite represent retrograde metamorphism in a fine-grained matrix of sericite, pyrophyllite and kaolinite. Width of field of view is 4 mm.

Figure 22. Orthogonal joints as seen on near-horizontal bedding surface of iron formation within the Mozan quartzite on the east side of Mahamba mountain. Coin is 2 cm in diameter.

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2.6 MAFIC INTRUSIONS

Mafic intrusions in the area may in part be coeval with the Usushwana intrusive suite north of the study area; some may be younger than the Karoo sediments. The mafic intrusions include diabase dykes and sills, a small-scale pyroxenite outcrop, and a layered intrusion previously thought to be a banded iron-formation.

The pre-Karoo dykes are mainly diabase and gabbro. These crop out in the central parts of the area, and all have a definite north-south trend. The post-Karoo dykes occur throughout the map area and do not have any obvious preferred trend.

One amphibolite outcrop forms a border between the Mahamba Gneiss and the Karoo sediments. The other outcrops are in the eastern part of the area, near the Vonde River (fig.23).

The diabase is composed mostly of calcic plagioclase and varieties of pyroxene, with or without olivine (Appendix 1, p.100). In thicker sills that have cooled slowly, some differentiation has occurred, allowing early olivine and pyroxene to sink under gravity to concentrate near the base of the sill. Upwards, plagioclase becomes more sodic and the pyroxenes more iron-rich, and small amounts of micropegmatite appear locally.

Figure 23. An amphibole-bearing pyroxenite belonging to the Mkhondo metamorphite, showing distinct layering. Coin is 2 cm in diameter.

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The Karoo dolerites have ages which range from 187 to 155 Ma; some of these dolerites may have acted as feeders to younger flows that have subsequently been removed by erosion, although the hydrostatic conditions may have been such that most of the dolerite formed sills that never reached the surface.

LATE TRIASSIC TO JURASSIC VOLCANIC AND HYPABYSSAL ACTIVITY

The sediments of the Karoo basin are topped by the Drakensberg Volcanics, which are exposed in the mountainous parts of Lesotho. The Drakensberg lavas have a maximum thickness of 1400m in northern Lesotho, and are made up of tholeiitic basalts with little or no olivine, but a few olivine-rich basalts occur near the base. Some individual flows are more than 40m thick. The tops of most lava flows are vesicular and thus susceptible to rapid weathering, but because little or no weathering appears to have taken place, rapid deposition of successive flows is inferred.

The volcanic rocks are part of a volcanic and hypabyssal suite that was emplaced during the late Triassic, through the Jurassic, and into the Cretaceous, embracing a time span of at least 100 million years.

The basal part of the Drakensberg volcanics has been dated at 187 $\pm$ 7 Ma by Fitch & Miller (1971), using the K-Ar method. The Lubombo mountains, along the eastern Swazi border, have two groups of basaltic flows which are separated by a thick sequence of more

felsic, predominantly rhyolitic, volcanics which have been dated at 190-194 +/- 12Ma (K-Ar). The spread in age of volcanics and related rocks elsewhere in Gondwanaland defines a spectacular province causally linked to the disruption of that supercontinent (fig.24).

The flows are typically horizontal, but in the Lubombo Mountains they have been warped into a monoclinal structure. The total thickness of this sequence is ~ 10km, with dips being greatest in the middle felsic part of the succession, which includes huge volumes of pyroclastics, and some ignimbrites.

The present phase of rift faulting in East Africa is of Tertiary or more recent age, but in a number of sectors of the East African rift system, a similar pattern of faulting of Jurassic age can be established. A change in the character of the igneous rocks is matched by a change in tectonic setting, from undeformed tholeiites of the Drakensberg to the deformed succession in the Lubombo, where felsic volcanic rocks and intrusions are present. In East African regions of major dislocations, alkaline rocks and carbonatites become increasingly important.

There is a relationship between the volcanic suite and the widespread assemblage of hypabyssal Karoo dolerites. In the Karoo basin, sills and near-vertical dykes are abundant (fig.25). To the north, distribution of the dykes is strongly influenced by fracture patterns

extending from the rift valley system of East Africa.

In the Karoo basin there is a gradation in attitude between concordant bodies and dykes, but many of the intrusions are broadly concordant and are thus sills. There are cases in which the sills transgress for short distances from one horizon to another.

The sill thicknesses range from < 1 to $> 300\text{m}$; individual dykes that have been traced for 85km are thin, between 2 and 10m in width. Rare examples of ring-dykes occur in the southeastern Karoo.

Mafic intrusions are very abundant in the study area. They have varying contents of ferromagnesian minerals; their general composition includes orthopyroxene, clinopyroxene, plagioclase, potassium feldspar, amphibole, sericite, chlorite, accessory opaques, and some contain minor amounts of serpentine. The rocks range from very fine to coarse grained, and the minerals range from anhedral to euhedral. Sample 14, an amphibole-bearing pyroxenite, (Fig.26), was collected southwest of the quartzite. It is from an outcrop of differentially weathered dark grey to black pyroxenite which is coarse-grained and surrounded by quartzite and granite. The rock is made up of orthopyroxene/clinopyroxene and actinolite. This is a fine to coarse grained sample in which most of the minerals are subhedral to euhedral. Amphibole is an alteration product of the pyroxene. Sample 19 is from a medium- grained dyke trending 040° . This dyke has onion- skin weathering and pock-marked differential weathering in different places. The plagioclase is euhedral, but the rest of the

minerals are anhedral. Sample 23 was collected from a dolerite inlier which forms high ground within the Nhlanguano Gneiss.

A coarse-grained gabbro dyke, one of the pre-Karoo mafic dykes (fig. 27), is exposed at the confluence of the Mlambomude and Mkhondo rivers. It is cut by an aplite dyke. In sample 31, relic olivines and pyroxenes can be seen in some of the amphiboles. Intergrowths of amphiboles are abundant, and could have formed from calcic pyroxene. Much of the plagioclase is antiperthitic; graphic intergrowths occur in granophyric patches. This is one of the pre-Karoo mafic dykes.

A very fine-grained dolerite dyke, 100m south of a thermal spring on the banks of the Mkhondvo River, trends 030° and is displaced by a fault. An outcrop of dolerite which shows differential weathering and blocky fracture patterns is exposed in a creek that flows north into Mkhondo River (fig.28). Sample 45 is coarse grained and contains graphic quartz; the plagioclase subhedral and antiperthitic, and a granophyric texture is locally developed. An occurrence of metamorphosed iron-formation (Sample 12, fig. 29) which is part of the Mozaan, has a dip of 76°W and a strike of 330°, the same as the 3.0 Ga quartzites that sandwich it. This sample contains green hornblende, actinolite, garnet, orthopyroxene, quartz, ilmenite and magnetite. Magnetite has exsolved from the ilmenite in distinctive rhombic patterns. The green hornblende poikilitically encloses actinolite.

Figure 24. Distribution of Karoo volcanic rocks in southeastern Africa (after Vail 1970).
Precambrian volcanic rocks not shown.

Figure 25. Karoo diabase dyke swarms in southern Africa (after Vail 1970).

Figure 26. An amphibolitic pyroxenite that is inferred to have intruded the Mozaan quartzite, showing distinct layering. Coin is 2 cm in diameter.

Figure 27. Mafic dyke, intruded by a narrow felsite dyke. Coin is 2 cm in diameter.

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**Figure 28. A post-Karoo dolerite showing pock-marked differential weathering.
Coin is 2cm in diameter.**

**Figure 29. Metamorphosed Mozaan iron-formation (now pyroxenite), with the layering still preserved.
Coin is 2cm in diameter.**

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2.7 THE KAROO SUPERGROUP

The Dwyka Formation

The lowermost unit of the Karoo Supergroup is the Dwyka Formation, which unconformably overlies rocks of Precambrian to Early Palaeozoic age.

The Permo-Carboniferous glaciation recorded throughout Gondwanaland has been studied extensively (Truswell, 1977). During this glaciation the Dwyka glaciers in Southern Africa carried fragments of enormous size, and deposited large volumes of poorly sorted till that consolidated to form tillite. Because rocks similar to tillite may form as a result of landslides, submarine mudflows, slumping, turbidity currents and debris flows, the term diamictite is now used for any such sediment. To differentiate, one has to look for glacial sedimentary indicators such as till fragments that are striated and faceted, or for a distinctively polished, grooved and striated pavement over which the glacier moved.

Dwyka diamictite is locally up to 1000m thick in the axis of the depositional trough in the southern part of the Karoo basin. The thickness decreases rapidly northwards, where the unit is erratically preserved only in thin layers and pockets. At the northern margin of the basin, these are overlapped by Ecca strata.

The diamictite is made up of poorly sorted coarse, angular fragments

of varying size set in a dense greenish-blue argillaceous matrix. The rock is unbedded to crudely stratified. In southern Africa, as elsewhere in Gondwanaland, the sediment must have been shed from one or more continental ice sheets.

Around the margin of the northern outcrop of the diamictite, several localities where the pavement over which the ice moved are preserved. Relevant data show that most of the ice movement in southern Africa can be related to an ice sheet centred in northern Transvaal, eastern Botswana, Zimbabwe and Zambia. A crude radial pattern out from this centre reached as far north as the Congo Basin and as far east as Madagascar. In South Africa and Swaziland, movement from this ice sheet has been interpreted to represent a succession of ice lobes moving broadly southwards from the Transvaal (fig.30).

The tillite (diamictite) consists of clasts in the pebble to boulder range in a silt/clay matrix. The clasts at the base of the Dwyka come from surrounding country rocks, but higher in the succession the clasts have been transported from farther (250km). The clast size increases to a maximum in the north. As the size increases, the clasts are better rounded. The pebbles and boulders are subrounded to rounded, with moderate to high sphericity, whereas sand-size grains are angular. In fresh samples of tillite, minerals that normally weather easily show little or no sign of alteration.

Ecca Group

The Ecca Group overlies the Dwyka Formation and consists of 100 to 300m of sediments which include coal seams, carbonaceous shales and feldspathic sandstones. The sediments are predominantly of lacustrine origin, and contain large coal seams (Truswell,1977). The clastic sediments of this unit have sources that lie almost entirely outside the coastline of southern Africa. The differing source areas are indicated by directional structures in the rocks and by differences in composition. Three facies, the Southern, Western and Northern facies, have been recognized by Truswell, 1977. Their distribution and directions from which the material was transported, are shown in figure 31.

The Southern Ecca facies is confined to the southern margin of the basin. The sediment that accumulated in this linear belt is very thick, indicating that the subsidence that accommodated the Cape rocks and the thickest development of the Dwyka had continued along this linear belt. The sediments of the Southern facies are greenish grey shales and greywackes. Sedimentary structures associated with the greywackes indicate that these sediments accumulated from turbidity currents during rapid subsidence. The Western Ecca facies in the western portion of the basin comprises bluish black shales and sandstones.

The Northern facies is distinctive. Besides the shales that represent the lower and upper units, the middle Ecca coal measures contain coarser detrital sediments such as conglomerates and arkosic sandstones, in association with the shale and coal beds.

Figure 30. Transport directions in lobes of Dwyka glaciation in Southern Africa (after Crowell and Frakes 1972).

The Eccra coal measures contain upward-coarsening cycles in which silts, alternating silts and sands, and sands that coarsen and become more massive upwards, reflect a single cycle. The cycle comprises a typical deltaic sequence, in which the coarser detritus represents building out of the delta above pro-delta muds. The top of the cycle is incised by fluvial channels. The sediments of this facies were transported from the east and northeast and deposited in a series of deltas. Many of the coal deposits of the Northern facies, such as those in northern Natal and Swaziland, represent the accumulation of rotting vegetation in back swamps and lakes upstream from deltas in the alluvial floodplain.

In the study area, the above mentioned units occur only in the subsurface, where the lower part of the Eccra rests conformably on the Dwyka. This represents the western edge of the vast Karoo Supergroup that extends southward into South Africa. According to data from a single drill hole in the study area, the Karoo may rest unconformably on the Mozaan quartzites, although the less-than-one-metre of quartzite penetrated before the hole was abandoned may be the top of a large quartzite dropstone in the Dwyka (The drilling technicians were under the impression that they had drilled through the Karoo and had reached the Pongola Supergroup).

Sample 47, from the chilled contact of dolerite with the Karoo shale (fig. 32), is fine grained and mottled. It is made up of 45% sericite, 20% orthopyroxene, 5% quartz, and 15% carbonaceous matter.

Figure 31. Facies and paleocurrent directions for the Ecca (after Ryan 1968).

Description of Borehole Core

The borehole yielded overburden down to a depth of 9.2m, before core was recovered. The core comprises 5.0m of dark carbonaceous shale and at the bottom of the hole, 80cm of orthoquartzite which was initially interpreted as the top of a large dropstone derived from the Mozaan Group. However, this quartzite (B3-01) shows bedding inclined to the same attitude as bedding in nearby exposures of Mozaan quartzites, supporting the interpretation of it as in situ bedrock. Made up of quartz, sericite, muscovite, andalusite and zircon, it is medium to coarse-grained and the grains are angular. The quartz has undulose extinction. The presence of andalusite further supports the view that the quartzite, metamorphosed at low pressure and temperature, is an extension of the Mozaan bedrock to the west.

B3-02 (13.5m depth), is an arkosic wacke that consists of quartz, sericite, plagioclase, microcline, muscovite, biotite, chlorite, iron oxides, lithic fragments, garnet, and minor carbonaceous material. It is fine to medium grained, angular to subrounded and is poorly sorted.

**Figure 32. An indurated shale at the contact of post-Karoo dolerite and Karoo Supergroup.
Diameter of coin is 2 cm.**

B3-03 (14m depth), is from arkose consisting of quartz, microcline, pyrite, sericite, plagioclase, biotite, muscovite, garnet, zircon, tourmaline, magnetite and lithic fragments. The sample is medium to coarse grained, subangular to rounded, and almost equigranular. Pyrite locally occurs along fractures and grain boundaries.

B3-04 (25.5m depth) is a very fine grained carbonaceous shale composed of quartz, muscovite, lithic fragments, calcite and carbonaceous material.

B3-05 (41.3m depth), is made up of quartz, plagioclase, microcline, muscovite, biotite, graphic quartz, sericite, carbonaceous material and opaques. It is fine to medium grained, angular to rounded and poorly sorted, but shows graded bedding. This sample is an arkose-rich shale. B3-06 (45.5m), is very fine and comprises quartz, muscovite, lithic fragments, chlorite, carbon-rich matrix, iron oxides and < 1 pyrite. Some of the quartz forms fine-grained clasts, and the lithic fragments are carbonate-rich.

B3-07 (49m), is also very fine-grained and consists of quartz, mica, calcite, lithic fragments, sericite and carbonaceous material. Graded bedding is well developed. Some quartz forms fine-grained clasts. The lithic fragments contain carbonaceous material and calcite. A calcite vein is almost parallel to the bedding. Graded bedding is well developed.

The sample is very fine grained. Some quartz is in clasts, and the lithic fragments are carbon and This sample, which is graded, is from the top of the Dwyka, and consists of quartz, sericite, mica, calcite, plagioclase, calcite and rock fragments, magnetite, pyrite and chalcopryrite. The sample is very fine to coarse grained. Iron oxides occupy veins parallel to bedding. This sample which is graded, belongs to the top of the Dwyka.

B3-10 (140m) comprises quartz, feldspar, sericite, mica, garnet, lithic fragments, chlorite, zircon, carbonaceous material, opaques and epidote. It is very fine to coarse grained, angular to subrounded. Lithic fragments are made up of quartz and calcite. One clast displays possible spinifex texture. Some of the clasts are of volcanic origin. A very fine-grained sample, B3-11 (167m), contains quartz, lithic fragments, mica, sericite, calcite, carbonaceous material and iron oxides, and shows cyclic graded bedding.

B3-12 (170m), is made up of quartz, lithic fragments, feldspar and calcite matrix. The lithic fragments are quartz-rich.

CHAPTER 3

METAMORPHISM AND PETROGENESIS

3.1 METAMORPHISM OF THE PONGOLA SUPERGROUP

Development of the Barberton greenstone terrane to the northwest was succeeded by deposition, metamorphism and folding of the Pongola Supergroup (Matthews, 1987). The Pongola displays progressive northward increase in grade of metamorphism in Natal, compared to Swaziland where much of the Pongola strata have been subjected to high-temperature metamorphism. The general trend of metamorphic grade of the Pongola Supergroup is from low grade (andalusite locally present) in the southwestern parts of the study area, to sillimanite grade in the northeastern part of the area. The metamorphic grade also increases from west to east, where the sediments have been affected by intrusion of the Hlatikulu Granite, resulting in the growth of sillimanite, which reflects medium-temperature metamorphism. Near the contact, the mineral assemblage sillimanite - muscovite - quartz, puts the metamorphic grade at amphibolite facies, reflecting moderate pressure and temperature (fig. 33). Along the western edge of the study area, sediments contain the assemblage andalusite-muscovite - quartz reflecting lower-temperature grade for this part of the area, farther from the granitic intrusion.

The aluminosilicate polymorphs, (andalusite, kyanite and sillimanite)

are sensitive to pressure and temperature changes because they have the same mineral composition but different crystal structures. Usually one of the polymorphs is present in a particular metamorphosed rock of appropriate composition. When a critical temperature or pressure is reached, one variety transforms to another.

Andalusite $\longrightarrow$ Kyanite (P increase, where T is constant)

Andalusite $\longrightarrow$ Sillimanite (T increase, where P is constant)

Kyanite $\longrightarrow$ Sillimanite (T increase, where P is constant).

The Mahamba samples do not contain the kyanite polymorph, and this means that the changes between andalusite and sillimanite were a result of variation in temperature. Pyrophyllite, kaolinite, sericite and chlorite are thought to be alteration products due to retrograde metamorphism. The absence of potassium feldspar in the sillimanite assemblage limits the temperature of metamorphism.

The scanning electron microscope (SEM) was used to identify the compositions of minerals of the different rock units in the study area. The SEM works by producing an electron beam from a heated tungsten filament located at the end of a column operated under high vacuum. The electrons are accelerated to a high energy by application of a large acceleration potential (20 Kv) to a grid cap located in front of the filament tip. The electrons travel down the column to the sample chamber, and are focussed to a fine probe by a series of apertures and lenses.

A solid state silicon detector, called the energy dispersive spectroscopy

(EDS) detector, detects X-rays produced by the high energy electron beam. The rays are converted to voltages and amplified. The voltages are proportional to the energy of the detected X-rays.

X-rays are collected from mineral standards and the net counts compared to those obtained from the sample.

Figure 33. Pressure - Temperature dependence of the three Al_2SiO_5 polymorphs.

3.2 THE OXIDES AS PETROGENETIC INDICATORS

Oxygen fugacity plays an important role in oxide formation. Frost (1991), described oxygen fugacity as a variable that tells us whether a mineral (like iron) is likely to occur in its native state, as a divalent cation in a silicate, or as a trivalent cation in an oxide. In natural systems, however, the presence of magnesium (5th most abundant element in crust) and titanium (10th most abundant) also play an important role in determining the relative stabilities of silicates, magnetite and ilmenite. Mg and ferrous iron substitute for each other in a variety of silicates. In many rocks, oxygen fugacity is a function of both the Mg/Fe ratio of the silicates, and the Ti content of the oxides.

The composition of Fe-Ti oxides and the composition of coexisting silicates thus provide a monitor of oxygen fugacity. Furthermore, oxygen fugacity is important in determining the composition of any fluid phase that is associated with igneous and metamorphic rocks. It should be noted that oxygen fugacity also has an important influence on the stability of graphite and sulfides.

The summary presented in the rest of this section is taken from Frost and Lindsley, (1991, pp 433-438).

Fe-Ti in plutonic rocks: The composition of spinel in plutonic rocks tends to re-equilibrate readily during cooling at high temperatures. The re-equilibration reactions involve inter-grain exchange of the

ferric iron and titanium between titano-magnetite and ilmenite, with consequent enrichment of Fe_3O_4 in the spinel, and of FeTiO_3 in the rhombohedral phase. At intermediate temperatures, Ti-magnetite re-equilibrates by the process of "oxyexsolution" in which the Fe_2TiO_4 component of the titanomagnetite oxidizes to form granules of ilmenite within or around the spinel. At lower temperatures the process produces lamellae of ilmenite in (111) planes of the host spinel. In addition to the decrease of Ti in the spinel during cooling, both the spinel and the rhombohedral oxide will lose Mg during exchange with the surrounding silicates. As a result of these processes, the host spinel in many plutonic rocks is nearly pure magnetite, with small amounts of Ti and only trace amounts of MgO and Al_2O_3 remaining in solid solution, except for oxide-rich layers in some intrusions which have relatively high contents of MgO , because they are isolated from the silicate minerals.

Fe-Ti in volcanic rocks: The dominant spinel phase in magmatic rocks is either chromite or titanomagnetite. In some basalts, both Cr-spinels and Fe-Ti spinels are present. In many assemblages the spinel is included in olivine and may be xenocrystic and not related to the surrounding phenocryst assemblage. In others there is evidence that crystallization proceeded from chromite to titanomagnetite during evolution of the melt. Chromite occurs in rocks that are relatively high in Mg as well as Cr, whereas titanomagnetite is characteristic of those rocks that are poor in Mg.

A wide range in the Ti content of spinels is shown by volcanic rocks. Almost pure ulvospinel occurs in highly reduced lavas. The spinels of tholeiitic rocks are enriched in Ti, with spinel from MORB rocks ranging from 50-80% ulvospinel. These spinels have a substantial amount of minor components, with Al_2O_3 ranging from 1-2 weight %; Cr_2O_3 , MnO and V_2O_5 from 0.5-1%. The MgO content varies from 0.1-3%, a reflection of the rapidity at which spinels undergo cation-exchange with other minerals. The spinels from andesites and dacites have lower Ti than tholeiites, although- minor element components are similar to those of tholeiites.

Alkalic rocks: tholeiites are richer in spinels than alkalic basalts, but alkali basalts are richer in minor components than tholeiites and andesites. Al_2O_3 and MgO contents are as high as 6% and 5% respectively. Spinel from these rocks contain moderate amounts of Ti, but there is no consistent trend in Ti content as a function of rock type. The distinctive feature is that the spinels have a tendency toward a high Al_2O_3 content. The high Al content of spinels from undersaturated rocks reflects the fact that the aluminium content of spinels from a rock containing olivine, clinopyroxene and plagioclase is a function of silica activity:

Decreasing silica activity will cause the reaction to move to the right,

thus increasing the activity of alumina in the spinel. The aluminium content of spinel from the assemblage olivine - diopside - plagioclase - spinel can act as a geobarometer, if an independent estimate of silica activity can be obtained.

Silicic volcanics: spinel from these rocks shows a complete range in Ti content from Usp100 in fayalite rhyolites to Usp10 in other rhyolites. The Ti content of titanomagnetite is a function of the oxygen fugacity at which the magma crystallized. The spinels of relatively reduced fayalite rhyolites have high Ti, whereas those of the more oxidized rhyolites with biotite and hornblende have low Ti.

Minor element abundances in spinels from silicic volcanics are variable. Al_2O_3 is in the range of 1-2% in spinels from fayalite rhyolites, but may be as high as 3% in peraluminous rhyolites. MgO is less than 1% in iron-rich fayalite rhyolites, but 2% in more magnesian rocks. MnO is normally less than 1%, but some lavas contain spinel with 2-3% MnO . Decreasing temperature increases the MnO from 0.5%-1%, but decreases MgO from 0.7 to 0.3%.

Ilmenite and hematite:

Tholeiitic volcanics: ilmenite is absent from some sea-floor basaltic rocks, but is present in others, where it usually has low to moderate Fe_2O_3 contents. The significant minor component in ilmenites from tholeiitic rocks is MgO with a range from 1 to 3%. MnO is normally

low in the ilmenite of tholeiites, ranging from 0.5 to 1%.

Calc-alkalic lavas: ilmenite is absent or rare in both high-Al basalts and basaltic andesites. The minor element contents of ilmenite from andesitic rocks are similar to those from tholeiites, with MgO = 1 to 4% and MnO = 0.5 to 1%.

Alkalic lavas: alkali basalts contain ilmenite with titanomagnetite as the only Fe-Ti oxide. The range of the Fe_2O_3 component of the ilmenite is 3 to 8 wt %. The most abundant element in the ilmenite is Mg (1 to 5 wt % MgO). MnO contents are similar to those of the tholeiites (0.5 to 1%.) **Silicic lavas:** ilmenite from silicic volcanic rocks shows a wide range of hematite content. Ilmenite from fayalite-bearing rhyolites has the lowest hematite contents, ranging from Hem_{1.5} to Hem₇. The evolved nature of these rocks causes the MgO content of the ilmenite to be low (<1%), and the MnO content to be high, (1 to 2%). Ilmenite with the lowest hematite content comes from assemblages that lack either quartz or magnetite, whereas that with the highest hematite content is from rocks having the assemblage q-opx-mt-ilm, with moderately magnesian opx.

Plutonic rocks: ilmenite from plutonic rocks may form in two ways. It may be a product of direct crystallization, or it may be produced by oxyexsolution from titanomagnetite. It is easy to recognise oxyexsolved ilmenite texturally when it outlines the familiar (111) lamellae, but ilmenite formed by oxyexsolution at high temperatures tends to

migrate to grain boundaries, where it can form separate grains or overgrowths on primary ilmenite. In some cases ilmenite formed by oxyexsolution can be distinguished from primary ilmenite by its high MnO content, but in other cases there is no consistent relationship between the composition of oxyexsolved ilmenite and discrete grains of ilmenite.

Mafic plutons: although absent in some mafic plutons, primary ilmenite is generally present in most mafic intrusions. In some it is the dominant Fe-Ti oxide. The ilmenite is poor in ferric iron, with a hematite component of less than 20 mole %. MgO in ilmenite of these rocks is variable (0.1-2.3%), as is MnO (0.1-1.4%). The MgO values are lower than in the volcanics, showing that the composition of ilmenite in mafic plutons has been significantly altered by ion exchange with the silicates during cooling.

Felsic plutons: ilmenite from felsic plutons has a compositional trend similar to silicic volcanics. In plutons containing fayalite, ilmenite is poor in ferric iron, with Hem₆ or less, especially in biotite- and hornblende-bearing granites, where magnetite is absent. Ilmenite from magnetite-bearing granitoids is moderately rich in ferric iron. The MnO content of ilmenite from felsic plutons is low in fairly unevolved granitoids (> 3%), and more variable in evolved fayalite-bearing rocks, ranging from <1% to 11%. Ilmenite from granites may be very rich in MnO (~ 30%).

2 of/de 2

PM-1 3½"x4" PHOTOGRAPHIC MICROCOPY TARGET
NBS 1010a ANSI/ISO #2 EQUIVALENT

PRECISION^{SM} RESOLUTION TARGETS

Samples from the study area:

The MnO content of ilmenite in sample 28-31, a pre-Karoo gabbro, is 2.7% but the sample lacks MgO. Ilmenite in sample 28-27 (Mahamba Gneiss) lacks MnO, but contains ~0.84% MgO. Ilmenite in the ultramafic pyroxenite (28-14) contains 1.4% MnO and 1.6% MgO. These values fall within the expected range for MnO and MgO in plutonic rocks. The compositions of these ilmenites must have been significantly changed by ion exchange with the silicates during cooling. Chromite from the pyroxenite falls outside the chart of Duke (1990), which is reproduced in Figure 34. This chart is for podiform and layered mafic intrusive units, and thus the chromite has probably undergone metamorphism.

Magnetite and ilmenite are adjacent to each other in pre-Karoo diabase (sample 31), and have given the temperature of formation of the diabase as $580^{\circ} \pm 10^{\circ}$ using the magnetite-ulvospinel and hematite-ilmenite chart of Lindsley (Fig.35).

Figure 34. Variation of $Mg/(Mg+Fe)$ and $Cr/(Cr+Al)$ in chromite from stratiform and podiform deposits. From Duke, 1990.

Figure 35. Composition of coexisting (Usp-Mt) and (Ilm-Hem) plotted as functions of T and oxygen fugacity. From Lindsley, 1976.

3.3 THE GARNET-BIOTITE GEOTHERMOMETER

Garnet - biotite thermometry has a potential value based on observations in metamorphites. Ferry and Spear, 1978, suggested that the technique should be restricted to garnets with low Ca and Mg, and biotite with low Ti, because solid solution involving additional components normally affects this thermometer.

Using SEM analyses of adjacent biotite and garnet grains for two locations in a single sample for geothermometry of the Mahamba gneiss (Appendix 2, p. 105-107), the temperature of formation was calculated using Mg and Fe^(Total) values. The Fe²⁺ values in biotite were estimated by multiplying the Fe atomic proportions by the factor 0.86, and the distribution coefficients (K_D) of Fe²⁺ and Mg in biotite-garnet were calculated using composition ratios (Ferry and Spear, 1978; Perchuk and Lavrent'eva, 1983; Kretz, 1990) for the two minerals:

$$K_D = \frac{X_{Mg}^{bt}}{1 - X_{Mg}^{bt}} \times \frac{1 - X_{Mg}^{gt}}{X_{Mg}^{gt}}$$

where X is Mg/Mg + Fe²⁺ in the biotite and garnet.

The atomic ratio Ca/Mg + Fe + Ca + Mn was also calculated, and then, using the formula from Kretz (1990, and unpublished results):

$$T (^{\circ}K) = \frac{2780 + 258[Ca]^{gt}}{\ln K_D + 0.235[Ca]^{gt} + 1.51}$$

the temperature of metamorphism was calculated to be 670°C.

This temperature is in agreement with upper amphibolite to granulite facies of the Early Precambrian, during which time the protoliths were originally formed as granitoid plutons, mostly of tonalitic composition.

SEM Analyses of Biotites in relation to Rock Types

Results of the SEM analyses of biotites are shown in Appendix 2. The charts show subtle compositional differences in the rock units; Al contents have atomic proportions in the range 0.21 to 0.24. The plot Fe vs. Fe/Fe + Mg (appendix 2.8-1 and 2.8-3) is 0.24 to 0.44 in the Mahamba Gneiss, 0.38 to 0.45 in the Nhlangano gneiss (appendix 2.8-2), and 0.3 to 0.4 in the Hlatikulu Granite and Karoo sediments (appendix 2.8-4). The Ti vs. Al plots (appendix 2.2), show that the two gneisses have the same range of values (0.01 to 0.05 Ti), but there is a smaller range in the granite and sediments (0.03 to 0.05). Al contents and Fe/Fe + Mg ratios are indicators of conditions that prevailed in the parent magmas, with Al indicating peraluminosity.

Separate charts were drawn, each to represent a different rock unit. The atomic elemental trends vs. Fe/Fe + Mg ratio were then studied in an effort to find the provenance of the Karoo sediments.

Sample 28/27 (Mahamba Gneiss) chart was eliminated as a possible source because its Fe (appendix 2.8-3) and Mn (appendix 2.6-3) contents are lower than those of the sediments, and Mg values are higher. The Al values for the sediments (Append. 2.1-4) are high because of the possible effect of the alteration of biotite into clay, and it is assumed that this takes place at the same time as the chloritization

of K (Append. 2.3). Mg of the Nhlanguano Gneiss (Append. 2.2-2) is lower than that of the sediments, and therefore this gneiss cannot be the source. The sediments and Hlatikulu Granite represented in chart 4 of each of the appendices 2, have most of the component elements falling in the same field for these two, thus making the granite the strongest possible contender for the source.

The calculations were done with the assumption that when the rocks cooled, the equilibrium state was reached. If this were not the case, then the margin of error will be high.

In the case of calibration, the experimental results of the Ferry and Spear equation are consistent, but the relation of K_D vs. T is not fully understood, and therefore some element of uncertainty exists. Also, analytical error could have been reduced if ten or more analyses had been obtained and averaged. However, a huge error might have resulted from averaging only six analyses per sample thin sections. The larger the number of analyses, the smaller the error range would be. Mg and Fe errors would propagate to the K_D calculation.

Ca analyses from which the $Ca/Mg+Fe+Ca+Mn$ ratios were calculated to obtain the Ca concentration, could have increased the error. Therefore a practical resolution of approximately 50° has been used (as proposed by Ferry and Spear, 1978) as the margin of error if all the above constraints are taken into account. This means that the garnet-biotite geothermometer is $670^\circ \pm 50^\circ$.

3.4 PONGOLA QUARTZITES

According to Wronkiewicz & Condie (1989), Pongola quartzites are enriched in SiO_2 , Ni and Cr, and depleted in TiO_2 , Sr, Th, V, Nb, Ta, Zr and Hf relative to quartzites elsewhere in the Archean. $\text{SiO}_2/\text{Al}_2\text{O}_3$, $\text{K}_2\text{O}/\text{Na}_2\text{O}$, La/Yb and Eu/Eu* ratios are also generally greater than those of Phanerozoic sandstones. In addition to higher SiO_2 and lower Al_2O_3 contents, the Pongola quartzites are also generally depleted in Se, V, Ni, Y, Nb, Ta, Hf and REE relative to Pongola pelites.

The Chemical Index of Alteration (CIA) of Nesbitt and Young (1982) for Pongola pelites falls near 78 (Wronkiewicz & Condie, 1989). These values are transitional between those of the older Moodies Group (CIA-59) and Witwatersrand pelites (CIA-83). Although uncertainties exist in interpreting the CIA because of possible mobility of alkali and alkaline earth elements, the CIA data appear to record moderate to intense chemical weathering during deposition of the Pongola and Witwatersrand Supergroups.

Pongola pelites have CIA values that decrease stratigraphically upwards (Wronkiewicz & Condie, 1989). Initially high values (CIA-79) may reflect erosion of deeply weathered source terrane during deposition of basal Pongola sediments. With increasing tectonic activity, relatively unweathered source material may have begun to enter the Pongola basin, resulting in a decrease in CIA values (71). This possibility is supported by deeply weathered palaeosols underlying

basal Pongola sediments in the north. During deposition of Mozaan sediments, tectonic and volcanic activity appear to have subsided, resulting in decreased erosion rates and increased flux of chemically weathered detritus into the basin. This explains the increase in CIA values (CIA-77) during Mozaan time.

Wronkiewicz & Condie (1989), examined palaeo-weathering conditions with the $K_2O-Fe_2O_3-Al_2O_3$ ternary diagram of Reimer (1985) (figure 36). Nsuze trend-A pelites may have been derived from weathered Al_2O_3 -rich detritus. Trend-B pelites appear to reflect erosion of a tectonically or volcanically active source, where a large proportion of relatively fresh rock was deposited in the basin. Thus, detritus may have changed from kaolinite - and/or illite-rich weathered rock to muscovite-rich, relatively unweathered granite or metamorphic rocks, with time.

Figure 36. Major element distributions from the Pongola Supergroup (after Reimer 1985).

PELITIC PROVENANCE

Although source-rock composition appears to be the dominant factor controlling the composition of fine-grained terrigenous clastic sediments, a complex interplay of a number of factors, including weathering, hydraulic sorting during transport, element adsorption onto clay particles, tectonic setting, diagenesis and metamorphism may also have had an effect. Condie & Wronkiewicz (1989) have shown that Pongola sediments have subtle but persistent geographic variations in some major and trace elements, indicating the presence of local source heterogeneities and contemporaneous volcanism. Some variations in pelite composition may be related to mineral fractionation in the sedimentary environment. The increasing REE contents are paralleled by increasing Al_2O_3 , suggesting that the amounts of clay minerals originally contained in these sediments may have controlled REE concentrations.

NASC (North American Shale Composite of Gromet et al., 1984) - normalised element distributions indicate a compositional similarity between Pongola pelites and Phanerozoic shales. These similar distributions suggest that the Pongola source was similar in composition to the average Phanerozoic upper-continental crust. The Pongola pelites originated from a source that was rich in granite, and only minor contributions were from tonalite and komatiite. The Nhlanguano Gneiss must have had a precursor that was tectonically

uplifted, producing sediment that was deposited in the Pongola basin. The upper crust of southern Africa apparently went through changes in composition during the Archean and Early Proterozoic, from largely mafic in the initial stages to largely felsic in the later stages of development. However, there must have been intermittent episodes of mafic magma injection to account for the schistose layers and lenses within the Mozaan.

An attempt to analyse biotites of the Mozaan pelites to obtain fingerprints to infer the pelitic provenance proved futile, because most of the biotite has been altered to chlorite.

3.5 THE KAROO

Detritus in the Late Permian-Early Triassic Eccu sediments of the Karoo seems to have been derived from younger granite plutons of the Hlatikulu type. SEM analyses of biotites from the Mahamba Gneiss, Nhlangano Gneiss and Hlatikulu Granite show that biotite of the Hlatikulu Granite is almost identical in composition to biotite of the Karoo sediments.

Sedimentary structures of the Karoo rocks have shown that lacustrine systems were widespread in the southern African subcontinent, and that their development was linked to major Gondwana fluvial systems (Truswell, 1977). During Eccu time, lacustrine environments of the White Band/ Whitehill Shale Formation stretched across two continents, extending over 300 000 km² in South Africa, the Kalahari

in Botswana, the Kaokoveld in Namibia, into equivalent deposits of the Parana basin of South America.

3.6 CONCLUSIONS

Sediments of the Pongola Supergroup have been difficult to deal with in the study area because of a lack of primary structures and heavy minerals. The sediments were extensively altered during metamorphism of the Mahamba and Nhlangano gneisses. The metamorphic grade is greenschist through upper amphibolite to granulite facies, as indicated by the biotite-garnet thermometer of the Mahamba Gneiss. Extensive granitization and granite intrusion have taken place, and in areas in which much of the metamorphic terrane has been modified by the intrusions, the development of a granite series has been established under conditions of regional metamorphism. The rocks in the eastern part of the study area were developed at high temperatures. The rocks are characterized by the presence of andalusite in the southwest, sillimanite in the northeast, and, as with the Mahamba Gneiss, garnets. Metamorphism took place before the 2.7 Ga intrusion of the Hlatikulu Granite, which has an isotropic fabric, and also contains xenoliths of the Mahamba Gneiss. Saggerson (1973) has noted that the Pongola Supergroup, which underwent metamorphism and folding, is a classic example of prograde metamorphism at low pressures following the pattern of early Precambrian shield areas.

Biotite of the Permo-Triassic carbonaceous shale of the Karoo Supergroup shows a strong resemblance to biotite of the Hlatikulu Granite. The other units were eliminated as potential source rocks for the Karoo because some of their components were either higher or lower than those of the Karoo biotite. The Hlatikulu Granite, therefore could have been the source of detritus for the sediments.

APPENDIX 1**MODAL COMPOSITIONS BASED ON MINERAL ESTIMATION
OF THIN SECTIONS****APPENDIX 1
MODAL COMPOSITIONS BASED ON MINERAL ESTIMATION
OF THIN SECTIONS
THE MAHAMBIA GNEISS**

Sample No.	27	29	35	40	15
Quartz	20	40	50	40	40
Microcline	0	0	5	0	<2
Plagioclase	<5	0	0	10	0
Sericite	8	0	20	0	30
Orthoclase	0	0	0	5	0
Biotite	20	30	15	5	15
Chlorite	5	5	<1	0	<4
Amphibole	0	0	5	30	0
Garnet	30	20	5	10	10
Zoisite	5	5	0	0	0
Muscovite	<5	0	0	0	0
					<1

Sample No	39	43	48	57
Quartz	50	30	45	60
Microcline	20	25	10	5
Plagioclase	5	20	5	5
Sericite	15	10	15	10
Orthoclase	0	0	0	0
Biotite	5	0	20	0
Chlorite	<1	<5	<2	10
Amphibole	5	10	0	5
Calcite	0	0	<5	0
Muscovite	0	0	<2	0

NHLANGANO GNEISS

Hhlatikulu Granite

Sample No.	13	22	32	33	34	36
Quartz	40	40	80	40	30	40
Microcline	15	20	0	15	0	30
Plagioclase	10	10	10	10	2	5
Orthoclase	10	5	10	0	5	0
Sericite	20	20	0	20	35	15
Biotite	<5	5	<1	10	0	0
Chlorite	0	5	0	<5	0	10
Epidote	0	0	0	0	5	0
Amphibole	0	0	<1	0	0	0
Muscovite	0	<1	0	0	0	0

Sample No.	19	23	31	37	38	45	46
Olivine	0	5	5	<1	<3	<1	<2
Orthopyroxene	5	15	10	0	30	20	30
Clinopyroxene	20	20	10	10	25	5	5
Plagioclase	50	40	10	20	3	55	40
Sericite	10	10	5	20	20	0	0
Orthoclase	15	5	<2	0	0	5	10
Chlorite	0	<3	<5	20	20	0	0
Serpentine	0	0	50	20	0	0	0
Graphic quartz	0	0	<3	0	0	5	0
Vein quartz	0	0	0	<5	0	0	0

DOLERITE

PONGOLA SUPERGROUP ARENTES

Sample No.	11	15
Quartz	94	80
Sericite	<2	10
Muscovite	<1	<5
Kaolinite	0	<2
Zircon	0	<1
Fe Oxide	0	<1
Calcite	0	0
Biotite	0	0
Andalusite	0	0
Sillimanite	<2	5

PELTIC SCHISTS AND IRON FORMATION

Sample No.	18	30
Quartz	10	10
Sericite	30	30
Kaolinite	15	15
Pyrophyllite	15	15
Chlorite	10	10
Sillimanite	10	10
Opauques	10	<10

ULTRAMAFICS

Sample No.	51	12	14
Clinopyroxene	3	20	30
Orthopyroxene	0	30	35
Magn/IImenite	0	20	2
Amphibole	0	20	30
Garnet	0	7	0
Quartz	0	2	0
Olivine	97	<1	<1
Orthoclase		0	2
Chromite		0	<1

AMPHIBOLITE

Sample No.	24	44
Quartz	40	15
Plagioclase	0	10
Sericite	20	20
Orthoclase	<2	0
Biotite	10	0
Chlorite	0	15
Amphibole	30	30
Opakes	0	<5
Muscovite	0	<5

APPENDIX 1b**Petrographic data for rocks of the study area**

Unit	Crystallinity	Grain-size	Crystal shape	Texture
Mahamba Gneiss	Holocrystalline	Fine- to medium-grained	Subhedral to anhedral	Ganoblastic
Nhlangano Gneiss	Holocrystalline	Medium- to coarse-grained	Subhedral to anhedral	Hypidomorphic
Hlabisi Granite	Holocrystalline	Fine- to coarse-grained	Subhedral to anhedral	Porphyritic
Mafic Intrusions	Hypocrystalline to holocrystalline	Fine- to coarse grained	Anhedral to euhedral	Ophitic to granophyric
Mozaan Quartzite		Medium- to coarse grained	Angular to subrounded	Polygonal to sutured
Palite		Very fine- to medium- grained	Angular to subrounded	Foliated
Karoo Supergroup Dwyke		Fine- to very coarse-grained	Angular to rounded	Matrix - supported
Arkose		medium-grained	Angular	
Shale		Very fine- grained		Cyclic layering

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APPENDIX · 2

SEM ANALYSES FOR BIOTITES OF THE STUDY AREA

	28-35#1	28-35#2	28-35#3	28-35#4	28-35#5	28-35#6	28-39#1	28-39#2	28-39#3	28-39#4
SiO2	33.94	33.70	33.80	34.04	33.35	33.39	34.87	34.65	31.19	34.71
Al2O3	15.95	14.98	15.29	16.34	15.37	14.88	14.96	14.66	15.78	15.16
TiO2	2.58	4.27	3.93	2.38	3.41	4.12	3.44	3.30	1.16	2.58
CR2O3	.12	.00	.00	.00	.00	.00	.00	.00	.00	.00
FeO	31.60	31.16	30.78	31.15	30.63	31.39	28.17	27.85	32.39	28.67
MnO	.24	.26	.21	.25	.28	.31	.50	.47	.38	.43
MgO	2.21	2.30	2.43	2.74	2.02	2.08	4.29	4.36	5.01	4.25
CaO	.00	.00	.00	.00	.00	.00	.00	.00	.00	.00
K2O	3.32	3.25	3.43	8.86	9.13	9.15	9.32	9.20	5.13	3.39
Na2O	.00	.00	.00	.00	.00	.00	.00	.00	.00	.00
· Cl	.00	.00	.00	.00	.00	.00	.07	.00	.00	.00
TOTAL	95.36	95.93	95.32	95.78	94.39	95.32	95.69	94.49	91.04	94.99
TOT-O	95.95	95.33	95.33	95.78	94.39	95.32	95.67	94.49	91.04	94.99

FORMULA (BASIS 22 OXYGENS)

Si	5.491	5.454	5.460	5.484	5.481	5.451	5.570	5.595	5.272	5.591
Al	2.509	2.546	2.540	2.516	2.519	2.549	2.430	2.405	2.728	2.409
	8.600	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	.533	.312	.371	.587	.459	.315	.386	.385	.415	.468
Ti	.514	.520	.477	.288	.421	.506	.413	.401	.147	.312
Cr	.016	.000	.000	.000	.000	.000	.000	.000	.000	.000
Fe	4.276	4.217	4.158	4.197	4.237	4.287	3.762	3.761	4.578	3.835
Mn	.033	.036	.029	.035	.039	.043	.067	.064	.055	.059
Mg	.532	.556	.600	.657	.496	.506	1.022	1.050	1.261	1.020
	5.704	5.641	5.635	5.764	5.652	5.657	5.650	5.661	6.456	5.694
Ca	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
K	1.925	1.911	1.944	1.825	1.915	1.906	1.914	1.895	1.107	1.929
Na	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
	1.925	1.911	1.944	1.825	1.915	1.906	1.914	1.895	1.107	1.929
Cl	.000	.000	.000	.000	.000	.000	.019	.000	.000	.000
	.000	.000	.000	.000	.000	.000	.019	.000	.000	.000
Mah.	Mah.	Mah.	Mah.	Mah.	Mah.	Mah.	Nhl.	Nhl.	Nhl.	Nhl.

FeO expressed as total Fe

TOT-O is O attributed to chlorine

APPENDIX 2

SEM ANALYSES FOR BIOTITES OF THE STUDY AREA

	Nhl.	Nhl.	Mah.	Mah.	Mah.	Mah.	Mah.	Mah.	Hla.	Hla.
	28-39# 5	28-39# 6	28-27# 1	28-27# 2	28-27# 3	28-27# 4	28-27# 5	28-27# 6	28-13# 1	28-13# 2
SiO ₂	32.71	34.30	36.43	35.00	35.80	36.12	36.78	36.30	36.08	35.86
Al ₂ O ₃	14.83	14.66	14.78	14.97	15.20	15.09	15.20	14.88	15.47	15.21
TiO ₂	2.36	2.28	3.51	3.08	3.48	3.35	3.54	2.98	2.82	3.44
CR ₂ O ₃	.00	.00	.00	.16	.00	.00	.00	.00	.00	.00
FeO	31.05	28.21	19.99	20.17	20.31	19.69	19.90	19.91	25.18	26.64
MnO	.48	.49	.00	.00	.00	.00	.00	.00	.18	.32
MgO	4.22	3.95	11.01	11.46	10.74	10.80	11.02	11.38	5.57	5.03
CaO	.00	.00	.00	.00	.00	.00	.00	.00	.00	.00
K ₂ O	6.83	3.35	8.53	7.53	8.42	8.83	6.35	8.59	9.66	9.57
Na ₂ O	.00	.00	.00	.00	.00	.00	.64	.00	.00	.00
CL	.00	.00	.00	.00	.00	.00	.00	.06	.06	.09
TOTAL	93.18	94.24	94.30	92.57	93.05	93.73	95.33	94.10	96.02	96.16
TOT-O	95.18	94.24	94.30	92.57	93.95	93.73	95.93	94.09	96.01	96.14

FORMULA (BASIS 22 OXYGENS)

Si	5.400	5.577	5.615	5.497	5.545	5.595	5.576	5.605	5.658	5.638
Al	2.600	2.423	2.385	2.503	2.455	2.407	2.424	2.395	2.342	2.362
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	.285	.387	.296	.270	.319	.348	.232	.313	.518	.456
Ti	.293	.401	.406	.354	.405	.390	.404	.347	.333	.407
CR	.000	.000	.000	.019	.000	.000	.000	.000	.000	.000
Fe	4.288	3.836	2.573	2.650	2.630	2.538	2.523	2.570	3.434	3.503
Mn	.067	.068	.000	.000	.000	.000	.000	.000	.025	.043
Mg	1.211	.956	2.527	2.685	2.479	2.492	2.492	2.619	1.303	1.178
	6.144	5.648	5.802	5.988	5.835	5.768	5.711	5.849	5.613	5.587
Ca	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
K	1.438	1.939	1.676	1.509	1.663	1.745	1.712	1.692	1.932	1.920
Na	.000	.000	.000	.000	.000	.000	.187	.000	.000	.000
	1.438	1.939	1.676	1.509	1.663	1.745	1.899	1.692	1.932	1.920
CL	.000	.000	.000	.000	.000	.000	.000	.015	.015	.024
	.000	.000	.000	.000	.000	.000	.000	.015	.015	.024

Fe expressed as total Fe

TOT-O is O attributed to chlorine

APPENDIX 2

SEM ANALYSES FOR BIOTITES OF THE STUDY AREA

	28-138 3	28-138 4	28-138 5	28-138 6	83-09	83-09	28-3381	83-09	28-27
SiO2	35.03	35.34	35.32	35.51	33.57	32.32	23.74	26.89	46.40
Al2O3	14.45	14.91	14.49	14.35	15.37	16.00	20.33	16.91	34.05
TiO2	2.75	2.60	3.67	3.98	3.60	3.04	.00	.22	.00
CR2O3	.00	.00	.00	.00	.00	.00	.00	.11	.00
FeO	26.31	26.59	27.40	27.48	28.43	29.61	39.64	35.24	.00
MnO	.30	.31	.21	.35	.27	.32	1.00	.60	.00
NgO	5.39	5.37	5.06	5.16	5.55	5.96	3.86	7.58	.00
CaO	.00	.00	.00	.00	.00	.00	.00	.09	17.29
K2O	9.19	9.41	9.46	9.49	6.26	5.34	.12	.17	.00
Na2O	.00	.00	.00	.00	.00	.00	.00	.00	1.43
Cl	.31	.08	.09	.11	.00	.00	.00	.00	.00
TOTAL	93.71	94.61	95.70	96.33	93.05	92.59	88.69	87.81	99.17
TOT-O	93.64	94.59	95.68	96.51	93.05	92.59	88.69	87.81	99.17

FORMULA (BASIS 22 OXYGENS)

SI	5.673	5.656	5.612	5.612	5.431	5.273	4.237	4.710	5.903
AL	2.327	2.344	2.388	2.388	2.569	2.727	3.763	3.290	2.097
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
AL	.431	.469	.326	.277	.361	.350	.515	.200	3.009
TI	.332	.313	.439	.472	.438	.374	.000	.029	.000
CR	.000	.090	.000	.000	.000	.000	.000	.015	.000
FE	3.563	3.560	3.640	3.622	3.847	4.041	5.917	5.163	.000
MN	.042	.042	.029	.047	.037	.044	.151	.089	.000
NG	1.500	1.281	1.198	1.211	1.339	1.450	1.027	1.978	.000
	5.668	5.665	5.632	5.629	6.022	6.259	7.610	7.474	3.009
CA	.000	.000	.000	.000	.000	.000	.000	.016	2.358
K	1.898	1.921	1.917	1.908	1.291	1.112	.028	.037	.000
NA	.000	.000	.000	.000	.000	.000	.000	.000	.353
	1.898	1.921	1.917	1.908	1.291	1.112	.028	.053	2.711
CL	.086	.022	.024	.029	.000	.000	.000	.000	.000
	.086	.022	.024	.029	.000	.000	.000	.000	.000
Hla.	Hla.	Hla.	Hla.	Kar.	Kar.	Hla.	Kar.	Mah.	

Fe expressed as total Fe

TOT-O is O attributed to chlorine

APPENDIX 2

SEM ANALYSES FOR GARNETS OF THE STUDY AREA

	28-35# 7	28-35# 8	28-27# 7	28-27# 8
SiO ₂	37.07	37.16	37.55	37.68
Al ₂ O ₃	20.51	20.63	21.21	21.54
TiO ₂	.00	.00	.00	.00
Cr ₂ O ₃	.00	.00	.00	.11
FeO	32.55	32.51	33.64	33.23
MnO	1.94	2.62	.62	.55
MgO	.53	.78	4.21	4.39
CaO	7.15	7.15	2.71	2.84
TOTAL	100.95	100.95	99.54	100.42

FORMULA (BASIS 24 OXYGENS)

Si	5.004	5.991	5.990	5.970
Al	.000	.019	.010	.030
	6.004	6.000	6.000	6.000
Al	3.916	3.994	3.979	3.992
Ti	.000	.000	.000	.000
Cr	.000	.000	.000	.014
	3.916	3.994	3.979	4.006
Fe	4.410	4.389	4.488	4.404
Mn	.266	.357	.084	.084
Mg	.201	.188	1.001	1.036
Ca	1.241	1.233	.464	.482
	6.118	6.167	6.037	6.006

Fe expressed as total Fe

The above are from two samples of the Mahamba Gneiss

APPENDIX 2

SEM Oxide analyses for Mahamba rock samples

	Pyroxenite			Phyllite		Diabase		Px	Diabase
	#28-14 1	#28-14 2	#28-14 4	#28-15 1	#28-15 4	#28-18 2	#28-31 2	#28-14 3	#28-31 1
TiO ₂	.19	.15	.16	98.49	.26	96.21	4.04	51.01	47.55
Al ₂ O ₃	46.93	44.76	44.27	.26	33.89	.00	.00	.36	.00
Cr ₂ O ₃	14.58	16.24	17.00	.60	22.93	.00	.23	.27	.00
Fe ₂ O ₃	5.38	6.12	6.53	.00	6.74	.00	59.32	3.23	7.90
FeO	21.02	20.64	21.01	.28	33.39	.22	34.06	41.27	39.97
MnO	11.88	11.54	11.39	.00	1.48	.00	.00	1.79	.00
MgO	.00	.00	.23	.00	.50	.00	.00	1.42	2.74
NiO	.54	.71	.48	.00	.00	.00	.00	.00	.00
ZnO	.00	.41	.49	.00	1.90	.00	.00	.00	.00
TOTAL	100.52	100.57	101.56	99.63	101.09	96.43	97.65	99.35	98.16

FORMULA (BASIS 32 OXYGENS)

FORMULA (BASIS 6 OXYGENS)

Ti	.032	.026	.027	15.850	.049	15.980	.955	1.923	1.847
Al	12.437	11.984	11.797	.065	10.059	.000	.000	.022	.000
Cr	2.592	2.917	3.039	.101	4.565	.000	.057	.011	.000
Fe+3	.910	1.047	1.111	.000	1.278	.000	14.033	.122	.307
	15.971	15.974	15.974	16.016	15.951	15.980	15.045	2.078	2.154
Fe+2	3.952	3.921	3.973	.050	7.032	.040	8.955	1.730	1.726
Mg	3.980	3.908	3.838	.000	.557	.000	.000	.134	.000
Mn	.000	.000	.044	.000	.106	.000	.000	.060	.120
Ni	.097	.130	.088	.000	.000	.000	.000	.000	.000
Zn	.000	.069	.081	.000	.354	.000	.000	.000	.000
	8.029	8.028	8.024	.050	8.049	.040	8.955	1.924	1.846

Chromite

Chromite

chromite

rutile

chromite

rutile

magnetite

Ilmenite

Ilmenite

APPENDIX 2.1

APPENDIX 2.1

Compositional charts from SEM analyses

APPENDIX 2.2

Compositional charts from SEM analyses

APPENDIX 2.3

APPENDIX 2.3

Compositional Charts from SEM analyses

APPENDIX 2.4

Compositional charts from SEM analyses

APPENDIX 2.5

APPENDIX 2.5

Compositional charts from SEM analyses

APPENDIX 2.6

Compositional charts from SEM analyses

APPENDIX 2.7

APPENDIX 2.7

Compositional charts from SEM analyses

APPENDIX 2.8

APPENDIX 2.8

1 Fe vs. Fe/Fe+Mg 28/35
Mahamba Gneiss

2 Fe vs. Fe/Fe+Mg 28/39
Nhlangano Gneiss

3 Fe vs. Fe/Fe+Mg 28/27
Mahamba Gneiss

4 Fe vs. Fe/Fe+Mg 28/13 & B3-09
Hlatikulu Granite + Karoo

Compositional charts from SEM analyses

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