CHAPTER 1: INTRODUCTION.

1.1: Location of the Study Area:

The area investigated in this thesis is situated in the Northern Province of the Republic of South Africa (Figure 1.1). The study area lies approximately 100km north-west of Pietersburg and 120km W.S.W. of Louis Trichardt, and covers approximately 2200km^2. This area, and the scope of the study, was determined by the project sponsors. The northern border of the area under investigation is 23°S, the western border is 28°30'E, the southern border 23°25'S, and the eastern border forms a N.N.E.-trending diagonal line from 23°25'S - 28°55'E to 23°S - 29°10'E (Figure 1.2). Much of the study area is flat-lying, with a height, typically, of 900m above sea level. There are, however, two notable topographic features within the study area: the Makgabeng plateau, which covers c. 225km^2 in the south of the study area and averages 1200m in altitude, and Blouberg mountain in the north east of the study area, which rises to 2051m (Figure 1.2). Whilst the flat-lying portions of the study area are readily accessible by gravel or sand roads, access to the more mountainous areas can only be gained on foot.

Flat, relatively low-lying areas of the study area are generally covered by Quaternary Kalahari sand or calcrete. Outcrop quality is considerably better around Blouberg mountain and on the Makgabeng Plateau, and consequently these areas form the main focus of this study.

1.2: Regional Geology:

Within the study area, outcrops can be assigned to five different lithological units. The *basement rocks* are comprised of Archaean granulite-grade gneisses of the Limpopo Mobile Belt, which are overlain by a series of younger, generally non-metamorphosed volcano-sedimentary and sedimentary Proterozoic successions: the *Blouberg Formation*, the *Waterberg Group* and the *Soutpansberg Group*. Some strata of the Phanerozoic *Karoo Supergroup* also occur locally, but the extent of their outcrop is minor. Before considering the detailed geology of the strata within Blouberg mountain and its surrounding area (Chapters 2,3,4,5, and 6), this section will provide an outline of the general characteristics of the lithological units which are most important to this study. During this chapter, the existing stratigraphic names and classifications used by the South African Committee for Stratigraphy (S.A.C.S., 1980) will be retained. However, in later chapters, evidence will be proposed for a revised stratigraphic nomenclature, which will be used throughout the remainder of this work.

1.2.1: The Limpopo Mobile Belt:

The 250 km–wide, E.N.E.–W.S.W. -trending Limpopo Mobile Belt is thought (Treloar *et al.*, 1992) to represent a Himalayan-style collisional event between the Kaapvaal Craton in the south and the Zimbabwe Craton in the north (Figure 1.3). Within the Mobile Belt, a central zone and two marginal zones can be identified by their individual geological signatures (van Reenen *et al.*, 1992). The Southern Marginal Zone (S.M.Z.) is composed of granite-greenstone material from the Kaapvaal Craton, metamorphosed under at granulite facies conditions. It is separated from the remainder of the Kaapvaal Craton by a northward-dipping shear zone, which contains down-dip lineations, and is thought to have formed by southward-verging thrusting of the Kaapvaal Craton during syn-collisional crustal thickening (van Reenen *et al.*, 1992). The Central Zone (C.Z.) of the Mobile belt is characterised by granulite-grade gneiss of the Beitbridge Complex (S.A.C.S., 1980). Generally the Beitbridge Complex is composed of quartzo-feldspathic gneiss, with highly altered meta-sediments such as metapelitic gneiss, marble, quartzite and magnetite quartzite. Mafic protoliths are represented by amphibolites (van Reenen, 1992). The boundary between the S.M.Z. and the C.Z. is marked by the Palala Shear Zone, which is composed of mylonitised C.Z. and Bushveld Complex (c. 2.05 Ga) rocks. Structures within the 10 km-wide Palala Shear Zone suggest sinistral strike-slip displacement (McCourt and Vearncombe, 1987). Similarly, the Northern Marginal Zone (N.M.Z.) is separated from the Central Zone by the Tuli-Sabi Shear Zone, which is interpreted as a gently-dipping dextral strike-slip fault (McCourt and Vearncombe, 1987). The N.M.Z. is separated from the Zimbabwe Craton by a southward-dipping, northwardverging thrust, and is composed of granulite-grade rocks, the protoliths of which are thought to have been granite-greenstone rocks of the southern edge of the Zimbabwe Craton. Thus the Limpopo Mobile Belt appears to be symmetrical in structure (Figure 1.3).

Although the structure of the Limpopo Mobile Belt is relatively well understood, there are conflicting interpretations concerning the age of the orogeny. Until recently, it has generally been held that the minimum age of the collision could be gained by dating the crystallisation of granites which cross-cut tectonic fabrics within the belt, and are thus thought to represent decompression melting during post-collisional exhumation. These chronological data suggest that the age of the contractional, late-stage of the orogeny is 2.65 Ga (McCourt *et al.*, 1995).

However, more recently it has been suggested that the age of the Limpopo orogeny may be considerably younger (e.g. Barton *et al.*, 1994; Holzer *et al.*, 1998). Kröner *et al.* (1999) found that granites throughout the Central Zone could be dated as having crystallised at a variety of ages between 3.3 Ga. and 2.5 Ga., all of which show fabrics indicating subsequent polyphase deformation. This, therefore, indicates that much of the tectonic history of the Central Zone took place in the early Proterozoic, rather than in the late Archaean. Undeformed granites in the Central Zone were dated, with an age of approximately 2.0 Ga being proposed (Barton *et al.*, 1994; Holzer *et al.*, 1998; Kröner *et al.*, 1999;). The older dates, including the 2.65 Ga event, are regarded as being related to earlier tectonic events but not to the cratonic collision. However it is argued (Treloar and Blenkinsop, 1995; McCourt and Armstrong, 1998) that such young dates (c. 2.0 Ga) may reflect a later phase of metamorphism, caused, for example, by the proximal intrusion of the c. 2.06 Ga Bushveld Complex in the Kaapvaal Craton, by peripheral mobile events (e.g. the c. 2.0 Ga Magondi orogeny on the northwestern edge of the Zimbabwe Craton, or the coeval -Kheis orogeny, south-west margin of the Kaapvaal Craton), or by the c. 2.0 Ga (Therriault et al., 1997) Vredefort impact event in the central Kaapvaal Craton. It is also argued that structures such as the Palala Shear Zone (which cuts the Bushveld

Complex) do not necessarily represent sutures between the zones of the Limpopo Belt, but may have formed during later events (McCourt and Armstrong, 1998).

1.2.2: The Blouberg Formation:

The Blouberg Formation is regarded (Jansen, 1976) as a siliciclastic and volcanic succession, which lies nonconformably on the gneiss of the Limpopo Mobile Belt, which is unconformably overlain by medial to upper formations of the Waterberg and Soutpansberg Groups (Jansen, 1976, 1982; Brandl, 1986). The extent of the outcrop is small, and occurs in isolated exposures approximately on the projected line of the Palala Shear Zone through the study area. The Blouberg Formation can therefore be considered to outcrop on the suture between the northern edge of the Southern Marginal Zone and the southern edge of the Central Zone. There is little evidence for metamorphism or mylonitisation of the Blouberg strata, suggesting that they post-date the Limpopo orogeny (Jansen, 1976,) and movement on the Palala Shear Zone (Callaghan and Brandl, 1991). The lithostratigraphic subdivision of the Blouberg Formation (S.A.C.S., 1980) is presented in Table 1.1, and the location of the outcrops of these constituent members is shown in Figure1.4.

Jansen (1976, 1982) considered the various members of the Blouberg Formation to have been laid down within partially isolated basins within a block-fault zone. This accounts for the variety of lithologies mapped within the Blouberg area and their isolated pattern of outcrops (Figure 1.4), and for the fact that the Blouberg Formation cannot be recognised as a continuous succession (Jansen, 1982). The lowermost, feldspathic clastic members of the Blouberg Formation generally outcrop in the southern foothills of Blouberg mountain, whereas non-feldspathic clastic sediments and volcanic members outcrop generally to the north and west of Blouberg mountain (Figure 1.4) (Jansen, 1976).

Jansen (1976, 1982) correlated the feldspathic Blouberg members (Basehla Arkose Breccia, Mananka Arkose, Thalalane Feldspathic Sandstone and Mmallebogos Grit members) with the middle Waterberg strata (in particular the Setlaole Formation (Section 1.2.3) correlates well with the feldspathic Blouberg members). The non-feldspathic and volcanic Blouberg members (Mositone Conglomerate, Varedig Sandstone, Semaoko Grit and My Darling Trachyandesite members; Table 1.1) were correlated by Jansen (1976, 1982) with the late Waterberg (in particular the Mogalakwena Formation (Section 1.2.3) correlates well with the non-feldspathic members). Callaghan and Brandl (1991) suggested that the non-feldspathic and volcanic members of the Blouberg Formation should rather be considered as a part of the Soutpansberg Group (Section 1.2.4).

1.2.3: The Waterberg Group:

The Waterberg Group is comprised dominantly of red coloured, coarse clastic strata, and outcrops mainly in the west of the Northern Province, and in eastern Botswana, with smaller outcrops in Gauteng and Mpumalanga Provinces, South Africa (Figure 1.5). The Waterberg Group is composed of twelve formations, some of which grade laterally into each other (Table 1.2). The formations are preserved in two basins, the Middelburg basin and the Warmbaths basin; the latter is sometimes considered to be made up of a Main basin, and the southern Nylstroom protobasin (Jansen, 1982; Callaghan *et al.*, 1991) (Figure 1.5). The thickness of the Group may locally be as much as 7km (du Plessis, 1987), though other workers (e.g. Cheney and Twist, 1986) argue that the thickness is less, up to 5km. A thickness of 5km or less is also supported by geophysical investigations (Stettler, 1991). The age of the Group is estimated as 1900-1700 Ma (Jansen, 1982), though no radiometric dating has been done on Waterberg lithologies, and ages are largely based on relationships with surrounding, dated lithologies (Barker *et al.*, in press).

The Waterberg Group (Warmbaths basin) is considered to have been deposited within a continental fault-bounded basin, which developed in the northern part of the Kaapvaal Craton and the southern edge of the Limpopo Mobile Belt. Important faults which appear to have controlled the location of the edge of the basin are the Murchison Fault Zone (part of the Thabazimbi-Murchison Lineament) in the south and the Melinda Fault Zone in the north (Callaghan *et al.*, 1991) (Section 1.2.5; Figure 1.5). Generally, the basin seems to have been filled by prograding sedimentary systems from south to north, as successively younger formations outcrop towards the Melinda Fault Zone (Jansen, 1976), although northern source areas are important locally.

The red colour of the Waterberg sediments is produced by haematite grain coatings, which appear to be early diagenetic in origin, as coatings themselves are deformed by late-diagenetic compaction, and are often covered by late diagenetic quartz overgrowths (Eriksson and Vos, 1979). This suggests that the sediments were deposited within an oxidizing environment, within 30° of the equator (Turner, 1980).

The Swaershoek, Wilge River and the lowermost part of the Sterk River formations (Figure 1.6) are thought to have been deposited in fan deltas and alluvial fans, which prograded into lakes, where they were subsequently reworked by lacustrine tides (Jansen, 1982; Callaghan *et al.*, 1991). The upper portion of the Sterk River Formation and the Alma Formation reflect synsedimentary fault activity along the Murchison Fault Zone, producing a steep fault scarp at the southern margin of the basin. North of the scarp, alluvial fans prograded northwards, composed of detritus eroding from the newly formed highlands to the south (Callaghan *et al.*, 1991). Fine detritus was washed further north where it either settled in lakes or was deposited in the distal parts of prograding lacustrine fan deltas (Callaghan *et al.*, 1991).

The Skilpadkop Formation (Figure 1.6) is interpreted as having been deposited in braided rivers. Local soft-sedimentary overturning and slumping within the Formation are thought to reflect either dragging by heavily laden, fast-flowing water above newly deposited sediment, or shock dewatering cased by activation of basin-bounding faults (Callaghan *et al.*, 1991). The Setlaole Formation (Figure 1.6) has also been interpreted as reflecting proximal braided river deposition, based on the immature clast-supported sedimentary rocks and a lack of planar bedding (Callaghan *et al.*, 1991.). The Setlaole Formation also locally contains tuffaceous beds. Thus the palaeoenvironmental conditions inferred for the Setlaole and the laterally equivalent (Table 1.2) Skilpadkop Formation are very similar (Callaghan *et al.*, 1991).

The Aasvoëlkop Formation (Figure 1.6) is generally upward-coarsening, and is interpreted to have been deposited within a lacustrine environment, with coarse fluvial detritus, washed into the lake, becoming increasingly common towards the top of the Formation (Callaghan *et al.*, 1991). The base of the Aasvoëlkop Formation is marked by a lahar deposit (Callaghan *et al.*, 1991), indicative of volcanic activity within the Waterberg basin at this time, possibly associated with faulting at the margins of the depository. The fluvial upper part of the formation exhibits contorted bedding, indicating movement along a nearby (Murchison) fault zone (Callaghan *et al.*, 1991).

The Makgabeng Formation (Figure 1.6) is laterally equivalent to the Aasvoëlkop Formation, and outcrops in the northern part of the Waterberg basin. It is characterised by very large-scale cross-bedding, and uniformly-sized fine- to medium-grained sandstones, lacking clay matrix (Meinster and Tickell, 1975). Also, local tabular bodies of massive sandstone are reported (Callaghan, 1987a). These facies associations are thought to reflect an aeolian palaeoenvironment. The multimodal palaeocurrent pattern, derived from the dip direction of the large cross-beds reflects superposed longitudinal dunes deposited during changing seasonal wind directions (Callaghan, 1987b); alternatively, they probably also reflect barchanoid dune forms (Simpson *et al.*, 1999). Small scale aqueous, or partly aqueous structures observed in the Makgabeng Formation, such as ripplemarks, interference ripplemarks, adhesion warts and desiccation cracks, may indicate gradational change to the lacustrine conditions of the Aasvoëlkop Formation to the south (Callaghan *et al.*, 1991), or could reflect interdune deposition (Eriksson *et al.*, 2000).

Both the Sandriviersberg and the Mogalakwena Formations (Figure 1.6) comprise trough and planar cross-bedded arenites, though the Mogalakwena Formation is more coarse, suggesting that it is more proximal. Both formations contain common pebble washes, with boulder conglomerates present in the most northerly outcrops of the Mogalakwena

Formation (Jansen, 1976). The average palaeocurrent direction for the two formations is from E.N.E. to W.S.W., whilst grain size generally decreases towards the south. (De Bruiyn, 1971). Both formations are interpreted to have been deposited in a braided stream palaeoenvironment; the consistent palaeocurrent directions, and the presence of matrix-poor coarse-grained sandstones with poorly rounded grains indicates that sedimentation was rapid, though not chaotic, and was deposited within a steadily deepening basin during uplift of a provenance to the north-east (Callaghan *et al.*, 1991).

The Cleremont Formation (Figure 1.6), characterised by texturally and mineralogically mature medium- to coarse-grained sandstones with local rounded pebble washes, is thought to have been deposited within a littoral palaeoenvironment (Callaghan *et al.*, 1991). The Vaalwater Formation (Figure 1.6) is very mature, suggesting that it may have been reworked from earlier Waterberg Group strata, and the greyish colour of these arenites indicates a less-oxidizing environment than the remainder of the Waterberg Group (Callaghan *et al.*, 1991). The postulated palaeoenvironment was a shallow siliciclastic sea, although the association of ripplemarks and trough cross-bedding may rather indicate estuarine conditions (Callaghan *et al.*, 1991). The Vaalwater Formation thus suggests fluctuating base levels within the late-Waterberg basin (Callaghan *et al.*, 1991).

The palaeoenvironmental interpretations suggested for the Waterberg Group reflect at least one major reactivation of the fault zones bounding the basin, as two fining-up sequences can be identified within the succession (Callaghan *et al.*, 1991), from the Swaershoek (basal) to the Aasvoëlkop/Makgabeng Formations, and from the Mogalakwena Formation to the Vaalwater Formation (Figure 1.6). The mid-Waterberg resumption of rapid erosion appears to coincide with extension of the basin to the north, and the increased supply of detritus from northerly source areas (Callaghan *et al.*,1991).

1.2.4: The Soutpansberg Group:

The volcano-sedimentary Soutpansberg Group outcrops in the far north of South Africa, mainly in the Soutpansberg mountains. The mountains form a long south-facing escarpment from Kruger National Park in the east to Vivo in the west. The Soutpansberg Group is preserved in an elongated basin, which extends from the western end of the present study area to Punda Maria (Figure 1.7). The lithostratigraphic subdivision of the Soutpansberg Group is shown in Table 1.3, and the distribution of the units in Figure 1.7. Generally, the Soutpansberg strata have a moderate to steep northerly dip, and are cut by several E.N.E.-W.S.W.-trending faults (van Eeden *et al.,* 1955).

The basal Tshifhefhe Formation is only locally developed at the eastern end of the Soutpansberg basin, and is only a few metres thick. It is comprised of strongly epidotised clastic sediments, including shale, greywacke and locally-derived conglomerate (Barker *et al.*, in press). The Sibasa Formation comprises subaerially extruded basalt, with intercalated pyroclastic and sandstone lenses. Generally the basalts are massive, epidotised and locally amygdaloidal (Barker *et al.*, in press). The pyroclastic lenses locally reach a thickness of 200m, whereas the laterally persistent clastic lenses locally attain a thickness of 400m (Barker *et al.*, in press).

The generally siliciclastic Fundudzi Formation is only developed at the eastern end of the basin (Figure 1.7). It is mainly comprised of arenaceous and argillaceous sedimentary rocks, though there are rare pyroclastic horizons, and basaltic lavas are intercalated with the sedimentary lithologies close to the top of the Formation (Barker *et al.*, in press). The Wyllies Poort Formation is composed of red-pink quartzite with minor pebble washes. The base is marked by a prominent agate pebble conglomerate, and in the east, minor basaltic and pyroclastic intercalations are present (Barker *et al.*, in press). The uppermost unit of the Soutpansberg Group is the Nzhelele Formation, which is volcanic at the base (400m), followed by argillaceous sedimentary rocks in the middle, and arenaceous rocks at the top (Barker *et al.*, in press).

The preponderance of inferred fluvial sediments and subaerial lavas suggest that the Soutpansberg Group was deposited within a continental setting. Although originally no unconformities were identified between the formations (Jansen, 1974), more recent work (Cheney *et al.*, 1990) identified a regionally-developed, low-angle unconformity beneath the Wyllies Poort Formation. Figure 1.7 shows that the base of the Wyllies Poort Formation lies on successively older rocks towards the west, indicating that the Soutpansberg Group may not represent continuous continental deposition (Cheney *et al.*, 1990).

1.2.5: The Melinda Fault Zone:

The Melinda Fault Zone strikes E.N.E.-W.S.W. across the region. In the west, it bounds the northern edge of the Palala Shear Zone at 23°22'S; 27°57'E, and in the east it appears to merge with several sub-parallel faults cutting the Soutpansberg strata and Karoo strata east of Vivo (Figure 1.7). In the Blouberg area, the fault bifurcates into two strands, which pass to the north and south of Blouberg mountain (termed the *northern* and *southern* strands respectively in this work). The northern strand, in turn, bifurcates into several splays in the north-eastern foothills of Blouberg mountain, where it cuts the Wyllies Poort Formation (Geological Survey 1: 250000 sheets 2326; Ellisras, 2328; Pietersburg and 2228; Alldays). The fact that the Melinda Fault displaces strata in an area situated approximately above, and parallel to the projected line of the Palala Shear Zone, may imply that the Melinda Fault Zone is a manifestation of reactivation along the Palala Shear Zone. It is thought that the Melinda Fault comprises two distinct phases of movement. Pre-Karoo movement along the fault caused Waterberg strata to the south to be downthrown, with a vertical displacement of about 1000m, and reactivation in post-Karoo times caused a downthrow to the north, with a vertical displacement of at least several hundred metres (Brandl, 1986).

1.3: Previous work in the Blouberg area:

Throughout sections 1.3 and 1.4, previously published work which has bearing on the present study will be examined, and many of the conclusions of earlier workers will be highlighted. Although the Blouberg area in Northern Province has not received much attention from workers during the last century (Section 1.3), more work has been done in surrounding areas, and on related lithologies, which have bearing upon the strata within the Blouberg area (Section 1.4).

The most detailed published work directly concerning the geology of the area around Blouberg mountain was that of Jansen (1976), and the discussion of that paper by Meinster (1977). Jansen (1976) based his paper on field data obtained from the Blouberg area by B. Meinster and S.J. Tickell between 1972 and 1974. The following sections summarise the reasoning and conclusions of the work of Jansen and Meinster.

1.3.1: Summary of the work of Jansen (1976):

Jansen (1976) recognised that the Blouberg area was located at the intersection of two major sedimentary basins on the Kaapvaal Craton: the Waterberg basin, which developed progressively from south to north, and the Soutpansberg basin, which developed progressively from east to west. As such, the Blouberg area presented an ideal opportunity to consider the relative ages of the Waterberg and Soutpansberg Groups, as the Blouberg area provides the only location where these two groups of strata are developed in close proximity to each other. Additionally, Jansen (1976) identified that this location appeared to be coincident with the northern margin of the Kaapvaal Craton.

1.3.1.1: Stratigraphic units examined by Jansen (1976):

Jansen (1976) considered three stratigraphic units within the Blouberg area: The Blouberg Formation, the Waterberg Group (in particular the Setlaole, Makgabeng and Mogalakwena Formations), and the Soutpansberg Group (in particular the Wyllies Poort

Formation). His paper aimed principally at solving age-relationships between these three units, and examining their depositional conditions and structural patterns.

Jansen (1976) proposed that the Blouberg Formation consists of an association of eight members (Table 1.1). The lowermost four members, which outcrop on the southern slopes of Blouberg are generally feldspathic siliciclastic sediments (Basehla Arkose Breccia, Mananka Arkose, Thalalane Feldspathic Sandstone, Mmallebogos Grit Members). The northern slopes of Blouberg, however, are underlain by three non-feldspathic siliciclastic sedimentary units (Mositone Conglomerate, Varedig Sandstone, and Semaoko Grit Members) which are overlain locally by a volcanic unit (My Darling Trachyandesite Member).

He proposed that the members of the Blouberg Formation were laid down within an active block-faulted terrain along the uplifted northern rim of the Waterberg basin (the 'Limpopo Rise'). The variability in maturity, inferred variability in distance from source area, variance in depositional conditions, and the patchy occurrence of the outcrops of the Blouberg Formation sediments could thus be explained by the varying presence of localised basins, by deposition within partially isolated basins, or 'negative fault blocks' (Jansen, 1976). As block-faulting was continuous throughout the deposition of the Blouberg members, small depositories were created at different times and in different places along the southern edge of the Limpopo Rise (Jansen, 1976).

Within the Waterberg Group, Jansen (1976) identified three formations: The arkosic, partially conglomeratic Setlaole Grit Formation, the Makgabeng Sandstone Formation, (which wedges out rapidly to the north, so that it is not present within the Blouberg block-fault zone), and the Mogalakwena Conglomerate Formation, characterised by the locally developed basal Sesalong Boulder Conglomerate Member, overlain by conglomerate and sandstones.

Jansen (1976) assigned all Soutpansberg Group strata in the Blouberg area to the Wyllies Poort Formation, which is characterised by light-coloured quartzite, with locally developed pebble washes, conglomerates and mudstone pellets. The presence of interbedded lavas, tuffs, argillaceous rocks and diabase sills was only inferred by the presence of rubble, as outcrop quality is exceptionally poor. Rarely outcropping lavas are coarse-grained, epidotised and often amygdaloidal. Significantly, Jansen (1976) assigns none of the lavas in the Blouberg area to the volcanic Sibasa Formation of the Soutpansberg Group.

1.3.1.2: Stratigraphic relationships proposed by Jansen (1976):

The only direct stratigraphic relationships that Jansen (1976) was able to observe between the three units in the field were:

1.) The Wyllies Poort Formation (Soutpansberg Group) unconformably overlies the non-feldspathic members of the Blouberg Formation on the northern slopes of Blouberg, and also unconformably overlies the basement.

2.) The Mogalakwena Formation (Waterberg Group) unconformably overlies the feldspathic members of the Blouberg Formation on the southern slopes of Blouberg, and also the basement.

3.) The Blouberg Formation unconformably overlies the basement.

4.) The contacts between the Waterberg and Soutpansberg Groups are faulted.

With such incomplete data (especially regarding the age relationship between the Waterberg and the Soutpansberg Groups), Jansen (1976) attempted to invoke other methods in order to draw up a stratigraphic column. His conclusions regarding stratigraphic relationships were based on lithological similarities between strata, palaeogeographical data, and age determinations on the basement.

Jansen (1976) correlated the four lowermost (feldspathic) members of the Blouberg Formation with the Setlaole Formation of the Waterberg Group, based on the fact that the

Setlaole is also locally feldspathic and conglomeratic, and lithologically similar to some of the members in the lower Blouberg Formation. Jansen (1976) suggested that a correlation of Blouberg units with the Magkabeng Sandstone Formation (Waterberg Group), which is aeolian in origin (Meinster and Tickell, 1975), and which is generally an arenaceous mature sandstone, is untenable.

Jansen (1976) noted that the feldspathic members of the Blouberg Formation, on the southern slopes of Blouberg, are unconformably overlain by the basal boulder conglomerates (Sesalong Member) of the Mogalakwena Formation. However, to the west of Blouberg, on the farm Kranskop 278 LR (23°09'S; 28°42'E to 23°09'S; 28°41'E: Appendix 1), Jansen recorded the presence of an abnormally thick succession of Blouberg strata (1200m of the Basehla Arkose Breccia Member). This succession is unconformably overlain by the sandstone and grit of the Mogalakwena Formation, rather than being overlain by the lowermost boulder conglomerate that typically marks the unconformity on the southern slopes of Blouberg. Jansen (1976) therefore suggested that the anomalously thick Kranskop strata may correlate with the Setlaole, the Makgabeng, and the lowermost part of the Mogalakwena Formations.

Jansen (1976) argued a syn-Mogalakwena age for the non-feldspathic members of the Blouberg Formation. He suggested that the southerly dip of the older feldspathic members of the Blouberg Formation indicated an area of positive relief to the north, thus providing evidence against contemporaneous deposition of the feldspathic and non-feldspathic Blouberg members. The fact that the non-feldspathic members cannot be obeserved to be overlain by the basal Sesalong Conglomerate Member of the Mogalakwena Formation, and the similarity between the non-feldspathic members of the Blouberg Formation and the upper sandstone and grit of the Mogalakwena Formation led Jansen (1976) to suggest that they could be correlated. Jansen (1976) also suggested a syn-Sibasa age for these rocks, as an alternative, as the non-feldspathic members can be observed to be unconformably overlain by the Wyllies Poort Formation of the Soutpansberg Group.

Thus, Jansen (1976) concluded that the feldspathic members of the Blouberg Formation are of middle Waterberg age (syn-Setlaole) and the non-feldspathic members are of late Waterberg age (syn-Mogalakwena) and also contemporaneous with the volcanic Sibasa Formation of the Soutpansberg Group. The Wyllies Poort Formation was considered to be younger, thus providing evidence that the Soutpansberg Group strata represent deposition over a relatively long period of time.

1.3.1.3: Mode of deposition proposed by Jansen (1976):

With a complete, ordered stratigraphical column of the lithologies in the Blouberg area, Jansen (1976) examined the inferred mode of deposition and palaeogeography of the Blouberg area at the time of deposition. As previously mentioned, Jansen strongly favoured the deposition of the feldspathic members of the Blouberg Formation within a block-faulted terrain positioned parallel to, and immediately south of the 'Limpopo Rise'. This rise, an area of strongly positive relief, was envisaged to have formed the north-eastern edge of the Waterberg basin, and to have been the source of many of the sediments within the Blouberg area. Sedimentation of the feldspathic members within the small isolated basins was inferred to have resembled that of yoked basins, with thick arkose wedges. Areas where there is no outcrop of Blouberg strata were explained by localised rises within the palaeotopography, and the 1200m-thick succession of arenaceous and arkosic strata on Kranskop 273LR (23°09'S; 28°42'E to 23°09'S; 28°41'E: Appendix 1) was explained by the encroachment of the main Waterberg basin into the localised Kranskop basin during Setlaole deposition.

Similarly, the main Waterberg basin was envisaged to have encroached over the entire block-faulted terrain upon the onset of Mogalakwena deposition, as the Sesalong Conglomerate Member outcrops above the feldspathic Blouberg strata. Evidence for the continued activity of the block-faulted terrain beneath is given by the irregular thickness and distribution of the Sesalong conglomerate, and by localised tilting of this stratum. Continued erosion of the Limpopo Rise mountain belt led to its progressive retreat northwards as Mogalakwena Formation deposition continued. Thus the Mogolakwena rocks are less rich in feldspar, reflecting erosion of increasingly distant source areas. Current directions within the Mogalakwena Formation were interpreted by Jansen (1976) to reflect fluvial currents both longitudinal and transverse to the Limpopo Mobile Belt (E.N.E.-W.S.W.) trend , and indicate an overall transport from the N.E., from an area occupied by the most elevated terrain of the Limpopo Rise. The now denuded area west of Blouberg was subjected to localised downwarping, creating a depository for the non-feldspathic members of the Blouberg Formation (the 'Lebu trough'). Deposition in the Lebu trough was thought to have been contemporaneous with the deposition of the sandstone and grit of the upper-Mogalakwena to the south. Following peneplanation of the Limpopo Rise, down-faulting continued, localised rifting provided conduits for lavas (My Darling Trachyandesite Member), and eventually culminated in the development of the Soutpansberg trough, previously envisaged by Jansen (1975b) to represent an aulacogen. Current directions recorded within the Wyllies Poort Formation were generally north to south, and S.W. to N.E. within the east-west trending Soutpansberg trough. Jansen (1976) interpreted this as reflecting, in part, longtitudinal current directions within the trough, and hence indicating a partial reversal of the Waterberg current directions.

Having dealt with the primary aim of his work (that of solving the enigmatic stratigraphic relationships and mode of deposition within the Blouberg area), Jansen (1976) used this as a basis to speculate on the basin evolution and structural patterns in the Blouberg area.

1.3.1.4: Basin evolution of the Blouberg area proposed by Jansen (1976):

Jansen (1976) noted that, in comparison to the relatively undisturbed nature of the Waterberg and Soutpansberg strata, the feldspathic members of the Blouberg Formation are frequently steeply tilted, and locally overturned. He interpreted this deformation as representing the first post-Limpopo deformational event in the Blouberg area, and ascribed the deformation to localised down-faulting within the block-fault zone. This earliest deformational event is envisaged to have continued from mid-Waterberg (syn-

Blouberg/Setlaole Formations) until the late Waterberg (syn-Sesalong Conglomerate Member), as the basal Mogalakwena is also tilted locally.

Similar down-warping and faulting is thought to have formed Jansen's (1976) next deformational phase in the area west of Blouberg. This led to the widening of the fault zone, and ultimately to the creation of horsts and grabens, most notably that of the 'Lebu trough', into which the non-feldspathic Blouberg members were deposited, and in which the My Darling trachyandesitc lavas were erupted. These rocks are locally unconformably overlain on the southern margin of the Lebu trough by the Wyllies Poort Formation. The Lebu trough is considered by Jansen (1976) to be a proto-trough to the main Soutpansberg trough. Continued regional down-warping and down-faulting culminating in the creation of the main Soutpansberg trough is the third structural event proposed by Jansen (1976) for the tectonic evolution of the Blouberg area. The developing Soutpansberg trough extended farther south than the Lebu trough, into the earlier Blouberg block-fault zone, now occupied by overturned feldspathic Blouberg members and sub-horizontal, locally-dipping Mogalakwena strata. Continued reactivation of the trough-bounding faults was envisaged to have occurred while regional down-warping of the Soutpansberg aulacogen proceeded. On account of this, the strata close to the trough-bounding faults are locally steeply tilted; the Blouberg strata bordering the faults on the southern edge of the trough became overturned, while Wyllies Poort quartzite on the northern margin attained dips up to 80° (Jansen, 1976).

1.3.1.5: Conclusions (Jansen, 1976):

On a cratonic scale, Jansen's model for the geological evolution of the Blouberg area can be considered in terms of the extension of a rift zone (Soutpansberg trough) into the domain of a cratonic basin (Waterberg Group), thus creating an intersection of two zones of crustal weakness. In turn, the development of the Soutpansberg and Waterberg basins was probably controlled by crustal events within the Limpopo Mobile Belt. Mobilisation within the Limpopo Mobile Belt between 2100-1800 Ma (the youngest U-Pb zircon ages to have been recorded in the basement in the Blouberg area at that time) led to isostatic uplift and the deveopment of a crustal arch over the Limpopo Mobile Belt (the Limpopo Rise). This was the main source of sediment for the developing Waterberg basin to the south. The isostatic uplift of the Limpopo Belt is envisaged to have promoted the sub-crustal flow towards it, and encouraged both the creation of the Blouberg block-fault terrain, and the northwards propagation of the Waterberg basin towards the Limpopo Belt. The Soutpansberg trough was presumed to have formed as a result of the foundering of the crustal arch, producing volcanic centres in the eastern Soutpansberg (Sibasa basalt) and at Blouberg (My Darling trachyandesites).

Jansen ended his work by proposing that the structural features examined should not be considered in terms of plate tectonics, as he was of the opinion that the Limpopo Mobile Belt was non-collisional in origin. Jansen (1976) rather favoured an intra-plate environment as the setting for the Waterberg and Soutpansberg basin development.

1.3.2: The work of Meinster (1977):

Meinster, who had performed much of the initial mapping on which Jansen (1976) had based his work, did not agree with many of Jansen's (1976) proposed age relationships. Rather than consider the Blouberg Formation as containing eight individual members, Meinster (1977) proposed that the non-feldspathic members be termed the 'Lebu Complex', and the feldspathic members be termed the 'Blouberg Complex', and Jansen's (1976) individual members be considered as formations.

Meinster (1977) recognised clasts of the Varedig Formation (Lebu Complex) within the Basehla Arkose Breccia Formation (Blouberg Complex). Additionally, the presence of well-rounded quartz pebbles within the otherwise immature Blouberg Complex was thought to be due to reworking of primary conglomerates within the Lebu Complex (e.g. Semaoko and Mositone Formations), and soft green material within the Blouberg Complex was thought to be highly weathered lava from the My Darling Formation. Thus, Meinster (1977) suggested that the Blouberg Complex was younger than the Lebu Complex, reversing the stratigraphy of Jansen (1976). Meinster (1977) did not look favourably upon the idea of block-faulting being a potential cause for overturning of the Blouberg Complex, suggesting that overturning caused by drag folding is a rarity, whilst overturned beds in the Blouberg and Lebu Complexes are common. Thus, he argued for a major deformational phase, postulating that steep upthrusts and overthrusts caused folding and overturning in the Lebu Complex, and created localised downwarping for the Blouberg Complex depository. Meinster (1977), however, was not able to show field evidence for such thrusts. The uplifted Lebu Complex was rapidly eroded into the nearby depository, accounting for the immaturity of the Blouberg Complex sediments, and the presence of the Lebu Complex detritus therein (Meinster, 1977). As compressional tectonics continued with time, the Blouberg Complex itself became tilted and overturned. Peneplanation of both the Lebu and Blouberg complexes following compression is regarded as having removed most of the evidence for such a deformational event. Significantly, Jansen (1977), in a reply to Meinster, argued that the Lebu Complex (non-feldspathic Blouberg members) was not particularly folded or tilted compared to the strata of the Blouberg Complex.

With regard to other correlations made by Jansen (1976), Meinster (1977) questioned the suitability of correlating the Blouberg Complex with the Setlaole Formation of the Waterberg Group, based solely on the presence of feldspars in both units. Correlation based upon such a common sedimentary mineral as feldspar was not, Meinster argued, sound practice. Additionally, with regard to this correlation, Meinster brought attention to the point that the geological map of the Blouberg area provided by Jansen (1976) shows Blouberg Complex strata unconformably overlain by Setlaole Formation rocks at 'Hill 3970' on the farm Beauley 260 LR (23 06.80'S; 28 59.40'E: Appendix 1). Meinster also noted the relative lack of deformation in the Setlaole Formation compared with the Blouberg Complex, and the lack of mudstones in the Setlaole, whilst observing that they are present (although rare) in the Blouberg Complex. Meinster (1977) did, however, propose that the Lebu Complex might correlate with the sediments of the Koedoesrand Complex in the Palala Shear Zone (located about 100km W.S.W. of the Blouberg area), as these rocks also show good evidence for having been affected by a strong phase of deformation. The fact that the Koedoesrand strata are intruded by the Bushveld Complex (c.2050 Ma) provided Meinster with evidence that the Lebu and Blouberg complexes were considerably older than the ages proposed by Jansen (1976).

The erosion and peneplanation of the Lebu and Blouberg Complex mountains was followed by deposition of what Meinster (1977), somewhat ambiguously, terms 'transitional strata' (they are not marked on his map), reported as flat-lying strata between the Lebu Complex and the Soutpansberg Group. These transitional strata appear on Jansen's (1976) map as part of the Varedig Member.

Of critical importance regarding the age-relationships in the Blouberg area, are Meinster's (1977) opinions regarding the relative ages of the Soutpansberg and Waterberg Groups. Meinster (1977) proposed that the similarity between the Lebu Complex and the Soutpansberg Group in terms of inferred mode of deposition, cyclicity, sedimentary rocks, lavas and their close spatial relationship, indicate their close relationship in time. Meinster (1977) refers to them as 'a single entity....separated from each other by a strong compressional phase, vigorous erosion, and subsequent deposition of the Blouberg Complex. Additionally, as the Soutpansberg strata are regarded as having been preserved, rather than having been deposited, in grabens, Meinster (1977) argues that rifting related to the graben formation would have produced plateau lavas over any pre-existing strata. The lack of plateau lavas within the Waterberg Group shows that these strata had not been deposited at this point. Therefore the logical succession of strata in the Blouberg area according to Meinster (1977) is:

5. Karoo Sequence
4. Waterberg Group
3. Soutpansberg Group
2. Blouberg Complex
1. Lebu Complex

With regard to the age of the Soutpansberg Group, Meinster refers to an age of lavas in the upper portion of the Group dated at ±2000 Ma, indicating a much earlier age for the

Soutpansberg Group than the Waterberg Group. Meinster (1977) argued that this age could be re-set, e.g. by the Bushveld intrusion, at 2050Ma from an even earlier date, and hints that the Soutpansberg Group may, in fact, correlate well with the Ventersdorp Supergroup. Such an early age is considered unlikely by Jansen (1977) in a reply to Meinster, due to the fact that the Soutpansberg sediments are generally red beds, thus reflecting sedimentation and diagenesis in an oxidising environment and thus within a younger time period (post c. 2000 Ma)(Jansen, 1977).

The contrasting stratigraphic and structural relationships proposed by Jansen (1976) and Meinster (1977), derived from the same field data, serve to illustrate the structural complexity and the lack of good stratigraphic markers across the study area, which has hampered previous work within the Blouberg region.

1.3.3: The work of Brandl (1991):

The contrasting views on issues concerning the geological history of the Blouberg area remained largely dormant until Brandl (1991) briefly re-examined the outcrops. He suggested that the non-feldspathic and volcanic strata within the Blouberg Formation/Complex (i.e. the Lebu Complex of Meinster, 1977) should rather be considered as part of the Soutpansberg Group. The My Darling Trachyandesite Member would therefore correlate with the Sibasa Formation (Brandl, 1986; Cheney *et al.*, 1990). The Blouberg Formation would thus only consist of feldspathic strata, and Brandl (1991) proposed that these should not be classified as separate members.

1.4: Relevant previous work in surrounding areas:

1.4.1: Previous work on the Limpopo Mobile Belt:

The Limpopo Mobile Belt has been subjected to intense study during the last sixty years, although early European miners began exploiting copper at Messina as early as 1904 (Barton, 1983). Söhnge (1945) began to map the rocks around Messina in 1940, and was the first to consider the high-grade metamorphism of those lithologies (Söhnge, 1940; Söhnge *et al.*, 1948). MacGregor (1953) coined the term 'Limpopo Orogeny', having identified that the metamorphic rocks formed a belt. Holmes and Cohen (1957) determined an approximate age of 2 Ga for the Limpopo Orogeny. Cox et al. (1965) termed it the 'Limpopo Orogenic Belt', and were the first to identify the Northern, Central, and Southern zones within the Belt, which terminology still remains in usage. Anhaeusser *et al.* (1969) applied the name 'Limpopo Mobile Belt' in order to avoid analogies with Alpine tectonic models. Bahnemann (1971) recognised basement and supracrustal successions within the Messina area, with deformed mafic dykes also present in the basement rocks.

Mason (1973) examined the Limpopo Belt as a whole, and envisaged the belt as being "a zone of crustal weakness throughout geological time and characterised by repeated shear deformation, igneous intrusion and extrusion". He recognised that the zones of the Limpopo Mobile Belt were separated from each other by shear zones (e.g. the Palala Shear Zone), and also proposed a faulted relationship between the Southern Marginal Zone and the Kaapvaal Craton. The Northern Marginal Zone was, however, believed to grade transitionally into the Zimbabwe Craton. In contrast, Coward *et al.* (1973) identified the northern margin as a dextral shear zone with up to 200km of displacement, which was also the view of Hepworth (1977).

Between 1973 and 1982, The South African Council for Scientific and Industrial Research (C.S.I.R.) sponsored the Limpopo Working Group as part of the National Geodynamics Project (N.G.P.) A summary of the work of the Limpopo Working Group is presented in Table 1.4. In 1983, the Geological Society of South Africa produced a special publication (van Biljon and Legg, 1983), which drew together much of the work of the Limpopo Working Group. The work in this publication is summarised in Table 1.5. and more recent work (1983 to 1999) is summarised in Table 1.6. The results of previous work on the Limpopo Mobile Belt which are most relevant to this study, were presented in the section on general geology (Section 1.2.1) and are not repeated here.

1.4.2: Previous work on the Waterberg Group:

1.4.2.1: Early work on the Waterberg Group (1872-1982):

Very early work on the Waterberg Group (1872-1965) is summarised in Table 1.7. The most relevant previous work began in the mid-sixties, when the Geological Survey of South Africa began a comprehensive study of the Waterberg strata. Much of the work in this large-scale study also concerned the investigation of the Blouberg Formation (Jansen, 1976; Meinster, 1977) and the Soutpansberg Group (Jansen, 1975b) which are dealt with separately in Section 1.3 and Section 1.4.3. The more recent work by the Geological Survey of South Africa (1965-1982) is summarised in Table 1.8, where publications which are of particular importance to the present study are highlighted in bold type. The work of Meinster and Tickell (1975), and Tickell (1975) are particularly relevant to the present study area.

Meinster and Tickell (1975) investigated the Makgabeng Formation on the Makgabeng Plateau. They reported the ubiquitous presence of fine- to medium-grained sandstone with rounded to well-rounded, spherical sand grains, with little or no matrix. Of particular importance are the sedimentary structures present within the Formation. Meinster and Tickell (1975) reported that the Formation is characterised by pronounced fine-scale laminations within very large-scale cross-bedded units. The large cross-bedded units are between 2 and 10m in set thickness, and their maximum recorded extent was recorded as 200m along foreset strike, and 400m perpendicular to strike. Meinster and Tickell (1975) also noted that the lower bounding surface of the cross-bedded units is typically marked by enigmatic 20cm to 8m-thick massive sandstone beds. Generally, they found that bounding surfaces between cross-bedded units were curved, though these were also observed to be irregular on a small scale, indicating erosion prior to deposition of the next cross-bedded unit. Foresets were reported to be generally tabular or shaped like very broad troughs. In cross-section, the foresets are concave, and typically dip at around 19°, though a maximum dip of 38° was recorded (Meinster and Tickell, 1975). The foresets flatten and merge towards the bottom of the set, so that the bottomsets are confomable with the massive beds beneath. Sparse ripple marks were reported on bottomset and foreset beds within the Makgabeng Formation. These are generally asymmetric, though many show truncated tops. Parting lineations were observed on the tops of massive beds. Rare mudstone lenses, up to 2 or 3 metres in width and only a few centimetres thick, were also recorded by these two researchers in their 1975 paper.

Meinster and Tickell (1975) considered, but rejected a sub-aqueous origin for the Makgabeng sediments, and favoured an aeolian palaeoenvironment. Massive beds were interpreted as having formed during temporary changes in wind direction, causing the dune to become aerodynamically unstable, and to be rapidly reworked. Such an interpretation was strengthened by the presence of anglar unconfomities between the massive bed and the earlier cross-bedded dune beneath, and the contrasting dip directions of subsequent dunes, reflecting changing wind directions (Meinster and Tickell, 1975). The presence of parting lineations on the top surface of the massive beds is not easily reconciled with the aeolian model of Meinster and Tickell (1975). They suggested that parting lineations may be able to form in subaerial conditions, due to rotating cellular vortices of air, in a similar way to parting lineations that are produced in fast-flowing water. The generally tabular shapes of the foresets in the Makgabeng Formation were taken to indicate that cross-bedded units represented fossil transverse dunes, and broad trough-shaped foresets were thought to be barchan dunes. Palaeowind directions, inferred from foreset dip directions, were thought to be dominantly from the N.E, (Meinster and Tickell, 1975).

Tickell (1975) examined the generally horizontally-dipping coarse sandstone and conglomerate of the Mogalakwena Formation. He found that the Mogalakwena Formation conformably overlies the Makgabeng Formation, with a 20m thick zone of transitional rocks developed between the two formations. However, Tickell (1975) also noted that in the immediate area of the Melinda Fault (Figure 1.5) there is a disconformity developed between the two formations. In the south-west, the Mogalakwena Formation is reported to grade laterally into the Sandriviersberg Formation, which is generally more yellowish in colour than the Mogolakwena, and contains no conglomerates. The maximum preserved thickness of the Mogalakwena Formation was reported to be 1250-1500m.

The Mogalakwena Formation is generally comprised of coarse sandstone, with several subordinate conglomerate members locally developed at a variety of stratigraphic heights (Tickell, 1975). The Tafelkop member is developed just beneath the base of the overlying Cleremont Formation, the Marken Member is developed about 50m above the base of the Mogalakwena Formation, and the Sesalong Member, at Blouberg, lies directly at the base. All of the conglomerate members were estimated to be about 100m in thickness, and Tickell (1975) reported that the Sesalong Member wedges out about 10km south of Blouberg.

The sandstone member of the Mogalakwena Formation was observed to be purplish brown in colour with pale pink patches, and with the presence of cross-bedding being highlighted by dark laminae within the foresets. Tickell (1975) reported an average composition of 64% detrital quartz, 33% matrix, and 3% lithic fragments, with feldspars generally being absent. The conglomerates were reported to contain rounded to well-rounded clasts of high sphericity, composed mainly of vein quartz, quartzite and quartz-mica schist. Clasts of quartz-haematite schist (banded iron formation pebbles), jasper and fuchsitic quartzite were found to be less common, and clast diameters between 3 and 10cm, and locally up to 80cm were recorded. The matrix in the conglomerate is similar to the sandtone member (Tickell, 1975).

Trough cross-bedding was reported to be present throughout the Formation, up to 2m in width, 60cm in set thickness, and up to 4m long. The dimensions of cross-bedded units were found to vary considerably within individual outcrops. Cosets were reported to be up to 3m in thickness, and were followed laterally by Tickell (1975) for up to 1km. The base of each unit was observed to be irregular, although the development of channels was not noted. Tickell used trough cross-bedding to define palaeocurrent directions, and suggests an overall direction from north-east to south-west, though trends in the Steilloopbrug (40km south-west of Blouberg mountain) area are more westerly. The conglomerate members were observed to consist of laterally persistent, sheet-like pebble layers, up to 3m in thickness, interbedded with sandstones. It was noted that there is a regional decrease in clast size in the Sesalong member, with thick boulder conglomerates developed in the extreme north (in the Blouberg area), decreasing to cobble conglomerates 10-30km to the south-west. The palaeocurrent directions derived from trough cross-bedding present in the interbedded sandstone in the conglomerate members are indistinguishable from those recorded in the sandstone member (Tickell, 1975).

The nature of the trough cross-bedded units, coupled with the lack of ripple marks and fine argillaceous deposits in the Formation led Tickell (1975) to suggest that the Mogalakwena Formation had been deposited in a braided river environment. Tickell (1975) pointed out that the Mogalakwena Formation bore considerable similarity to molasse sediments deposited adjacent to the Himalyas. The northerly source of the sediments led Tickell (1975) to propose that the Mogalakwena developed as a molasse complex from the Limpopo mountains to the north.

A summary of the re-investigation of the Waterberg Group (1965-1982) by the Geological Survey of South Africa was provided by Jansen (1982), which highlights the key issues raised during this period of investigation. The results of this large-scale project greatly improved knowledge of the Group, and formed the basis on which modern work on the various Waterberg strata was undertaken.

1.4.2.2: Recent work on the Waterberg Group (1977 to present):

Other work continued on the Waterberg Group during this time that was independent of the Geological Survey. Coertze *et al.* (1977) proposed that sedimentation of the Transvaal (Supergroup, c. 2.6 – 2.1 Ga) succession and the Waterberg succession may have been at least partially continuous in the Otse basin of Botswana, and that sedimentation was interrupted only by as little as 180 Ma during the intrusion of the Bushveld Complex in the main Transvaal and Waterberg basins. Vos and Eriksson (1977) proposed a fluvial fan model for the depositional environment of the Waterberg Group, and suggested that the general reddish colour of these rocks was due to authigenic-diagenetic alteration of iron-bearing detrital silicates.

Cheney and Twist (1986) re-interpreted the stratigraphy of the Waterberg Group, proposing the presence of unconformities between each of the major formations in the lower part of the main basin. Only the Mogalakwena, Cleremont and Vaalwater formations were viewed as a continuous succession. Importantly, unconformities were thought to be present between the Setlaole and Makgabeng Formations, and between the Makgabeng and Mogalakwena Formations. This proposal was based on the fact that each of the five identified unconformity-bounded sequences (U.B.S.) rests on one or more older sequences, and also on basement rocks, indicating erosion between each U.B.S. The consequences of these proposed unconformity-bounded sequences within the Waterberg are that the present outcrop of Waterberg may bear little resemblance to the previous extent of the basin (i.e. the present outcrop is preservational rather than depositional). Models involving the northward migration of the depocentre (e.g. Jansen, 1975a) could no longer be applied accurately. Additionally, the presence of proto-basins could be questioned due to obscuring of the original basin architecture by inter-sequential erosion. Cheney and Twist (1986) proposed that the maximum thickness of the preserved Waterberg Group does not exceed 5km.

Cheney and Twist (1986) also provide good evidence constraining the age of some of the Waterberg strata: the Makgabeng Formation unconformably overlies the Abbotspoort Fault, a southern reactivation of the Palala Shear Zone. The Abbotspoort Fault cuts the Palala Granite, and deformation of the granite has been dated at 1770±60 Ma. (Cheney and Twist, 1986). This data indicates that the Makgabeng and younger formations probably post-date 1770±60 Ma, assuming that deformation of granite was coeval with the Abbotspoort Fault.

Du Plessis (1987) re-examined the Gatkop area of the Waterberg basin, east of Thabazimbi. He observed structures present along the Thabazimbi-Murchison Lineament (T.M.L.), a zone of major crustal weakness, that is thought to have actively controlled basin margins on the Kaapvaal Craton throughout the Late Archaean and much of the Proterozoic. The structures under investigation in the T.M.L. cut Bushveld granite, close to the edge of the Nylstroom protobasin (Figure 1.5), indicating that the T.M.L. may have been active at the time of Waterberg Group sedimentation, producing the Nylstroom protobasin and Alma trough as a pull-apart basin. Du Plessis argued that the unconformities present within the Waterberg Group need not necessarily represent basin-wide erosion, as basin edges are typically associated with marginal unconformities which grade into conformable relationships towards the centre of the basin. If a basin is expanding, each successively younger sedimentary succession will onlap onto the basin floor. Du Plessis therefore postulated that the presence of an unconformity indicates only the proximity of a basin edge, and not basin-wide erosion. It was therefore proposed that the present extent of Waterberg rocks does indeed closely resemble the extent of the Waterberg depositional basin, and that protobasins can be recognised (du Plessis, 1987); as a consequence, models of Waterberg basin evolution involving the northward migration of centres of thermal subsidence (e.g. Jansen, 1975a) were again proposed.

Stettler (1991) used gravitational and aeromagnetic data to determine the thickness of the Waterberg Group. The geophysical data suggested a maximum thickness no greater than 5km, in support of Cheney and Twist (1986). Callaghan (1987b) and Callaghan *et al.* (1991) summarised the sedimentology and petrography of the Waterberg Group strata, and investigated cassiterite–bearing placer deposits. Van der Neut *et al.* (1991) studied the Wilgerivier Formation in the Middelburg basin, and proposed that it had been deposited in a distal alluvial fan-braidplain environment within a graben. Eriksson *et al.* (1997) examined the economic potential of the Waterberg sediments, and discussed palaeoplacer deposits of titanomagnetite-ilmenite-zircon in the Cleremont Formation, cassiterite in the Gatkop area, and U-Cu and manganese deposits close to the T.M.L. Van der Neut and Eriksson (1999) calculated palaeohydrological parameters in the Wilgerivier Formation, and compared tham with calculated parameters from Phanerozoic braided river deposits in South Africa. The calculated palaeohydrological parameters were used to infer the palaeogeography and palaeoclimate. Barker *et al.* (in press) summarise the present knowledge concerning the Middelberg and Main Waterberg basins.

1.4.3: Previous work on the Soutpansberg Group:

Very early work concerning the Soutpansberg strata (1908-1955) is presented in Table 1.9. Original mapping of the Soutpansberg area was by Van Eeden *et al.* (1955), though the most comprehensive study of the Group is that of Barker (1979). Important work concerning the tectonic setting of the Soutpansberg Group has also been undertaken by Jansen (1975b), Barker, (1983) and Cheney *et al.* (1990).

Although the volcano-sedimentary nature of the succession is accepted by all workers, there has been heated debate in the past concerning the tectonic environment in which the Soutpansberg Group was deposited and preserved. Jansen (1975a) recognised that both the Waterberg and the Soutpansberg basins were intra-cratonic in origin (i.e. both are continental), but saw that there were considerable differences between the two basins. He appreciated that the Waterberg basin was developed in broad undulations in the craton, whereas the Soutpansberg basin appears to be developed within a distinctly narrow, E.N.E.-W.S.W. (Limpopo-parallel) trending trough, which may be partially fault bounded on the northern and southern margins. The apparent lack of internal unconformities within the Soutpansberg Group (Jansen, 1975b) led him to believe that the Soutpansberg Group represented a single sedimentary and volcanic cycle. Jansen (1975b) was also able to identify syn-sedimentary faulting, as some of the faults within the Soutpansberg Group are only developed in older formations.

Considering all the available data, Jansen (1975b) proposed that the Soutpansberg Group was deposited in an aulacogen (failed rift), which had developed due to reactivation between the Central and Southern Marginal Zones of the Limpopo Belt. The onset of rifting was marked by normal faulting (evidence for which is now masked by subsequent reactivation), and widespread volcanism (Fundudzi Formation) which continued sporadically throughout the lifespan of the trough. Jansen (1975b) believed that the older members of the Soutpansberg Group at the eastern end of the trough pre-dated the Waterberg Group, whereas relationships in the west at Blouberg suggested that the Wyllies Poort Formation was younger than the Waterberg Group (Jansen, 1975b, 1976) (Section 1.3.2). Jansen (1975b) therefore proposed that the aulacogen was long-lived (spanning the Waterberg Group sedimentation) and that the trough spread gradually from east to west through time. The fact that successively younger Soutpansberg strata appear to have been deposited non-conformably on basement strata towards the west seemed to corroborate this suggestion. The Soutpansberg aulacogen compared favourably with aulacogens developed on other cratons.

Jansen's (1975b) work did not remain unchallenged. Barker (1976) suggested that the aulacogen model should be regarded with caution. He noted the absence of chemical and marine clastic sediments within the Soutpansberg succession, providing evidence against aulacogen formation adjacent to a continental margin. Barker also noted that available palaeocurrent data indicated transport generally from the north, whilst aulacogen environments generally favour transport along the axis of the trough. Barker (1976) suggested that the evidence from the Soutpansberg Group favoured a rifted or yoked basin model, rather than an aulacogen. In addition to a rifted or yoked basin, Barker (1983) also proposed that the Soutpansberg Group may have been deposited with a half-graben with a southern hanging wall.

Meinster (1977) did not believe that the mature sediments of the Wyllies Poort Formation (typically recrystallised to quartzite with well rounded quartz pebble washes) could have formed within an active fault-bounded rift, and proposed that these sediments had been transported a considerable distance, rather than from the proximal edges of a graben. To account for such mature sediment, Meinster (1977) proposed that the Soutpansberg had been preserved, rather than having been deposited, in a graben type structure, hinting that the Soutpansberg Group may have been developed over a wider area than the present extent of outcrop suggested.

Such an idea was explored by Cheney *et al.* (1990), who identified a large regional low-angle unconformity below the Wyllies Poort Formation, which had remained unidentified by earlier workers. The presence of the unconformity was based on the observation from maps that the Wyllies Poort Formation rests on successively older strata towards the west, though this could rarely be identified directly in the field (the same relationship had been explained by Jansen (1975b) as being due to westward propagation of the Soutpansberg aulacogen). The presence of such an unconformity provided a strong argument against an aulacogen model, as continuous down-faulting should lead to continuous sedimentation. Major gaps in sedimentation (represented by the unconformity) suggested that the Soutpansberg Group was once, as Meinster (1977) earlier hinted, more widespread, possibly across the Limpopo, Zimbabwe and Kaapvaal provinces. Cheney *et al.* (1990) proposed correlations with other isolated red-bed sequences on the Kaapvaal and Zimbabwe Cratons that had previously been thought of as representing separate basins. The Olifantshoek Sequence in the west of the Kaapvaal Craton, and the Palapye Group in eastern Botswana were proposed as correlates of the Soutpansberg Group.

Barker (1983) considered the development of the Soutpansberg trough from the perspective of its evolution within the Limpopo Mobile Belt. He proposed that rising granite diapirs, formed during anatexis at about 2.7 Ga, were responsible for uplifting the Central Zone of the Limpopo Mobile Belt (granulite-grade gneisses are presently exposed at the surface, indicating around 15 km of uplift). Rapid uplift continued from 2.1 to 1.8 Ga, when isostatic settling along the margins of the Central Zone led to the creation of the Soutpansberg trough within an asymmetrical yoked basin, with a faulted northern boundary. Post -Karoo readjustment is proposed as the tectonic event responsible for the present-day steeply dipping Soutpansberg strata, erosion of which has created the escarpment in the Soutpansberg mountain range.

The geochemistry of the Sibasa and 'Ngwanedzi' (lower Nzhelele) Formation basalts was examined by Crow and Condie (1990). They found evidence that the magmas were derived from multiple mantle sources, which had been enriched in subduction-zone components during an earlier (Archaean) phase of arc collision. These conclusions were based on the ratios of incompatible trace-elements in the Soutpansberg basalts.

Barton and Pretorius (1997) related the Soutpansberg Group to rifting, possibly associated with mantle plumes. In agreement with Cheney *et al.* (1990), Barton and Pretorius (1997) also envisage the extent of the Soutpansberg Group to be much more widespread than the extent of the preserved basin suggests. Barton and Pretorius (1997) note that the age of the Soutpansberg Group (c. 1.85Ga) is coeval with the emplacement of the Schiel Alkaline Complex, just east of Louis Trichardt (Barton *et al.*, 1996).

Barker *et al.* (in press) propose that the most appropriate explanation for Soutpansberg basin evolution may be a combination of the models outlined above. They suggest that the lower Soutpansberg Group (Sibasa and Fundudzi Formations) represent a graben or rift environment. The upper Soutpansberg Group (Wyllies Poort and Nzhelele Formations), which is unconformable on the lower part of the Soutpansberg Group, was laid down within a more extensive basin, only a portion of which is now preserved, after reactivation of the original Soutpansberg rift (Barker *et al*, in press).

1.5: Aims of the Study:

Although the main basins of the Waterberg and Soutpansberg Groups have been the focus of considerable attention in recent years, the geology of the Blouberg area, (the only location where these two groups are found in close proximity to each other) remains poorly understood . The summary of previous work undertaken in the Blouberg area (Jansen, 1976; Meinster, 1977; Brandl, 1991; discussed in Section 1.3) serves to emphasise the lack of consensus of opinion regarding the Blouberg area. Fundamental geological questions, such as the age-relationships between the strata, remain largely unresolved.

Whilst the previously published data give a good overview of the geology of the Blouberg area, there is an absence of integrated analysis of sedimentary, structural and geochemical data from the study area. Advances in geological methodology in recent years present the opportunity for a new focus with which to re-examine the geological history of the Blouberg area. The present study therefore aims to collect new field data, including a detailed map of the area, structural data, measurement of sedimentary architectural elements and determination of their component lithofacies and their three-dimensional relationships, and to re-interpret the geological evolution of the Blouberg area.

Unravelling the tectonic and depositional history of the area around Blouberg mountain may help to constrain models for the age of the Limpopo Mobile Belt, which is presently the focus of much debate. Furthermore a model for the geological evolution of the Blouberg area may serve as a model for basin development and basin inversion on or near an inter-continental suture zone (or terrane boundary).

A second focus of attention of this work concerns the strata of the Makgabeng Formation on the Makgabeng plateau. An interpretation as an aeolian deposit has been proposed (Meinster and Tickell, 1975; Callaghan *et al.*, 1991), but the Formation has not been examined in detail. The considerable age of these sediments (c. 1920–1700 Ma; Jansen, 1982) indicates that the Makgabeng plateau may be one of the oldest (e.g., Eriksson and Simpson, 1998) and best preserved records of aeolian deposits.

1.6: Methodology:

1.6.1: Field work:

Field work was carried out between November 1997 and August 1999. The field area shown in Figure 1.2 was mainly mapped on foot, and geological data were recorded in notebooks and plotted on field base maps. Navigation in the field was generally accomplished by use of a G.P.S. (global positioning system) receiver.

General geological data were collected during the course of the fieldwork, and mostly comprised primary (sedimentary) structures and secondary (tectonic) structures. Primary sedimentological features were recorded for the identification of sedimentary facies, in order to attempt to correlate strata in different parts of the study area and characterise the depositional conditions. Directional data from primary structures such as ripple marks and cross-bedding orientation were recorded in order to infer palaeocurrent directions. The sedimentary lithofacies and palaeocurrent directions were considered together to produce a classification of architectural elements (Miall, 1996). Cross-bed set thickness, channel size and clast size were measured in order to calculate palaeohydrologic parameters, such as palaeoslope (Section 1.6.3).

Pebble surveys of conglomerates in the Mogalakwena Formation were undertaken with the use of a $1m^2$ grid, with vertical and horizontal strings stretched at 10cm intervals across the grid, thus producing 100 points of intersection. Counting the presence and composition of pebbles or cobbles at each of these points gave an estimate of the ratio of clasts to matrix, and the variability of clast composition within each m^2. Relative clast size could be estimated by the measurement of the b-axis (intermediate axis) of the largest clast within each m^2. Vertical variability of these parameters in the Mogalakwena Formation could be assessed by choosing accessible vertical profiles and conducting a metre by metre survey through the exposed vertical section. Lateral variation in these parameters could be assessed by comparison of data collected from vertical profiles from different locations.

Plane-table mapping was used for accurate large-scale maps of cross-bedding geometry in the Makgabeng Formation. Borehole data were logged in order to provide details on vertical lithological variation within the Waterberg Group. Dip and dip-direction of bedding were frequently recorded, and the orientation of tectonic structures (e.g. slickenside lineations, foliations, lineations and veins) was measured to aid with the reconstruction of tectonic events.

1.6.2: Photogeology:

A set of 1:50,000 –scale aerial photographs of the study area was used to assist with field-based mapping. The use of aerial photographs was most beneficial in mapping large areas west of the Makgabeng plateau, occupied by flat-lying Mogalakwena Formation sandstones and conglomerates. Outcrops of these lithologies could readily be identified from the surrounding Quarternary cover by stereoscopic examination of an overlapping pair of photographs. Dykes cutting the Waterberg Group could also be examined by the use of aerial photographs. As these basic igneous rocks weather much faster than the siliciclastic sedimentary wall rocks, they give rise to a reddish soil, which appears darker on monochrome aerial photographs. Dykes could therefore be easily identified, followed for many kilometres along strike, and plotted straight onto maps thereby avoiding unnecessarily long periods in the field.

1.6.3: Geological maps and cross-sections:

Collected field data and interpretetations from aerial photographs were transferred to 1:50,000 –scale topographic maps (South African Government Printer) of the study area, and data were then transferred to a digital format and plotted on digitised topographical maps, using ArcView G.I.S. (Geographic Information System) software. The final geological map is presented in Appendix 1, and cross-sections, based on interpretation of the maps in Appendix 1, are presented in Appendix 3.

1.6.4: Calculations of palaeohydrological parameters:

A variety of different formulae have been proposed by several workers with which to measure hydrological parameters in ancient fluvial deposits. Generally the field data required for these formulae are grain size analyses (sandstones) and clast size analyses (conglomerates and gravels) and the set thickness of cross-bedded units (e.g. Ethridge and Schumm, 1978). It is important to emphasise, however, that the calculated results are only estimates of the hydrological conditions at the time of deposition of the fluvial sediment. Use of different formulae will likely produce different results, and results may also vary considerably with a variety of data set sizes. It should therefore be stressed that the calculated results are most useful for comparative purposes between different fluvial deposits, and should not be considered as quantitative values for the palaeohydrological parameters being considered.

A variety of formulae will be used to calculate palaeohydrological parameters during this work, in order to minimise the error of using single formulae alone. Methods involve both utilisation of measurements from clast-filled channels, and the set thickness of cross-beds. The use of multiple methods also provides a range of parameters rather than a single value, which produces a more accurate framework from which to model the palaeogeography and palaeoclimate.

1.6.4.1: Palaeohydrological parameters that can be calculated from clast-filled channels:

Clast sizes can be used to calculate the average velocity of water and the unit stream power, according to the following equations:

$$v = 0.2 \times (di)^{0.455} \tag{1}$$

$$W = 0.009(di)^{1.686} \tag{2}$$

(where v is velocity of water in ms^{-1}, W is the unit stream power, measured in Watts/m^2, and di is the length of the intermediate axis of the largest visible clast) (Costa, 1983).

Provided that the cross-sectional surface area of a channel containing a clast can be measured (or estimated), the discharge can be calculated by the following equation:

$$Q = \bar{v} \times A \tag{3}$$

(where Q is the discharge in m^3s^{-1}(cumecs), and A is the cross-sectional surface area of the channel (approximated by $d_m \times w$) in m^2, where d_m is the mean depth of the channel measured in metres, and w is the width of the channel in metres)(Costa, 1983).

When the intermediate axis of a clast in the channel is measured, then palaeoslope can be calculated by the following equation:

$$W = \left(\frac{9800 \times Q \times S}{w}\right) \qquad (4)$$

(where S is the palaeoslope, measured in m/m) (Costa, 1983).

1.6.4.2: Palaeohydrological parameters that can be calculated from cross-bed set thickness:

The mean water depth can be calculated by the following formula:

$$h = 0.086(d_m)^{1.19} \qquad (5)$$

(where h is the mean set thickness of cross-beds in metres (Allen, 1968).

The ratio between channel width and depth can be estimated by:

$$F = 225M^{-1.08} \qquad (6)$$

(where F is the ratio between channel width and depth, and M is the sediment load variable, i.e. the percentage of silt and clay in the channel perimeter, which can be assumed to be a constant of 5% for such coarse bedload as inferred from the sedimentary rocks of the Blouberg area (Schumm, 1968a,b; Van der Neut and Eriksson, 1999)). This gives a fixed channel width to depth ratio, $F = 40$.

The estimation of the width of channel, w, can be calculated by:

$$w = Fd_m \qquad (7)$$

(Schumm, 1968a).

It is then possible to estimate the average daily discharge (also called the mean annual discharge by some workers):

$$Q_m = vA \qquad (8)$$

where Q_m is the average daily discharge, measured in m^3s^{-1}, and A is the mean cross-sectional surface area (approximated by $d_m \times w$) in m^2; v is the velocity of water in ms^{-1}, and can range between 0.5 and 1 ms^{-1} in conditions where large subaqueous dune bedforms migrate (i.e. when cross-bedding is formed). (Leopold *et al.*, 1964). For the purposes of this study, an intermediate velocity of 0.75 ms^{-1} is assumed.

Mean bankfull channel depth can be estimated by:

$$d_b = 0.6M^{0.34}Q_m^{0.29} \quad (9)$$

(where d_b is mean bankfull channel depth in m) (Schumm, 1969).

Calculation of bankfull channel width can be estimated by:

$$w_b = 8.9d_b^{1.40} \quad (10)$$

(where w_b is the bankfull channel width in m) (Leeder, 1973).

This allows for a recalculation of Q_m (average daily discharge), which can be compared to the results of equation 8, when the following equation is applied:

$$Q_m = 0.027w_b^{1.71} \quad (11)$$

(Osterkamp and Hedman, 1982).

Palaeoslope can be calculated by:

$$S = 60M^{-0.38}Q_m^{-0.32} \quad \text{(Schumm, 1968a)} \quad (12)$$

and by

$$S = 30\left(\frac{F^{0.95}}{w^{0.98}}\right) \quad \text{(Schumm, 1972)} \quad (13)$$

(where S is the palaeoslope).

Assuming that the estimation of Q_m is gained only from equation 11 (which is likely to be more accurate than that derived from equation 8), two different estimates of palaeoslope can be obtained from equations 12 and 13. Thus an approximate range of palaeoslope can be gained.

Using the two values of palaeoslope derived from equations 12 and 13, two estimates of bankfull water discharge can be made, using the equation:

$$Q_b = 4.0A_b^{1.21}S^{0.28} \tag{14}$$

(where Q_b is bankfull water discharge, and $A_b = d_b \times w_b$) (Williams, 1978).

The drainage area (catchment area) of a river system can be estimated by:

$$Q_b = A_d^{0.75} \tag{15}$$

(where A_d is the drainage area in km^2) (Leopold *et al.*, 1964).

Principle stream length (from source to depositional site) can be estimated by:

$$L = 1.4A_d^{0.6} \tag{16}$$

(where L is the stream length in km) (Leopold *et al.*, 1964).

By substituting the two different values of Q_b obtained from equations 12 and 13 into equations 15 and 16, two different values for drainage area and stream length can be calculated.

1.6.5: Laboratory methods

Lithological samples from all the main stratigraphic units encountered in the field area were collected for thin section analysis. Additional samples were collected for thin section analysis of sedimentary and tectonic structures, which required closer examination. Orientated samples were taken in order to determine kinematic directions, fault and recrystallisation history.

Orientation data of bedding and tectonic structures were analysed by stereographic projection in order to identify any dominant tectonic trends. Analyses of stereographic data were undertaken using *Spheristat 2* software. Each stereographic projection presented in this work is a lower hemisphere projection on a Schmidt (equal area) net. Palaeocurrent analyses and dyke trend analyses were plotted on rose diagrams, and vector mean (principal directions) were calculated, also using *Spheristat 2* software. Data concerning the trend of dykes intruding the field area were collected directly from collated maps. A 1.5cm^2 grid was placed over a 1: 150000 geological map, and the trend of each dyke within each grid square was measured with a protractor, and plotted on a rose diagram. Thus, the most laterally extensive dykes are recorded the most frequently within each grid square, and are in turn represented preferentially on the rose diagrams. The azimuth of the principal direction is indicated, where appropriate, on rose diagrams by a single line. The standard deviation away from the principal direction is indicated by small ticks on both sides of the principal direction line.

Geochemical investigation of Sibasa Formation basalts and dyke swarms intruding the Waterberg Group were undertaken by use of the I.C.P.M.S. facility at the University of Cape Town (incompatible trace elements) and by X.R.F. at the University of Pretoria (major and trace elements).

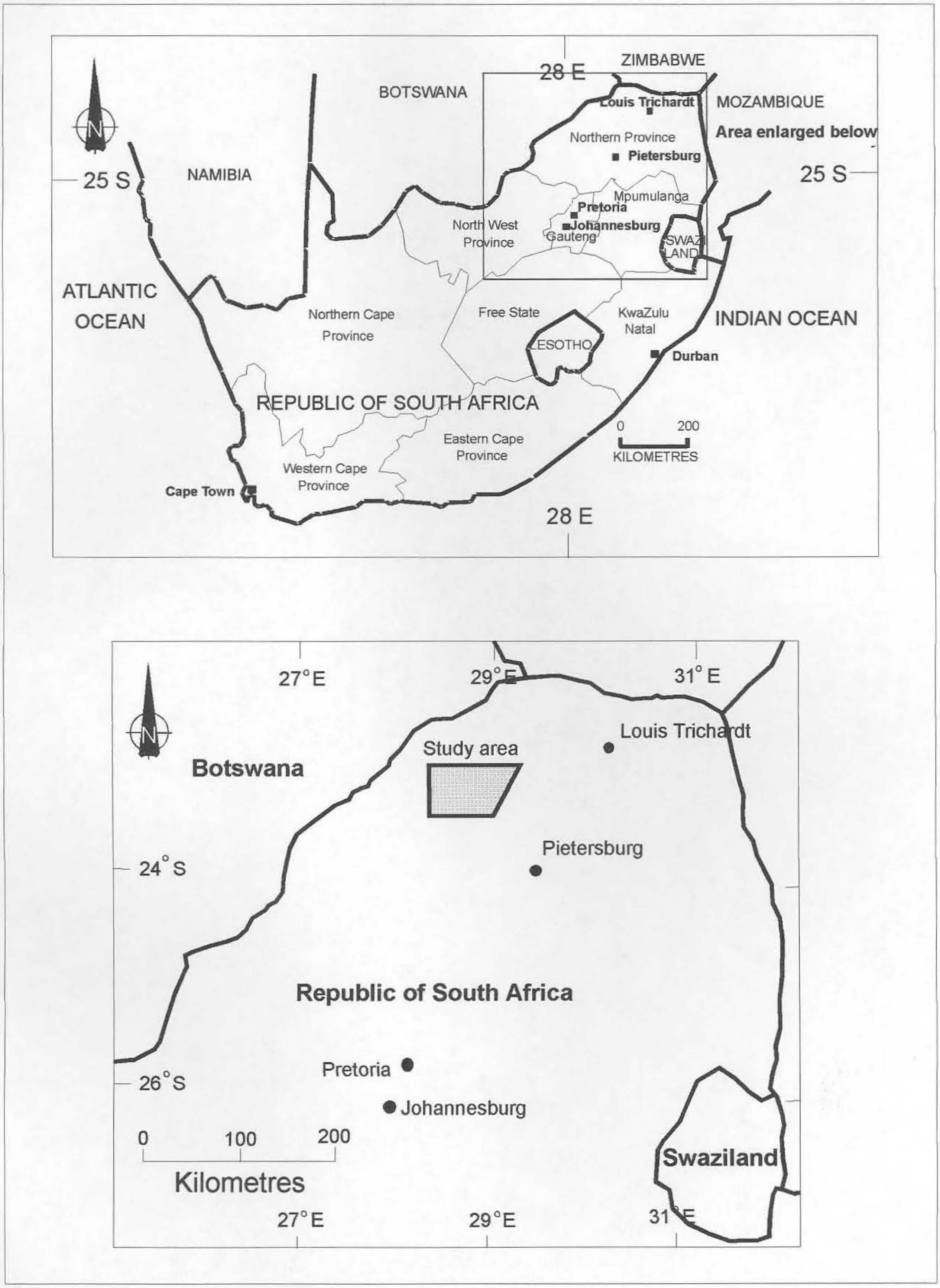

Figure 1.1: Geographical location of the study area.

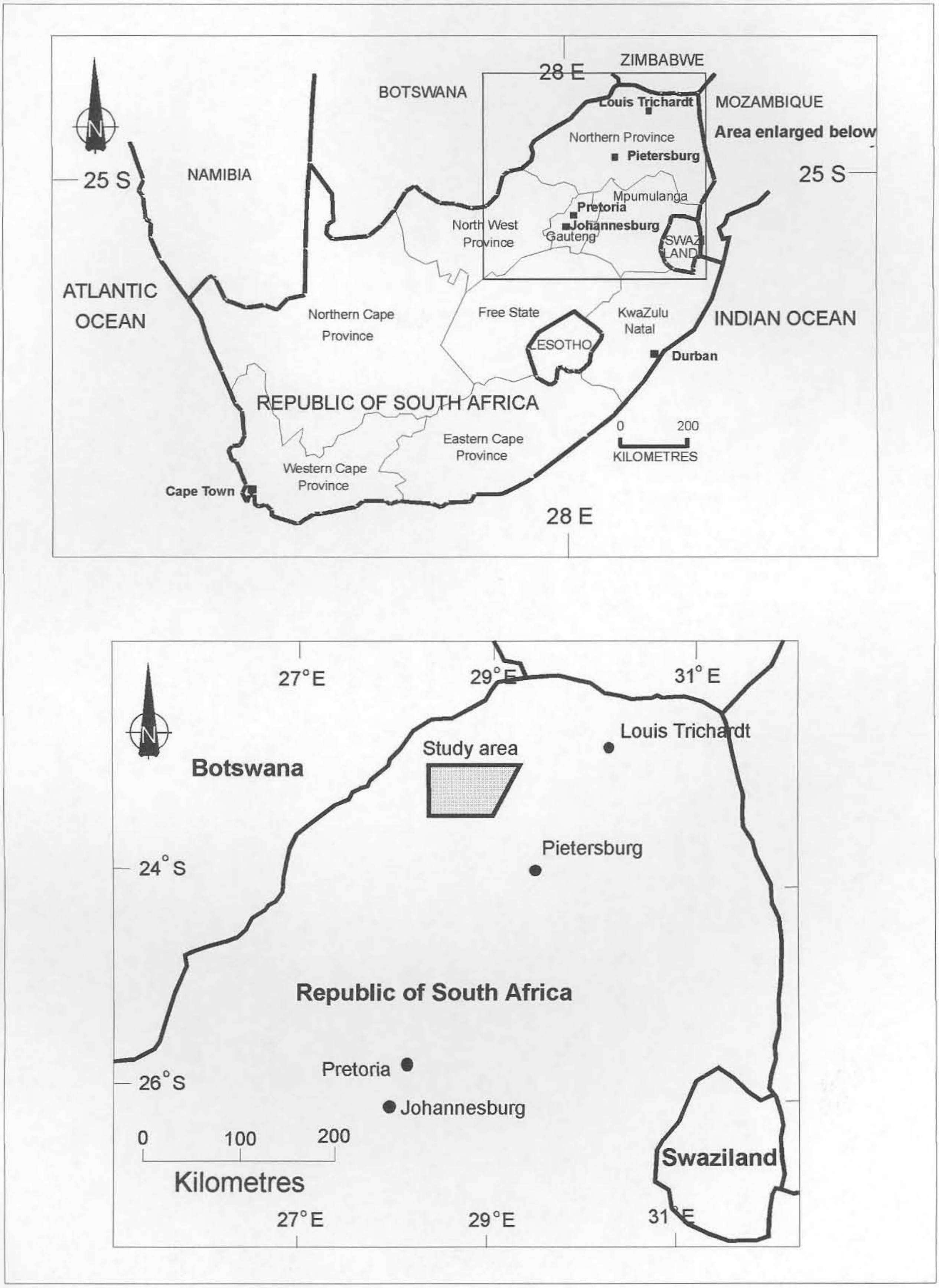

Figure 1.1: Geographical location of the study area.

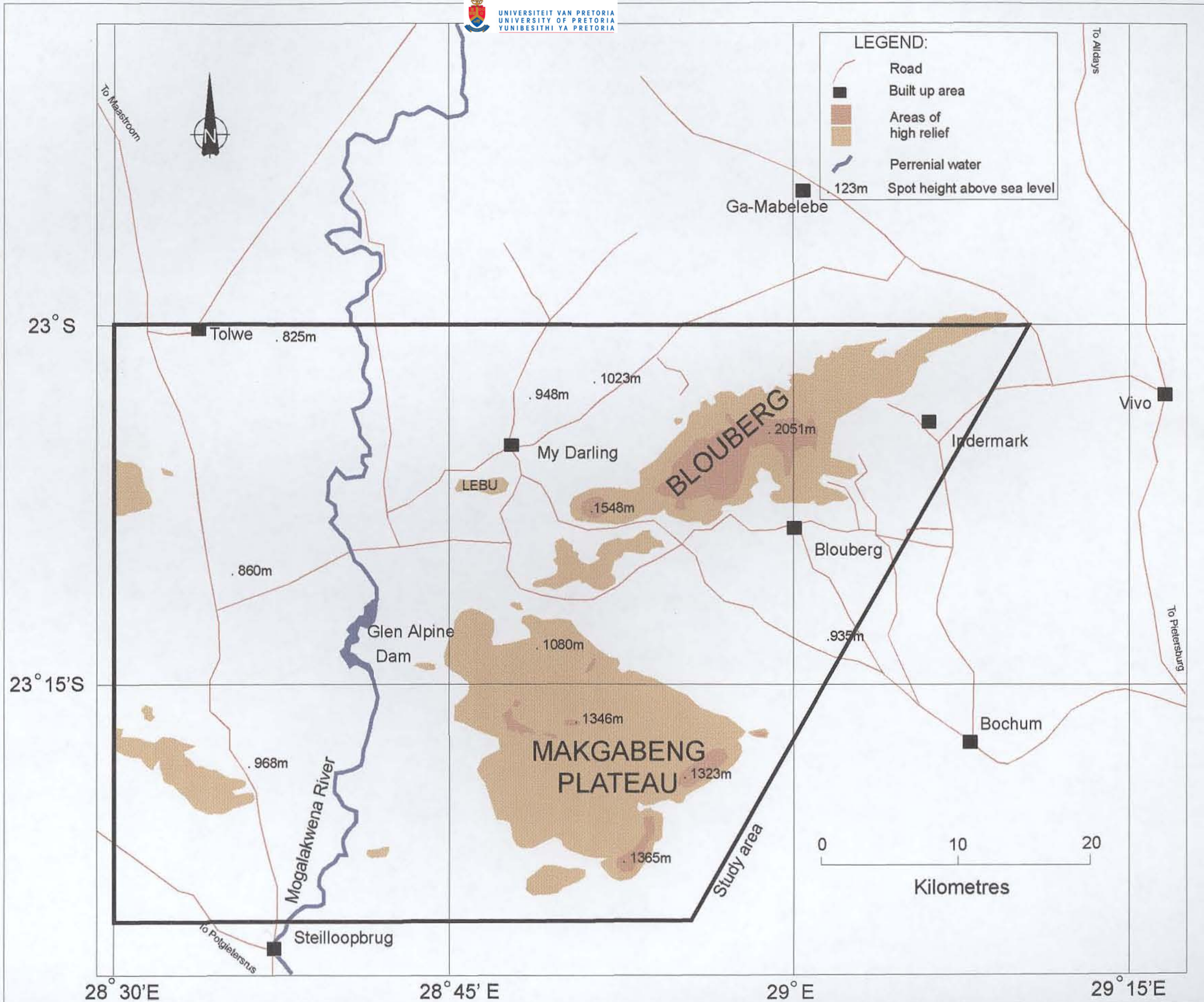

Figure 1.2: Map showing geographical features of the study area.

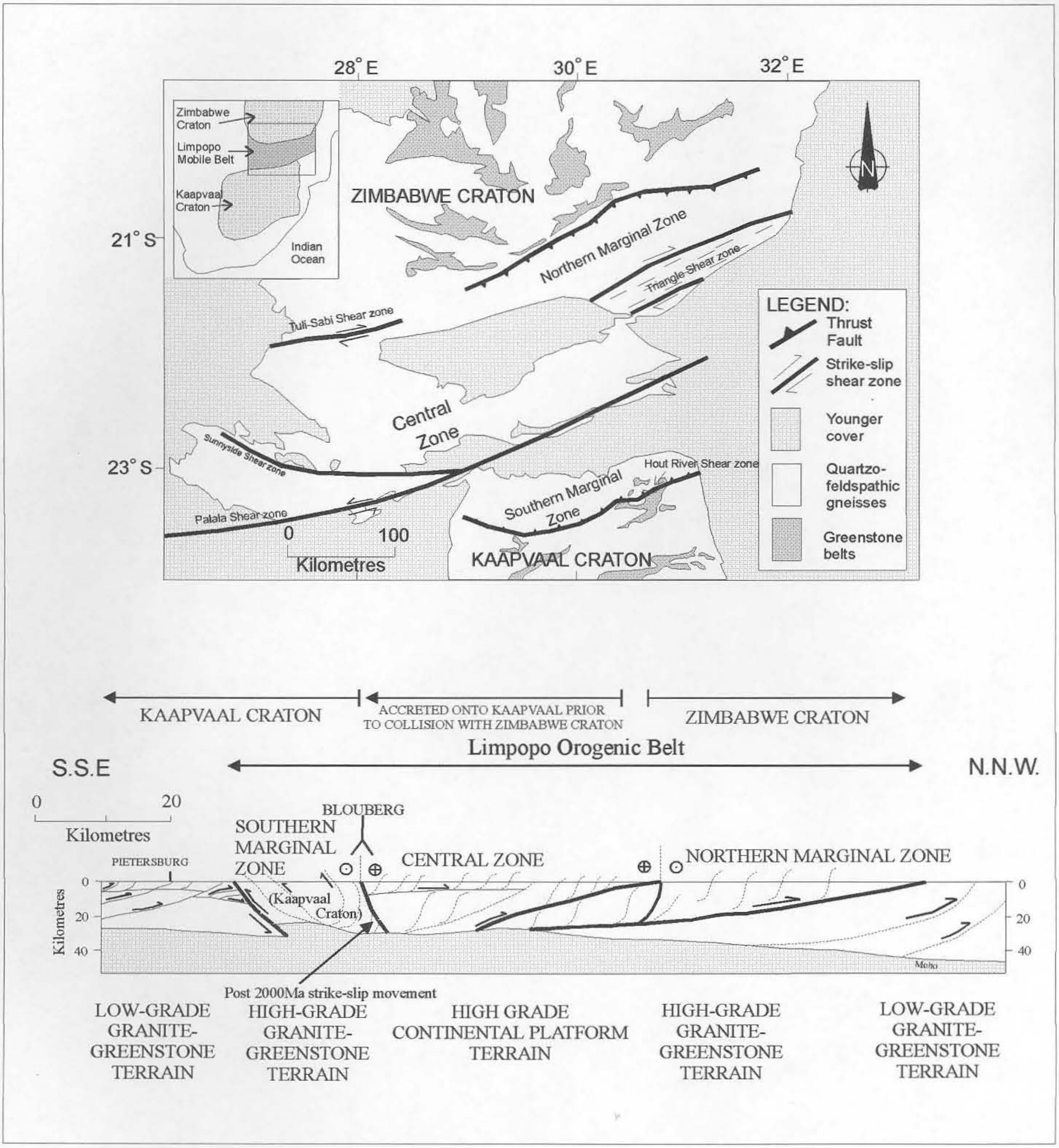

Figure 1.3: Simplified geological map and cross-section showing features of the Limpopo Mobile Belt (after Kröner *et al.*, 1999 and Roering *et al.*, 1992).

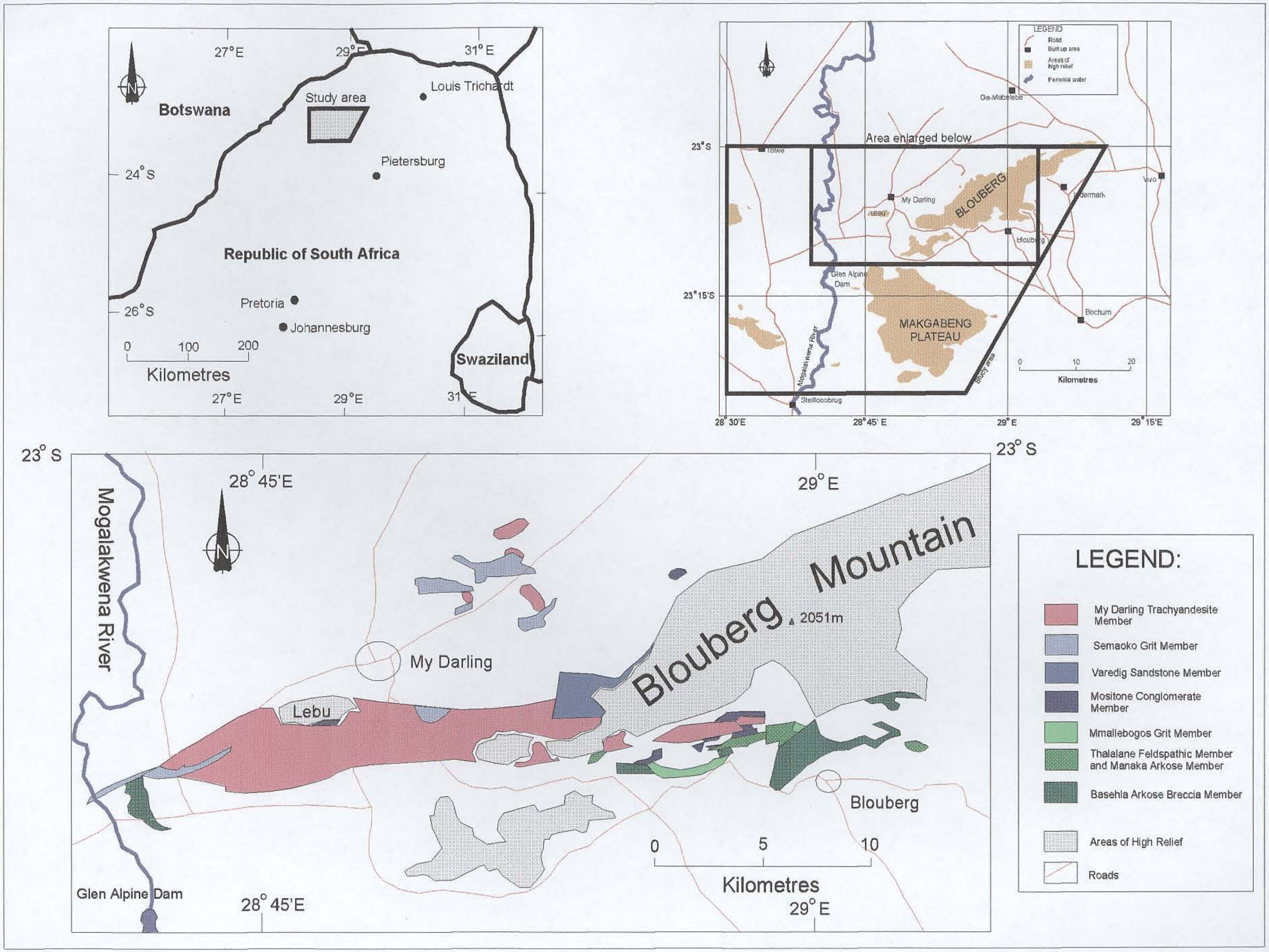

Figure 1.4: Map showing the distribution of the lithostratigraphic units in the Blouberg Formation (after Jansen, 1982).

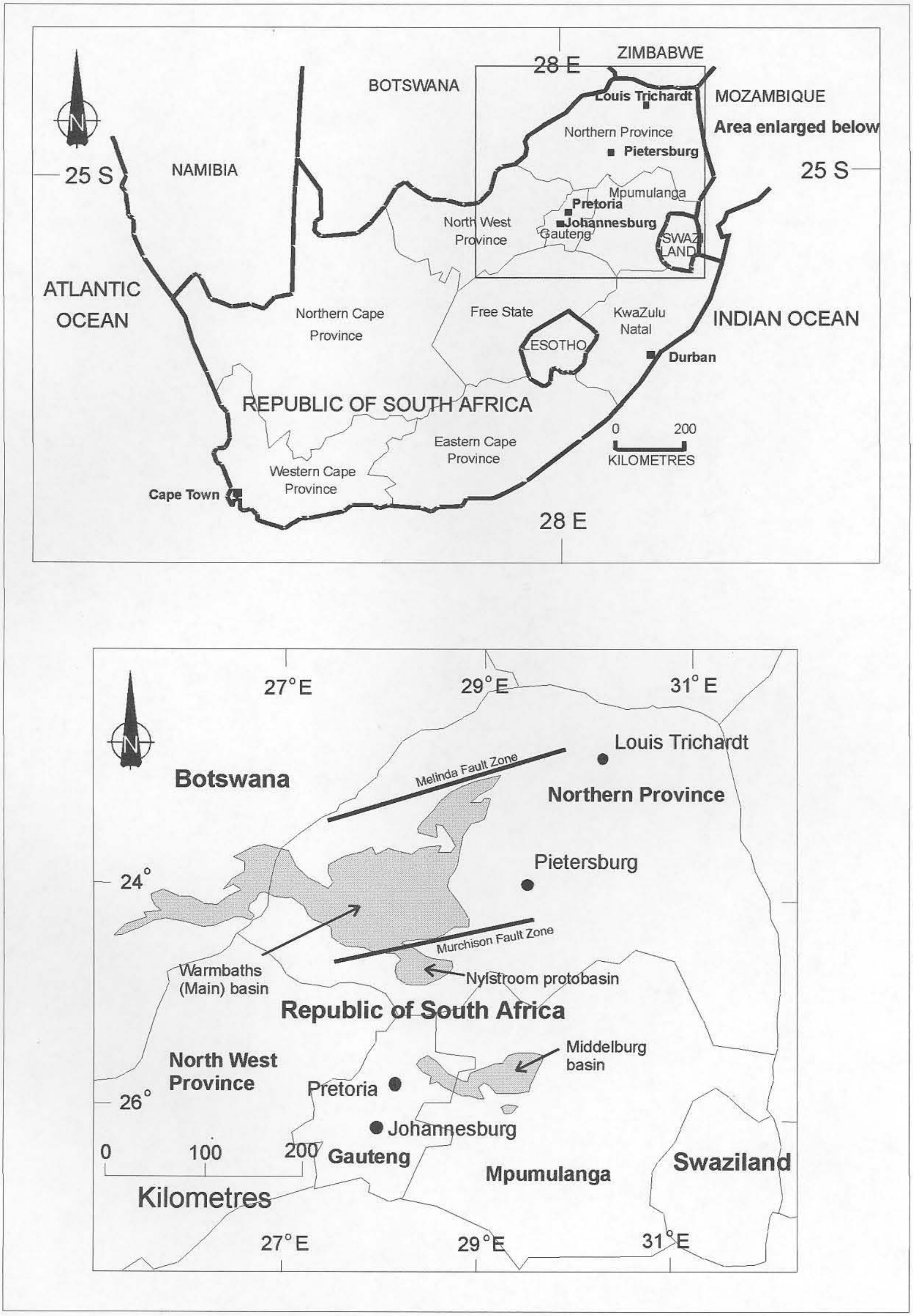

Figure 1.5: The distribution of the Waterberg Group (after Callaghan *et al.*, 1991).

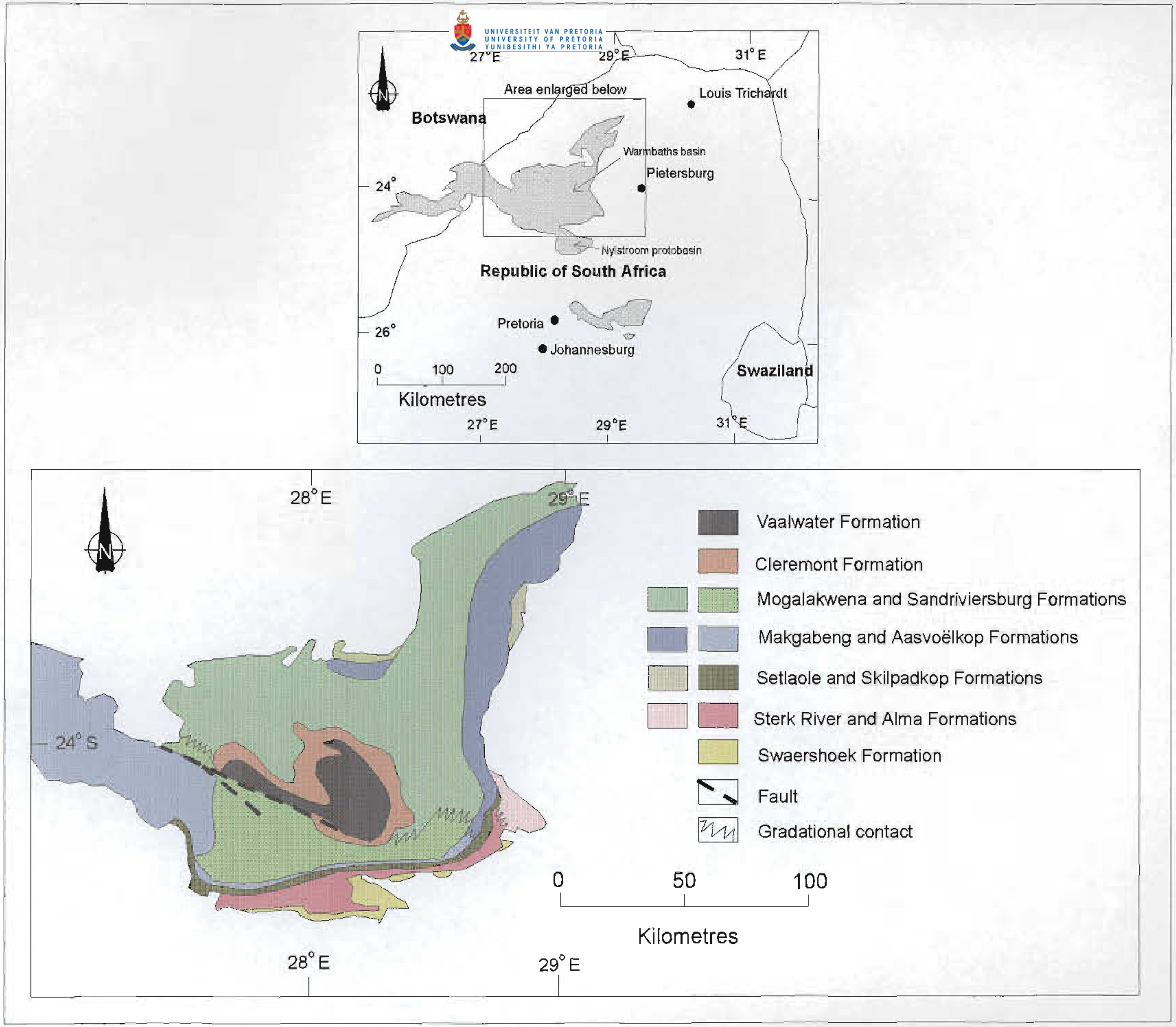

Figure 1.6: Map showing the distribution of the lithostratigraphic units in the main Warmbaths basin of the Waterberg Group (after Callaghan *et al.*, 1991).

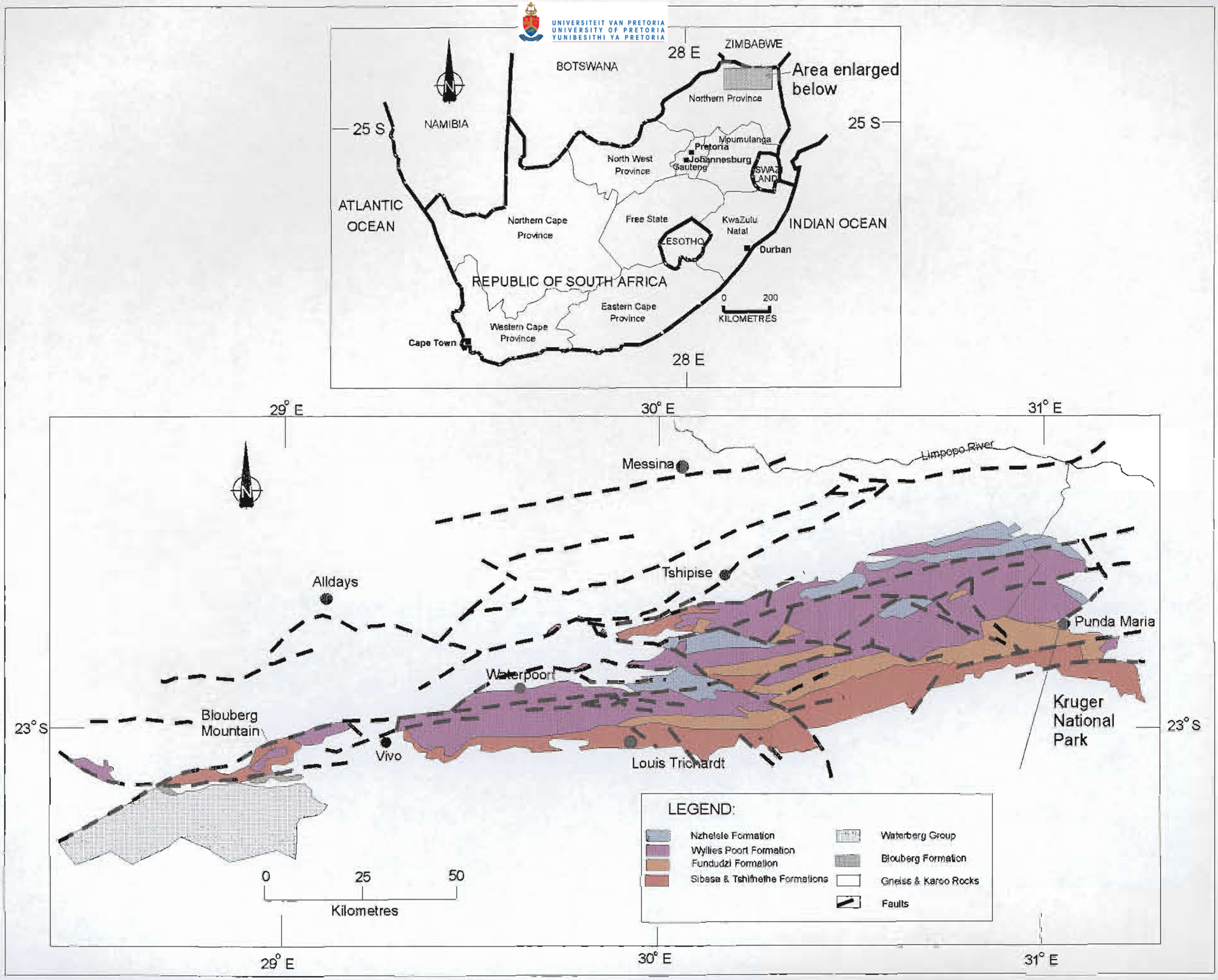

Figure 1.7: Map showing the distribution of the lithostratigraphic units in the Soutpansberg Group (after Barker *et al.*, in press).

FORMATION	MEMBER
BLOUBERG FORMATION	Semaoko Grit (300-400m)
	My Darling Trachyandesite (Max. 800m)
	Varedig Sandstone (350-450m)
	Mositone Conglomerate (100-150m)
	Mmallebogos Grit (400m)
	Thalalane Feldspathic Sandstone (450-600m)
	Manaka Arkose (150-250m)
	Basehla Arkose Breccia (Max. 1200m)

Table 1.1: Stratigraphic subdivision of the Blouberg Formation (after S.A.C.S., 1980).

GROUP	SUBGROUP	South / southwest & central parts	Southeast & central parts	North / northeast & central parts	Nylstroom area	Middelburg area
WATERBERG GROUP	KRANSBERG	Vaalwater Formation (475m)				
		Cleremont Formation (125m)				
		Sandriviersberg Formation (1250m)	Sandriviersberg/ Mogalakwena Formations	Mogalakwena Formation (1250-1500m)		
	MATLABAS	Aasvoelkop Formation (300-600m)	Aasvoelkop/ Makgabeng Formations	Makgabeng Formation (380-1000m)		
		Skilpadkop Formation (450-600m)	Skilpadkop Formation (450-600m)	Setlaole Formation (450m)		
	NYLSTROOM	Alma Formation (3000m)	Sterk River Formation (500-1500m)		Alma Formation	
		Swaershoek Formation			Swaershoek Formation (2500m)	Wilge River Formation

Table 1.2: Stratigraphic subdivision of the Waterberg Group (after Callaghan *et al.*, 1991).

GROUP	FORMATION
SOUTPANSBERG GROUP	Nzhelele Formation (1000-2000m)
	Wyllies Poort Formation (1000-4000m)
	Fundudzi Formation (0-2800m)
	Sibasa Formation (0-3300m)
	Tshifhefhe Formation (0-9m)

Table 1.3: Stratigraphic subdivision of the Soutpansberg Group (after S.A.C.S., 1980).

Date	Author(s)	Nature of investigation
1977	Barton and Ryan	Review of geochronology
1977	Barton *et al.*	Rb-Sr dating of ancient dykes
1977	Du Toit and Van Reenen	Metamorphism and structure in the S.M.Z.
1977a 1977b	Van Biljon	General Introduction Relationship between plate tectonics and ancient mobile belts
1977	Van Reenen and Du Toit	Mineral reactions and timing of metamorphism
1978	Barton *et al*.	Relationship between Rb-Sr and U-Th-Pb dates in Sand River gneiss
1978	Van Reenen	High-grade metamorphic petrology of the S.M.Z.
1978	Van Reenen and Du Toit	Metamorphic petrology of granulites
1979a 1979b	Barton	Geochemistry, Rb-Sr isotopes and tectonic setting of post-kinematic igneous rocks. Crustal evolution
1979a 1979b	Barton *et al*.	Rb-Sr and U-Pb dating of Singelele and Bulai gneisses Geology, age and tectonic setting of the Messina Layered Intrusion
1979	Du Toit	Geology and structure of the Levubu and Bandelierskop intrusions
1979	Fripp *et al.*	Structure and origin of the Singelele gneiss
1980	Barton	Pattern of Archaean crustal evolution
1980	Fripp *et al.*	Deformation in the Archaean Kaapvaal craton
1980	Horrocks	Archaean supracrustal rocks in the L.M.B.
1981a 1981b 1981c	Barton	Status of isotopic investigations Pattern of Archaean crustal evolution Limpopo excursion guidebook
1981	Barton and Key	Tectonic development of the L.M.B., and the Archaean crustal evolution.
1981	Fripp	Sand River gneiss
1981	Horrocks	Geology between Messina and Tshipise
1981	Robertson and Du Toit	General report on Limpopo Belt
1982	Fripp	Geology around Sand River, Messina

Table 1.4: Summary of the work of the Limpopo Working Group, 1977-1982.

Date		Author(s)	Nature of investigation
1983	a	Barton *et al.*	Isotopic studies of the Sand River gneiss
		Barton and Key	Ages and geological setting of Central Zone rocks in eastern Botswana
	b	Barton *et al.*	Ages and chemical composition of deformed mafic dykes in the Central Zone
	a	Barton	Lead isotope evidence for age of Messina Layered Intrusion
	b	Barton	Lead isotope studies of Banded Iron Formation in the Central Zone
		Barton and McCourt	Rb-Sr evidence for the age of the Palala Granite
		Ryan *et al.*	Isotope studies on copper deposits near Messina
	c	Barton *et al.*	Geochronology in the S.M.Z.
		Watkeys *et al.*	Overview of the Central zone
		Horrocks	Geology of the area between Tshipise and Messina
		Fripp	Geology of the area around the Sand River, Messina
		Brandl	Geology and geochemistry of supracrustal rocks of the Beitbridge Complex
		McCourt	Archaean lithologies in the Koedoesrand area
		Du Toit *et al.*	Geology, structure and metamorphism of the S.M.Z.
		Van Reenen	Metamorphic petrology in the S.M.Z.
		Key *et al.*	Evolution of the S.M.Z. In Botswana
		Coward	Tectonics of the Limpopo Belt
		Barker	Geotectonic model for the Soutpansberg Group
	c	Barton	Summary of work

Table 1.5: Summary of the work on the Limpopo Mobile Belt in the Geology Society of South Africa special publication (Van Biljon and Legg, 1983).

Date	Author(s)	Nature of investigation
1982	Light	Continental collision as cause of Limpopo Belt
1985	Watkeys and Armstrong	Deformed lamprophyric dykes in the Central zone
1987	McCourt and Vearncombe	Shear zones bounding the Limpopo Belt
1987	Van Reenen *et al.*	Deep crustal response to collision in the Limpopo Belt
1988	Brandl	Geological setting and southern boundary of the Central zone
1990	Barton *et al.*	Significance of 3 Ga mafic dykes in the Central zone
1990	Boryta and Condie	Geochemistry and origin of the Beitbridge complex
1990	Brandl	Geology of the Alldays gneiss and associated dykes
1990	Brandl and Reimold	Pseudotachylite occurrences in the Palala Shear zone
1990	Retief *et al.*	Zircon ages in the Limpopo Belt
1990	Van Reenen *et al.*	Granulite facies rocks of the Limpopo Belt
1992a 1992b	Barton and Van Reenen	Age of the Limpopo Orogeny Significance of biotite and phlogopite ages for the thermal history of the Central zone and S.M.Z
1992	McCourt and Vearncombe	Shear zones in the Limpopo Belt
1992	Roering *et al.*	Tectonic model for the evolution of the Limpopo Belt
1992	Stevens and Van Reenen	Origin of metapelitic granulites in the S.M.Z.
1992	Treloar *et al.*	Himalayan-Tibetan analogies for the evolution of Limpopo Belt
1993	Rollinson	Terrane interpretation of the Limpopo Belt
1994	Barton *et al.*	Metamorphic events: implications for P-T pathway application in complex metamorphic terrains
1994	Mkweli *et al.*	Timing of thrusting in the N.M.Z.
1995	Berger *et al.*	Geochemistry, geochronology of charoenderbites in the N.M.Z.
1995	Holzer	Petrology of the Bulai pluton and the 2.0Ga overprint in the Central zone
1995	Ichihashi and Miyano	P-T trajectory of the Central zone
1995	Kamber and Biino	Evolution of high T, low P granulites in the N.M.Z.
1995a 1995b	Kamber *et al.*	2.0 Ga event in the Triangle Shear zone Proterozoic transpressive deformation in the N.M.Z.
1995	Rollinson and Blenkinsop	Magmatic, metamorphic and tectonic evolution of the N.M.Z.
1995	Tsunogae and Yurimoto	Zircon chronology
1996	Barton	Messina Layered Suite: Crustal contamination of an Archaean anorthosite complex
1996	Blenkinsop and Frei	Mineralisation and tectonics at Renco mine, N.M.Z.
1996	Holzer *et al.*	2 Ga event in the Limpopo Belt
1996	Layer *et al.*	Cooling history of the Central zone
1997	Barton and Sergeev	Discordia in zircon dates
1997	Chavagnac *et al.*	Nd systematics in migmatites
1997	Holzer *et al.*	Granulite facies metamorphism: Evidence for Proterozoic collision
1997	Jaeckel *et al.*	Granitoid magmatism and high grade metamorphism in the Central zone
1998	Hofmann *et al.*	Field relationships of Sand River gneiss
1998	Holzer *et al.*	Pb dating of metamorphic minerals in the Central zone
1998	Kröner *et al.*	Field relationships and age of Beitbridge Complex
1999	Kröner *et al.*	Zircon ages for the granitoid gneisses in the Central zone

Table 1.6: Summary of recent work on the Limpopo Mobile Belt (1982-1999).

Date	Author(s)	Nature of investigation
1872	Moodie	Presence of copper/lead occurrence
1897	Harger	Strata in Middelburg basin (Wilge Rivier Fm) unconformably overlie Transvaal strata, and are, in turn, unconformably overlain by Karoo strata.
1898a 1898b 1901 1904	Molengraaff	Named the rocks the 'Waterberg Sandstone Formation', and later the 'Waterberg Series', after the Waterberg District of Northern Transvaal.
1904 1905a 1905b 1907 1908 1909a 1909b 1910	Mellor	Relationship between Waterberg and surrounding strata Geology of Springbok Flats and Rhenosterkop Geology of Middelburg Discussion of Merensky's paper Geology west of Potgietersrus Description of the basal Alma Formation and its nonconformable relationship to the underlying Bushveld granite. Geology north of Nylstroom
1904	Holmes	General geology of Northern Transvaal
1904	Jorrisen	Mentioned in discussion of the Chuniespoort Group.
1905	Hatch and Corstorphine	General geology of South Africa
1908	Merensky	Tin deposits around the Bushveld granite.
1910	Anderson	General geology between Nylstroom and the Limpopo River.
1912	Kynaston and Mellor	Geology of the Warmbaths area (Map explanation).
1924	Daly and Molengraaff	Refuted claims that the Bushveld granites were intrusive into the Waterberg strata at Gatkop.
1932	Hall	Estimated thickness of Waterberg strata: max. 2700m north of Warmbaths
1942	Le Roex	Geological Survey report, Potgietersrus
1942 1948	Strauss	Dolerite and granophyre near Potgietersrus Geological Survey report, Potgietersrus
1944	Du Preez	Structural geology east of Thabazimbi.
1960	Cullen	Distribution of Waterberg strata in southern Bechuanaland Protectorate (Botswana).
1965	Glatthaar	Pyroclastics south east of Rust de Winter

Table 1.7: Summary of early work on the Waterberg Group, 1872-1965.

Date	Author(s)	Nature of investigation
1963	Wilke	Geology of sheet 2428B
1963 1966 1967	De Villiers	Geology of sheet 2428B Sedimentary geology around Potgietersrus Sedimentology of Loskop and Waterberg strata
1969 1970 1973	De Vries	Stratigraphic subdivision (current usage) Geology of the southern Waterberg Sedimentary structures in the southern and central Waterberg
1970 1971 1972a 1972b 1972c	Frick	Identified strata in the south east as having been deposited within a littoral, possibly tidal palaeoenvironment Heavy mineral deposits in the Waterberg basin Geology and geochemistry of the south eastern Waterberg Heavy mineral deposits in the Waterberg Group
1969 1970a 1970b 1971 1972 1975	Meinster	Waterberg bedding and cross-bedding Deformed cross-bedding in the Swaershoek Formation Geology between Matlabas and Buffelsdrift Geology between Heuningfontein and the Makapansberge Geology of Gatkop, near Thabazimbi Structures at Gatkop
1975	**Meinster and Tickell**	**Aeolian deposits in the Makgabeng Formation**
1975	**Tickell**	**Braided river palaeoenvironment proposed for the Mogalakwena Formation**
1971a 1971b 1972a 1972b	De Bruiyn	Geology around Vaalwater Geology between Mokamole and Blouberg Folded cross-bedding in the Waterberg Group Report on Nooitgedacht lead mine
1972a 1972b	Du Plessis	Relationship between Waterberg Group and Bushveld Complex Geology of the area north-east of Warmbaths
1969 1970a 1970b 1975a **1975b** **1976** **1982**	**Jansen**	Structural evolution of the southern Waterberg basin Volcanic and sedimentary rocks in the south of the Waterberg basin Geology of the Nylstroom area Sedimentology and structures of Precambrian basins on the Transvaal craton. **Aulacogen model for Soutpansberg Group** **Geology of the Blouberg area (Waterberg and Soutpansberg groups).** **Review of all available data. Termination of this period of study.**
1970 1972	Jansen *et al.*	Geology between Thabazimbi and Rankins pass Waterberg tectonics and sedimentation

Table 1.8: Summary of work on the Waterberg Group undertaken by the Geological Survey of South Africa, 1963-1982 (items in bold are most relevant to this study).

Date	Author(s)	Nature of investigation
1908	Mellor and Trevor	Description of some rocks in the Soutpansberg area, and general impressions regarding structure.
1925	Rogers	Geology and petrography of lavas
1931	Janisch	Notes on Fundudzi Lake
1938	Taljaard	Physiographic description
1939	Kent	Description of a copper occurrence on Bosch 407
1955	Van Eeden *et al.*	Explanation of Sheet 42 (Soutpansberg) Geological Survey.

Table 1.9: Summary of early work on the Soutpansberg Group, 1908-1955.