

CHAPTER 8: GEOLOGICAL HISTORY.

In this chapter, the geological history for the area in the vicinity of Blouberg mountain will be considered. The data from each of the stratigraphic units presented in Chapters 2-6, and the structural data from Chapter 7 will be combined and analysed to produce a provisional model of basin evolution for the study area.

8.1: Tectonic setting:

8.1.1: Tectonic interpretation of basement rocks:

The foliated rocks of the basement are positioned along the projected extension of the E.N.E.-W.S.W. Palala Shear Zone. In the Palala area (23°20'S; 28°05'E), these rocks consist principally of mylonitic and ultra-mylonitic rocks (McCourt and Vearncombe, 1992). The foliation planes in the banded gneiss recorded in the study area are generally parallel to the foliation in the mylonite at the Palala Shear Zone. The higher-grade gneiss recorded in the study area suggests, however, that deeper crustal levels in the Shear Zone are represented at Blouberg. The general sinistral sense of movement recorded in the banded gneiss at Blouberg compares favourably with the sinistral sense of movement recorded at the Palala Shear Zone (e.g. McCourt and Vearncombe, 1992; Broekhuizen, 1998).

The map of foliations in the gneiss shown in Figure 2.10 shows foliations with at least two contrasting strikes. From Figure 2.10 an earlier foliation can be identified, which strikes N.E.-S.W., and is cut by a secondary foliation, which strikes W.N.W.-E.S.E. This secondary foliation may be related to the Sunnyside Shear Zone, which is thought to merge with the Palala Shear Zone in this area (Figure 1.3; Kröner *et al.*, 1999). Broekhuizen (1998) also reported a secondary mylonitic (dextral) foliation with an E.S.E strike in the Palala area.

The Palala Shear Zone is generally regarded as representing the suture between the northern edge of the Kaapvaal Craton (Southern Marginal Zone) and the Central Zone

of the Limpopo Mobile Belt (McCourt and Vearncombe, 1992). The Central Zone is regarded as having docked with the Kaapvaal Craton during c. 2.7 Ga (McCourt *et al.*, 1995; McCourt and Armstrong, 1998) or c. 2.0 Ga (Kröner *et al.*, 1999) transpressional collision. The structural data exhibited by rocks of the Palala Shear Zone provide evidence for the kinematics of this event (i.e. they indicate sinistral-reverse oblique slip). Having identified the rocks beneath Blouberg mountain as representing a continuation of the Palala Shear Zone, this suggests that the strata in the northern part of the study area are located on the suture on the northern edge of the Kaapvaal Craton. Any model for basin evolution constructed for strata in the study area should consider this fundamental structure as a possible tectonic control.

8.1.2: Tectonic interpretation of the Blouberg Formation:

The oldest of the sedimentary strata developed nonconformably on the basement gneiss are the rocks of the Blouberg Formation, which is subdivided into two contrasting members (Lower and Upper; Chapter 3). The extent of the outcrop of both of these members is restricted to a narrow zone across the study area (Appendix 1). If the c.10km-wide Palala Shear Zone is projected eastwards through the Blouberg area, it can be seen that the extent of the preserved Blouberg strata is restricted to the area likely to be underlain by the Shear Zone.

Of the two members of the Blouberg Formation, only the Lower Member was found to outcrop over a reasonably large area. The dip-directions of bedding planes of the Lower Member in the vicinity of Blouberg mountain, shown in Figure 7.7, and especially the presence of northward-dipping overturned beds in the Lower Member of the Blouberg Formation, suggest that southward-vergent thrusts have affected the Blouberg and older rocks (Section 7.2). This is corroborated by the presence of low-angled, southward-vergent thrust faults (e.g. Figures 7.4, 7.5 and 7.6), which have locally thrust basement rocks over Blouberg strata, and by the presence of southward-vergent reverse faults cutting the Lower Member of the Blouberg Formation (e.g. Figures 7.10, 7.11 and 7.12). Such high-angled reverse faults could also, however, be interpreted as being related to positive flower structures, caused by transpressional

movement along the Palala Shear Zone. Additionally, geophysical evidence from the Melinda Fault outside the field area suggests that the Melinda Fault is vertically dipping (G. Brandl, pers. com.). However, it generally appears that the Lower Member of the Blouberg Formation has been subjected to a positive basin inversion, resulting from a stress field with an approximately north-south orientated compression.

The contrast in lithofacies between the generally sandy Lower Member and generally conglomeratic Upper Member of the Blouberg Formation, represents a clear coarsening-upwards succession for the Formation as a whole (Figure 3.1). The conglomeratic Upper Member consists of large, angular cobbles of quartz, quartzite and feldspathic, foliated rocks which are likely to have been derived from a proximal source of basement gneiss. As shown in Figure 3.1, the contact between the Lower and Upper Members is sharp, and can be interpreted to reflect depositional changes as a response to sudden tectonic uplift of a source area nearby; a single coarsening-upward succession as suggested above is thus not pertinent.

In contrast to the steeply-dipping Lower Member, rare outcrops of Upper Member strata around Blouberg mountain have bedding planes with only low angles of dip (Figure 7.7; Appendix 1). Thus the Upper Member appears to be unaffected by the tectonic event which was responsible for the southward-vergent thrusting and overturning of the older Blouberg strata. It is possible that the southward-vergent event, and the event responsible for the tectonic uplift of the Blouberg Upper Member source area were the same, as both events post-date the Lower Member, but pre-date the deposition of the Upper Member.

In contrast, the 1400m of Blouberg strata recorded in the Kranskop section appear to be unaffected by any southward-vergent tectonic event. Instead, bedding planes dip consistently and steeply to the west, which is difficult to reconcile with the tectonic model proposed above. However, the considerable distance (c. 25km westwards) between the Kranskop strata and the outcrops in the area of Blouberg mountain, and the fact that the Kranskop strata lie adjacent to the Melinda Fault (where the northern

and southern strands have merged), suggest that the tectonic history in this area may have been different. It seems that Kranskop strata were only overturned *after* the deposition of the Upper Member of the Blouberg Formation, as both Lower and Upper members exhibit comparable deformation. Although difficult to interpret, this may represent a continuation of syn-Blouberg deformation into post-Blouberg times. The change in dip direction between the Blouberg mountain strata (which dip to the south or are overturned and dip to the north) and the Kranskop strata (which dip steeply towards the west) may be due to localised rotation or drag folding alongside the Melinda Fault. As Blouberg strata in both areas are unconformably overlain by gently-dipping Mogalakwena strata, this suggests that the age of all the events proposed above can be constrained to at least pre-dating deposition of the Mogalakwena Formation.

8.1.3: Tectonic interpretation of the Waterberg Group (Setlaole, Makgabeng and Mogalakwena formations):

Generally these strata are not particularly deformed in the southern part of the study area. However it is important to note the fact that the Makgabeng Formation does not appear to be preserved less than 4km south of the southern strand of the Melinda Fault. This may have been due to the presence of an elevated palaeotopography to the north in the immediate vicinity of the southern strand of the Melinda Fault, so that erosion rather than deposition prevailed there during Makgabeng times. The earlier, syn-Blouberg basin inversion (southward-vergent thrusts) may have produced such high relief northwards of the southern strand of the Melinda Fault, and implies a small time gap between Blouberg and Setlaole / Makgabeng sedimentation.

Outcrops of the younger Mogalakwena Formation, however, underlie areas much further north than the Makgabeng Formation. Adjacent to the southern strand of the Melinda Fault, bedding begins to dip more steeply (Appendix 1), and contrasting facies of the Mogalakwena Formation are juxtaposed against each other, separated from each other by brittle faults (Figures 2.14, 2.15, 7.1, 7.2 and 7.3). Though no

foliation exists in the crush breccia fault rocks, the following evidence suggests that these faults are extensional (normal) dip-slip faults.

The presence of reverse fault planes dipping steeply to the north in the Blouberg Formation (Figure 7.10) suggest that a northward-dipping fault plane was already established beneath the Mogalakwena rocks. The E.N.E.-W.S.W striking veins throughout the fault rocks, and the generally E.N.E.-W.S.W.-striking post-Mogalakwena, syn-Sibasa dyke swarm cutting the Waterberg Group (Figure 6.2; see Section 8.1.4 below), suggest that the syn-Blouberg reverse faults may have reactivated during syn-Sibasa times under a stress field with approximately N-S orientated extension. Thus the faults are likely to be northward-dipping normal faults, which were active prior to the deposition of the Soutpansberg Group in the area. These normal faults are envisaged to have juxtaposed the upper (sandstone and granulestone) Mogalakwena in the hanging wall to the north, against the lower (conglomeratic) Mogalakwena strata in the footwall.

An alternative model for the juxtaposition of distal facies of the Mogalakwena Formation against inferred proximal facies is that dextral strike-slip movement occurred along the southern strand of the Melinda Fault. However, in order to juxtapose these sediments, a displacement of around 25km is necessary in order to juxtapose such relatively mature sediments from distal parts of the basin with more proximal rocks. Such a large dextral displacement is considered unlikely in view of the narrow, localised extent of the crush breccia.

The southern strand of the Melinda Fault therefore appears to comprise northward-dipping faults which were active as reverse faults in syn-Blouberg times, and as normal faults in post-Mogalakwena / syn-Sibasa times. Generally it appears that the later event must have occurred as a reactivation along the older fault planes. Locally, however, field relationships suggest that the reverse faults failed to reactivate, e.g. in the vicinity of Dantzig, at the eastern end of the southern foothills of Blouberg mountain ($23^{\circ}06.5'S$; $29^{\circ}01.5'E$). Here, it seems that the normal, syn-Mogalakwena component is only poorly developed, as basement rocks occupy areas of relatively

high topography in the hanging wall to the north, compared with gneiss north of the southern strand of the Melinda Fault further to the west, which outcrops only at low altitudes. This indicates that little or no syn-Sibasa normal displacement has occurred in the hanging wall in the Dantzig area, and evidence for only the syn-Blouberg reverse faults was recorded (Section 7.2).

8.1.4: Tectonic interpretation of the Soutpansberg Group:

The lack of outcrops of the Sibasa Formation with any significant preserved structures has not allowed for any direct tectonic inferences to be made. However, a comparison between the incompatible trace element pattern shown in Figure 5.5 (Sibasa basalt) and Figure 6.4 (dykes cutting the Waterberg Group), shows that both igneous rocks have comparable geochemistry (Figure 8.1). In particular, Figure 8.1 shows that relative ratios of incompatible trace elements (Nb to Yb) for the six samples (3 dolerite dykes, 3 basaltic Sibasa lavas) remain constant, throughout varying degrees of fractional crystallisation. This indicates a common parental magma for both the dykes and the Sibasa lavas. Similarly, a plot of Zr against Y, shown in Figure 8.2, suggests that all six samples likely had the same parental magma, as these highly incompatible elements maintain the same ratios relative to each other throughout successive episodes of fractional crystallisation. Thus it seems that the dyke swarm cutting the Blouberg Formation and Waterberg Group to the south of the southern strand of the Melinda Fault was a feeder to the lavas of the Sibasa Formation above. Such an interpretation strengthens the hypothesis that the Soutpansberg Group post-dates the Waterberg Group, as Soutpansberg-aged rocks appear to intrude the Waterberg strata. Additionally, it must be borne in mind that the dyke swarm, which intrudes the Waterberg Group so intensely, cannot be demonstrated to cut the Wyllies Poort Formation (Appendix 1). Reliable age-dating of Sibasa basalts and dolerite from the dykes may confirm the suggestion that the dykes cutting the Waterberg Group acted as feeders to the Sibasa Formation, and in addition provide age constraints for the minimum age for the deposition of the Waterberg Group (Mogalakwena and older formations).

Trace element signatures from basalts of the Sibasa formation and inferred feeder dykes to the lavas suggest a subduction-related, calc-alkaline signature. However, it must be borne in mind that many igneous rocks throughout the Kaapvaal Craton (e.g. volcanics of the Transvaal Supergroup and mafic intrusives of the Bushveld Complex) also possess a subduction-related geochemical signature (e.g. Harmer and von Gruenewaldt, 1991). Such signatures are thought to be inherited from lithospheric source rocks, which had been generated through subduction-related processes at the time of Archaean crustal development, rather than indicating the actual tectonic regime under which the magmas were generated (Harmer and von Gruenewaldt, 1991).

The orientation of the strike of dykes developed in the Blouberg and Waterberg strata is generally E.N.E-W.S.W (Figure 6.2). This suggests that, at the time of dyke intrusion and eruption of the Sibasa basalts, an extensional stress field prevailed and the extension direction was orientated perpendicular to the plane defined by the dyke swarm (i.e. approximately north-south extension; Section 8.1.3).

It is important to note that neither the Sibasa Formation nor the Wyllies Poort Formation can be demonstrated to outcrop south of the southern strand of the Melinda Fault in the study area. Similarly, no Soutpansberg strata have been recorded southwards of the projected line of the Palala Shear Zone anywhere else along the Soutpansberg basin (Cheney *et al.*, 1990). However, outcrops have been recorded further to the north, including xenoliths that have been correlated with the Soutpansberg Group within the Venetia Kimberlite pipe near Messina (Barton and Pretorius, 1996).

The structural data from the northern strand of the Melinda Fault, where it cuts the Wyllies Poort Formation, suggest that the northern strand may be a dextral strike-slip fault, though with a considerable dip-slip component developed locally. This can be inferred from the rose diagrams shown in Figures 7.26 and 7.27, and stereographic projections shown in Figures 7.28 and 7.29. Figure 7.26 shows that joint planes in the Wyllies Poort Formation have a dominant strike towards the N.W, suggesting

maximum extension towards the N.E. and S.W., with the direction of maximum compression from the north-west and south-east. Similarly the strike of veins (Figure 7.27) suggest compression from the N.W. and S.E. Figure 7.28 show that the dominant planes of small-scale faults match the geometry of the northern strand of the Melinda Fault (a steeply-dipping E.N.E.-W.S.W.-striking fault), and Figure 7.29 shows that slickenside lineations recorded from these fault planes dominantly show an azimuth directed E.N.E.-W.S.W, with generally shallow plunges. However many of these lineations (those with a steep plunge) also demonstrate a considerable dip-slip component developed on many of the fault planes. Bedding orientations from outcrops of the Wyllies Poort Formation in the vicinity of the northern strand of the Melinda Fault, plotted in Figure 7.17, suggest that the Wyllies Poort Formation is locally folded, with a horizontal fold-axis with a N.E.-S.W. trend.

Together this evidence suggests that a N.W. to S.E. -directed compression was responsible for general dextral (though also oblique) displacement along the northern strand of the Melinda Fault (Figure 8.3) at a time which post-dates the deposition of the Wyllies Poort Formation. The regional pattern of displacement of the Wyllies Poort Formation by the northern strand of the Melinda Fault, especially where the projected line of the Fault displaces the Soutpansberg strata north east of Vivo (c. 22°58'S; 29°21'E), by up to 17km, appears to confirm the dextral sense of movement proposed here.

8.1.5: Conclusions regarding the age relationships and tectonic setting in the study area:

The age-relationships between the stratigraphic units that can be inferred from this work are shown in Table 8.1, where they are compared with the existing stratigraphic subdivision, as proposed and debated by Jansen (1976) and Meinster (1977). The map in Figure 8.4 shows the location of an idealised cross-section from north to south through Blouberg mountain, which is shown in Figure 8.5. This sketch cross-section illustrates the structural geology and is constructed with regard to the proposed tectonic history of the study area described above. Similarly, cross-sections through

the Blouberg mountain area, constructed from the map in Appendix 1, are provided in Appendix 3.

8.2: Depositional Palaeoenvironments:

In this section, the depositional setting of each of the sedimentary units recorded in the study area will be extrapolated from the data presented in earlier Chapters.

8.2.1: Depositional setting of the Blouberg Formation:

The Lower Member of the Blouberg Formation is characterised by relatively small sets of trough cross-bedded purplish sandstone with gravel-filled channel forms developed locally (Sections 3.2 and 3.3). The purple colour in areas which have not been reduced indicates the presence of syn-diagenetic oxygen, which may also have been present in the atmosphere at the time of deposition, although a considerable time period may elapse between deposition and final lithification of sedimentary rocks. The content of carbon dioxide in the Precambrian palaeo-atmosphere decreased in a complex manner over geological time, as illustrated by palaeosol data (Holland et al., 1986; Rye and Holland, 1998). Detrital pyrite and uraninite deposits, which are unstable under oxidising conditions, are restricted to deposits older than c. 2.0 Ga, and BIF, with few exceptions, pre-date 1.85 Ga (Kasting, 1991; Condie, 1997). Red beds *sensu stricto* (i.e., those with clasts partially covered by iron-stained clays rather than just having red pigmented intergranular material) were virtually absent prior to c. 2.0 Ga (e.g. Eriksson and Cheney, 1992). The appearance of red beds thereafter suggests a substantial increase in atmospheric oxygen levels (Kasting, 1993; Rye and Holland, 1986), as also supported by palaeosol compositions which support a rapid rise in atmospheric oxygen close to c. 1.9 Ga (Holland, 1994). Thus the presence of red beds throughout the entirety of the siliciclastic successions in the study area suggests that none of the sediments is older than c. 1.9-2.0 Ga.

The sedimentary facies associations in the Lower Member of the Blouberg Formation are composed of facies of planar and trough cross-bedded sandstone (Sp and St),

together with trough cross-bedded pebble conglomerate (Gt). Although it was impossible to establish architectural elements within the Blouberg Formation, which can be used in order to discriminate fluvial styles (Miall, 1992) the facies comprising the Blouberg Formation can be interpreted, instead of architectural elements, in order to infer a depositional setting for the Blouberg Formation. Facies St can be interpreted to have formed due to the migration of aqueous dunes under lower flow regime conditions, using the scheme of Miall (1977) (Table 8.2a). Similarly, facies Sp (planar cross-bedded sandstone) can be interpreted to reflect deposition from linguoid bars or sand waves, and stratified gravel channels (Gt) can be interpreted as having being deposited as minor channel fills (Table 8.2a). These interpretations can be applied throughout the Lower Member of the Blouberg Formation, which is comprised of associations of these three facies. These interpretations, when considered together with the fact that the Blouberg strata are red beds (and therefore most likely of continental origin), have relatively unimodal palaeocurrent directions (in the Blouberg mountain area; Figure 3.33) and the predominance of bedload material over suspended sediment in the preserved rocks (muddy sandstones are rare in the Blouberg Formation) suggest a fluvial depositional palaeoenvironment that closely resembled that of low-sinuosity distal braided sheetflood systems (Miall, 1992). Models for Blouberg rivers being meandering or having anastomosing channel systems, which require resistant river banks in order to form, are considered unlikely in view of the lack of vegetation and soil producing biota in the Palaeoproterozoic. The paucity of preserved argillaceous rocks in the Blouberg Formation would seem to qualify this argument. In the general absence of terrestrial vegetation, and with enhanced mechanical and chemical weathering processes inferred for the Precambrian (Condie, 1997; Corcoran *et al.*, 1998), it is thought that Precambrian rivers had a high bedload and low bank stability (e.g. Mueller and Corcoran, 1998; Van der Neut and Eriksson, 1999). Therefore braided river patterns are generally accepted for pre-vegetative Precambrian rivers (Schumm, 1968b; Cotter, 1978; Long, 1978; Eriksson *et al.*, 1998).

It should again be stressed that palaeohydraulic data, presented in Section 3.5, cannot be used as an indicator of absolute values of calculated parameters, but can rather be

used to compare parameters as they vary across an individual basin, or even between basins. The palaeohydraulic data presented in Figures 3.36, 3.37 and 3.38 show how calculated parameters vary from east to west, along the inferred axis of the Blouberg basin. Parameters calculated using clast sizes within channels (Figure 3.36) show reasonable systematic variation across the basin. Stream power (W) seems to increase towards the west. This is compatible with westerly-increasing discharge (Q) values (Figure 3.36) which correlate well with an approximately westward palaeocurrent direction (Figures 3.33 and 3.35). As discharge values essentially reflect the size of the catchment area (e.g., Van der Neut and Eriksson, 1999), it is inferred that more braided channels on an increasingly broad fluvial system developed towards the west; stream power would have increased along with an enhanced volume of water derived from the larger catchment area. The fact that estimations calculated for water velocity (v) appear to decrease towards the west, whereas palaeoslope (s) values increase in the same general direction (Figure 3.36), serve to illustrate the inherently large potential errors in calculations of palaeohydraulic parameters (e.g., Van der Neut and Eriksson, 1999).

Palaeohydrological parameters that were calculated from cross-stratification set heights within the Blouberg Formation show little systematic E-W variation along the probable axis of the Blouberg basin. Discharge (Q) and bankfull water discharge (Q_b) appear to increase towards the centre of the basin (Location B in Figure 3.35; Figures 3.37 and 3.38). Drainage area (A_d and $A_d(1)$) and stream length also appear to be greatest at the centre of the basin (Figure 3.38), whilst palaeoslope (s and $s(1)$) appears to be steepest at the margin of the preserved basin (Locations A and F in Figure 3.35; Figure 3.37), all of which are logical general conclusions for most basins. Channel depth (d_b) and width (w_b) show little variation (Figure 3.37).

The general increase in discharge, drainage area, and stream length towards the centre of the preserved Blouberg basin (Location B in Figure 3.35), and the concomitant lowering of palaeoslope, are compatible with braided fluvial deposition within a pull-apart basin, as also envisaged by Brandl (1986b). Pull-apart basins typically form within active strike-slip fault zones (see Section 8.1.2), have their margins bound by

fault scarps (i.e. steep palaeoslopes at the margins of the basin) and are characterised by current directions which flow towards the centre of the basin (Crowell, 1974). Rapid lateral facies changes and rapid subsidence are also characteristics of pull-apart basins (Miall, 1996). The Blouberg Formation appears to contain many of these features, especially the characteristic of rapid subsidence (e.g. 1400m of immature sediment in the Kranskop area; Section 3.2). The localised extent of the outcrop, restricted to areas overlying the Palala Shear Zone, shows that the Blouberg Formation has been deposited within an active fault zone. Assuming that the area around locations B and D (Figure 3.35) is the approximate centre of the Blouberg basin, the paucity of reliable palaeocurrent directions from the Lower Member in the western half of the Blouberg basin mean that a centrally-flowing pattern of palaeocurrent directions cannot be established. However, palaeocurrents from the Lower Member in the east of the basin are generally more systematic, and would seem to reflect transport along a basin axis. The lack of thickly developed Blouberg successions in the east, compared to the Kranskop strata, may indicate that the basin deepened westwards.

The facies present in the Upper Member of the Blouberg Formation contrast with the generally sandy-bedload braided river palaeoenvironment which is thought to have prevailed at the time of deposition of the Lower Member. The change of facies between the Lower and Upper Members is rapid; although the exact facies change is not exposed in the Kranskop section, the upper-most beds of the Lower Member, and the lower-most beds of the Upper Member are developed within 5m of each other. This suggests that palaeoclimatic conditions, which tend to be more gradual, are unlikely to have been responsible for the change in sedimentation regime. The sudden change to more immature sediment for the Upper Member can rather be interpreted as a response to tectonic activity in the source area for the Blouberg sediments (Section 8.1.2).

Within the Upper Member of the Blouberg Formation poorly stratified, matrix-supported, sub-angular to sub-rounded cobbles imply that debris flows were now the predominant depositional process (Miall, 1996). Planar-bedded and cross-bedded

coarse sandstone and granulestone red-beds can be identified locally, and trough cross-beds can be seen to be developed within channel-fills, which were erosive into the matrix-supported conglomerate. This suggests that aqueous dunes, linguoid bars and sand waves were migrating across the basin of the Upper Member, and again, considering the Precambrian conditions, would suggest a braided, fluvial palaeoenvironment between periods of debris flow. The general upwards-fining, cyclical nature of the conglomerates, which pass vertically into sandstones (where each cycle is generally less than 1m thick) suggests a fairly rapid cyclicity in discharge, perhaps due to fluctuating precipitation. Flood events generating debris flow deposits within a vegetation-free and soil-poor palaeoenvironment were followed by traction current deposition of more sandy detritus as floods abated.

8.2.2: Depositional setting of the Setlaole Formation:

The facies recorded from the few outcrops of the Setlaole Formation, which are generally similar to those found in the Blouberg Formation, suggest a fluvial setting. The presence of trough cross-bedded sandstones as a dominant facies, reflecting migrating dunes, a relatively unimodal southerly palaeocurrent data (Figure 4.3) support this interpretation. The disparity between palaeocurrent directions of the Blouberg Formation and Setlaole Formation is one of the few criteria that could be to discriminate between these two very similar strata. As with the Blouberg model, the Precambrian age would have favoured braided systems rather than anastomosing or meandering. Again, the lack of argillaceous sediment preserved within the Setlaole Formation concurs with a braided fluvial environment.

8.2.3: Depositional setting of the Makgabeng Formation:

The five facies associations identified within the Makgabeng Formation (Chapter 4) are: a.) Large-scale trough and planar cross-bedded sandstone; b.) Horizontally-bedded and rippled mudstone and sandstone; c.) Rippled and cross-bedded sandstone; d.) Massive sandstone, and e.) Pebbly sandstone. Of these, the large-scale trough and planar cross-bedded sandstone is the most abundant. The inverse-grading of sand

grains in laminations is diagnostic of wind-ripple strata deposited during aeolian sedimentation (Hunter, 1977, 1981; Kocurek and Dott, 1981), indicating that the Makgabeng Formation was generally laid down as aeolian dunes in a palaeodesert. Ripple indices recorded from ripplemarks developed on foresets are consistent with aeolian transport (McKee, 1945, 1979). Planar cross-bedded strata with steep angles of inclination are likely to represent straight-crested, transverse aeolian dunes, whereas the large-scale trough cross-bedded strata are likely to represent sedimentation in either sinuous-crested (aklé) or barchanoid sand dunes (McKee, 1979). The wedge-shaped strata present amongst the wind-ripple strata most likely represent grain flow strata, where tongue-shaped dry avalanches of sand fell from an over-steepened dune crest (Hunter, 1977, 1981; Kocurek, 1981).

The horizontally-bedded and rippled mudstone and sandstone facies, is generally found overlying very low-angled or horizontal dune foresets (i.e they were deposited at the base of the sand dunes). This facies association can be interpreted to reflect a drying-up sequence. Lower-most massive sandstone, stratified sandstone and current-rippled sandstone in this facies association may reflect deposition sheetflows during periods of heavy precipitation (Eriksson *et al.*, 2000). Upper wave and combined flow ripples with subordinate wind-rippled sandstone, adhesion warts, desiccation cracks, evaporite casts and roll-up structures appear to represent ponding of the sheetfloods, followed by gradual drying. The fact that this facies association overlies horizontally inclined foresets of dune deposits suggest that this facies association was deposited in flat-lying interdune areas, which may be prone to flooding during periods of heavy, periodic precipitation events (Eriksson *et al.*, 2000). The muddy roll-up structures recorded in the upper beds of this facies are of interest, as usually desiccating mud polygons attain a maximum curl of 90-120°, before disintegrating (Schieber, 1998). These roll-up structures, which attain a maximum curl of 720°, are most likely to represent the growth, desiccation and subsequent resedimentation of algal mats over the upper surface of the drying lake muds (Eriksson *et al.*, 2000). The growth of cells of cyanobacteria over the surface of the mud is thought to have added sufficient cohesion to the substrate to prevent disintegration upon desiccation. The parallel orientation of the roll-up structures may reflect their subsequent transport by flowing

water, so that the roll-ups are now orientated parallel to a palaeocurrent direction. Significantly these roll-up structures from the Makgabeng palaeodesert may represent the earliest recorded colonisation of a wholly terrestrial (and harsh desert) environment by microbial organisms (Eriksson *et al.*, 2000).

The various facies incorporated within the rippled and cross-bedded sandstone facies association can be interpreted to reflect the following depositional processes: strongly asymmetric ripples = aqueous current ripple migration; slightly asymmetric ripples = aqueous combined flow ripple migration; inversely-graded laminations = aeolian wind ripple strata; symmetrical ripples = wave ripple migration; massive sandstone = suspension (Simpson *et al.*, under review). These facies are consistent with an interpretation of the deposits accumulating in a playa lake setting, as all the facies described above have also been identified together in both modern and ancient playa deposits (Lowenstein and Hardie, 1985; Renaut and Last, 1994; Wedge *et al.*, 1994; Demicco and Hardie, 1994; Turner and Smith, 1997; Sweet 1999; Irmen and Vodra, 2000). The small pits recorded in these inferred playa lake deposits, filled by sandstone, mudstone or siltstone and characterised by preferential cementation, are typical of features associated with thin salt crusts (Smoot and Castens-Seidell, 1994), where wind-borne clay particles aggrade on hydroscopic films on salt crystals (Simpson *et al.*, under review). Intercalated facies of wind-ripple strata are thought to reflect the local encroachment of aeolian dunes over the margins of the dried-up playa lakes during times of low precipitation or during sandstorms, during which sand flux may fill a playa lake within hours (Wedge *et al.*, 1994).

The great thickness attained by this facies association (30m) indicates that such playa lake deposits were relatively long-lived within the Makgabeng palaeodesert, and their stratigraphic relationship with the underlying interdune deposits (Section 4.3.3.2) may indicate progressively higher rates of seasonal (or longer interval) precipitation during deposition of the Makgabeng sand sea.

The genesis of the massive sandstone facies association is more enigmatic. The presence of dewatering structures, desiccation structures and parting lineations on the

upper surface of the massive beds, and the close association with soft-sedimentary deformation of steeply-dipping aeolian foresets (which is inconsistent with the failure of dry sand), suggest that the massive sand was wet during its deposition (Simpson *et al.*, under review). Third-order (reactivation) surfaces (McKee, 1966; Brookfield, 1977) have been related to a temporary change in wind direction, and may also record a hiatus in aeolian sedimentation (Brookfield, 1977). The local onlap of massive beds onto third-order surfaces indicates that massive beds are deposited whilst aeolian deposition had been interrupted. The geometry of the massive sandstone beds, which are lenticular at the bottom of aeolian palaeodunes and channelised higher up the preserved dune face, suggests that water-saturated sand slumped down the lee face of sand dunes and became channelised, perhaps due to the funnel-shaped nature of barchan sand dunes (Simpson *et al.*, under review). The erosive nature of the massive channel-fills suggests that flow was turbulent in these channels. As the angle of repose decreased, the channels are thought to have spread out into lobe-shaped deposits over the dune plinth. The decrease in flow depth, which would have accompanied the transition from channelised to lobate geometry, would have led rather to a short-lived and localised increase in flow energy, as evinced by rare occurrences of lobate massive sandstone eroding the underlying wind-ripple strata (Simpson *et al.*, under review). Thereafter, reduced gradients led to laminar flow conditions for the massive sandstone lobes.

The triggering mechanism behind the slope failure of sand dunes, considering the fact that the massive deposits are likely to have been transported as water-saturated flows, is likely to have been periodic torrential rainfall (e.g. Loope *et al.*, 1999). The presence of interdune deposits (with rainspots locally preserved on muddy beds) grading upwards into large playa lake deposits, indicates that intermittent rainfall was significant in the Makgabeng desert's palaeoclimate. It is thought that during periods of rain, steeply-dipping foresets close to the crest of the dune and which were stable under dry conditions, became saturated with percolating meteoric water. As the maximum angle of repose of dry sand is greater than that of its wet counterpart, as the rainstorm continued failure of the top part of the dune could have occurred. Rainfall must have fallen at a rate that exceeded the rate at which water could have percolated

through the aeolian dune sand, otherwise meteoric water would have drained down to the water table, and the sand could not have become sufficiently saturated to lead to slope failure. Loope *et al.* (1999) suggest that such rainfall rate may be of the order of 250mm in 6 hours. Such a model is consistent with modern events reported from Sandhills in Nebraska, where stable, vegetated dunes collapsed after a period of heavy rainfall (Loope *et al.*, 1999). It is also possible that microbial organisms played a role in stabilising some dune crests before saturation and failure occurred.

The fact that the massive sandstone facies are absent in the lower part of the Makgabeng Formation, and can generally be seen to become more common towards the top, is evidence for long-term climatic change throughout syn-Makgabeng times, and suggests that the desert became steadily wetter through time. The increase in the presence of massive sandstones towards the top of the Makgabeng Formation is also recorded in the borehole log (Figure 4.12), where massive sandstones are only recorded in the top-most 570m of Makgabeng strata, and are absent from the lower-most 280m. This overall increase in precipitation towards the top of the Makgabeng formation is also recorded in the increasingly common transition from inferred interdune deposits to playa lake deposits (discussed above). At this period in geological history, where land plants were generally absent, even high rates of periodic precipitation might not necessarily lead to the cessation of aeolian sedimentation. Increasingly wet weather might also explain the gradual dominance of barchan dunes over transverse dunes towards the top of the Makgabeng Formation. Barchan dunes generally form rather than transverse dunes under dwindling sediment supply (McKee, 1979). If rainfall rates increased, more sand might have been removed from source areas by fluvial action which bypassed the erg, rather than being retained as aeolian bedforms.

The facies present within the pebbly sandstone facies association are difficult to reconcile with any aeolian process. The relatively large quartz pebbles were not recorded in any other facies association in the Makgabeng Formation, though quartz grains included in the generally massive sandstone of this facies association are identical to those associated with the dune deposits. This suggests that dune sand has

been reworked into this facies, and that quartz pebbles were transported into the erg. The presence of small-scale trough and planar cross-bedding, preserved channel forms and parting lineations on planar-bedded and laminated sandstone surfaces indicate the migration of sub-aqueous dune and bar bedforms, channel scouring, and upper flow regime plane-bed flow. Thus, the features of this facies association are consistent with those of ephemeral rivers (Miall, 1996), which may have flowed across the Makgabeng desert. In particular, the presence of plane-bed laminated sand with parting lineations is indicative of high-velocity, flashy discharge which is a characteristic of ephemeral streams (Miall, 1996). Quartz and quartzite pebbles could not have been brought into the desert by aeolian processes, and thus fluvial processes are considered likely. The trend of parting lineations and channels, and the dip-direction of foresets suggest that these ephemeral rivers flowed southwards. This palaeocurrent direction matches those from the Setlaole Formation beneath, and it seems that both these formations were deposited when there was an area of higher relief to the north from which rivers drained.

The widespread evidence for an increasingly wet palaeoclimate recorded in the upper strata of the Makgabeng Formation also provides evidence for the cessation of aeolian conditions. Increasing precipitation may initially have caused the transition from transverse to barchan dunes, and increasingly common ephemeral rivers may have removed material away from aeolian source areas altogether, thus ending deposition in the Makgabeng sand sea.

8.2.4: Depositional setting of the Mogalakwena Formation:

The conglomerates and interbedded trough cross-bedded sandstones and granulestones of the Mogalakwena Formation (Facies Gmm, and St) are grouped together in sheet-like architectural elements (CHS; Miall, 1992; Table 3.1), and locally, conglomerate filled channels occur (architectural element CHR; Miall, 1992; Table 3.1; Section 4.4.1.1). Trough cross-bedding, preserved in the sandstone and granulestone of the Mogalakwena Formation implies migrating aqueous dunes (Table 8.2a) and the ubiquitous presence of architectural element CHS is compatible with the

migration of braided fluvial channels (Miall, 1992; Table 8.2b). Unimodal palaeocurrent directions recorded from the Mogalakwena Formation (Figures 4.55, 4.61, 4.70) suggest that rivers were flowing from the N.E. and E.N.E, and this palaeocurrent direction is reflected by a facies change towards the S.W.: The dominant conglomeratic lithofacies in the vicinity of the southern strand of the Melinda Fault (Section 4.4.1.1) become subordinate towards the S.W., and more distal outcrops in the southwestern portion of the study area are dominated by trough cross-bedded sandstone and granulestone, locally with heavy mineral concentrations developed on foresets (Section 4.4.1.2). The conglomerate-dominated facies in the east can therefore be regarded as proximal, and the trough cross-bedded sandstone and granulestone in the west can be regarded as distal facies equivalents.

The Mogalakwena Formation is developed considerably farther northwards than any of the older formations of the Waterberg Group (Section 4.4.2). The basal pebble and cobble conglomerates are generally compositionally (quartz, quartzite and banded iron formation cobbles and pebbles) and texturally mature, suggesting that the sediment had travelled some distance. The conglomerates appear to be most fully developed at the base and close to the southern strand of the Melinda Fault in the southern foothills of Blouberg mountain, so it is likely that this fault exerted a controlling factor over the deposition of the conglomerates. Conglomerates to the north of the southern strand of the Melinda Fault are relatively thin (max 10m)(Section 4.4.2), similar to those developed in distal areas in the S.W. of the study area. The dominant trough cross-bedded sandstones north of the southern strand of the Fault, characterised by heavy mineral concentrations on foresets, are also identical with the facies from the distal outcrops of the Mogalakwena Formation in the S.W. This suggests that basal conglomerates of the Mogalakwena Formation thin northwards, and younger, distal facies equivalents are more fully developed in this area to the north of the southern strand of the Melinda Fault. This implies that the Mogalakwena Formation overlapped northwards over the southern strand of the Melinda Fault. The brittle northwards-dipping normal faults identified along the southern strand of the Melinda Fault (Section 8.1.3) appear to have juxtaposed the distal facies from the top part of the Mogalakwena strata in the hanging wall against

the basal conglomeratic facies in the footwall to the south after the end of Mogalakwena deposition.

Considered as a whole, the strata of the Waterberg Group can be seen to gradually encroach northwards over time, as successive formations appear generally to have their preserved margins established progressively further to the north, ultimately onlapping over the southern strand of the Melinda Fault. This would suggest that an area of high topography established to the north, approximately over the Central Zone of the Limpopo Mobile Belt, was gradually being peneplaned throughout the Waterberg deposition. Such a model is qualified by the general veering of fluvial palaeocurrent directions from southwards (Setlaole and Makgabeng Formations), i.e. directly away from the inferred palaeohigh, to southeastwards (Mogalakwena Formation), i.e. obliquely away from a less developed topography.

8.2.5: Depositional setting of the Wyllies Poort Formation:

The Wyllies Poort Formation appears to be comprised of two dominant facies associations; trough cross-bedded sandstones with pebble washes developed locally, and planar bedded sandstones with ripplemarks and very low-angled cross bedding developed locally. Both these facies associations are developed in sheet-like architectural elements, and are inferred to have developed in braided fluvial conditions (Miall, 1992; Table 8.2b). However, these two different facies associations contain evidence for contrasting palaeocurrent directions. Palaeocurrent directions recorded from trough cross-bedded sandstone generally suggest flow towards the S.W. (sub-parallel to the trend of the southern strand of the Melinda Fault and the Palala Shear Zone). Asymmetric ripplemarks, in contrast, suggest that weak currents flowed from the N.W and S.E. Similarly, symmetric ripplemarks suggest small -scale wave action from the N.W or S.E. Ripplemarks recorded from the northeastern foothills of Blouberg mountain suggest a main flow from the south, with a minor flow from the north.

The fact that no outcrops of any Soutpansberg strata have been identified south of the projected line of the Melinda Fault, whilst outcrops are common to the north, suggests that the Soutpansberg strata may have been deposited in a half-graben environment, with the hanging wall to the north. Leeder and Gawthorpe (1987) state that one of the criteria for recognising sedimentation in a half-graben environment is that contrasting facies contain contrasting palaeocurrent directions. It is possible that the approximately southwards-flowing palaeocurrents (Figure 5.19) (the planar-bedded sandstones with ripplemarks and locally developed low-angled cross-beds) represent fast-flowing water under upper flow regime conditions, down the gently-dipping hanging wall side of the half-graben. Large- and small-scale trough cross-beds are likely to represent reworking of sediment by aqueous dunes along low-sinuosity, longitudinally-flowing trunk rivers, which flowed parallel to the half-graben axis.

Graben-type models have been proposed earlier for the Soutpansberg Group (e.g. Jansen, 1975; Barker, 1976), though they were refuted by Cheney *et al.*, (1990), who interpreted the Soutpansberg Group as representing a remnant of a once craton-wide cover sequence (i.e. the Soutpansberg Group was *preserved in*, rather than *deposited in* the Soutpansberg trough). One of the arguments against a Soutpansberg depository developed along the Palala Shear Zone put forward by Cheney *et al.* (1990), was that no fault had been identified to the south of the Soutpansberg mountains which could have acted as a southern graben margin. Additionally, several outcrops of Soutpansberg strata had been identified further north than Jansen's (1975) proposed northern margin to the Soutpansberg graben.

Evidence presented here from the Blouberg area suggests that the southern margin of the Soutpansberg Group may be bound by the northward-dipping normal faults along the southern strand of the Melinda Fault, as demonstrated in Sections 8.1.3 and 8.2.4. This normal fault is thought to have displaced the Mogalakwena strata as a response to approximately north-south orientated extension that accompanied the intrusion of the Sibasa feeder dyke swarm, and may have acted as a southern-bounding fault to the Soutpansberg half-graben.

8.3: Provisional basin evolution model for the study area:

During this section, the interpretations of the structural setting and the sedimentary palaeoenvironments which have been described above, will be considered together to produce a provisional model for basin evolution over time of the study area around Blouberg mountain.

The Palala Shear Zone appears to be a factor affecting both the structure and the sedimentation of the study area. Faults (e.g. the Melinda Fault Zone and other reverse faults or thrusts) appear to have formed above the Palala Shear Zone, and have a parallel strike. The extent of all of the sedimentary units discussed in this work appears to be at least partially controlled by structures developed along the Palala line. Palaeocurrent directions in the sedimentary strata are often either parallel to (e.g. the Mogalakwena Formation and the trough cross-bedded facies of the Wyllies Poort Formation) or perpendicular to (e.g. the Setlaole Formation and the ephemeral river facies of the Makgabeng Formation) the Palala line. Therefore it is likely that the geological evolution of the study area has been mainly controlled by periodic reactivation along the suture between the Kaapvaal Craton and the Central Zone of the Limpopo Mobile Belt.

The generally sinistral strike-slip sense of movement recorded in the banded gneiss in the basement is likely to have formed in the collision between the Kaapvaal Craton and the Central Zone of the Limpopo Mobile Belt. The highly ductile deformation of this transpressional event is in contrast to all the younger, brittle deformation, which has affected the generally unmetamorphosed sedimentary rocks in the overlying strata. The fact that sedimentary strata were deposited over such high-grade metamorphic rocks in the field area, indicates that a considerable period of time had elapsed between collision and the onset of Blouberg sedimentation, for such deep crustal rocks to have been exhumed prior to sedimentation. The following paragraphs consider the basin evolution and tectonic history of the Blouberg area following the exhumation of the basement gneiss. This history is also shown schematically in Figure 8.6.



The extent of the Blouberg Formation appears to be restricted to basins developed only above the Palala Shear Zone, and evidence from the Kranskop area suggests that, whilst limited in aerial extent, the basin may have subsided rapidly (at least 1400m of immature sedimentary rocks are preserved) prior to basin inversion and subsequent erosion. The general immaturity, varying palaeocurrent directions and apparently variable palaeohydraulic inferences, and the fact that these rocks are developed over a major transpressional shear zone, show that the strata of the Blouberg basin have many characteristics in common with those associated with pull-apart basins (Figure 8.6a). No strike-slip faults were identified which may have bounded a Blouberg pull-apart basin, though it is likely that any such faults would have been reactivated several times in the subsequent deformational history, and kinematic indicators were likely to have been destroyed. It is therefore not possible to identify whether such a postulated pull-apart basin may have formed as a response to dextral or sinistral strike-slip movement.

The southwards-vergent basin inversion affecting the Lower Member of the Blouberg Formation indicates that a compressional regime, orientated approximately perpendicular to the previously transpressional Palala Shear Zone (Figure 8.6b), affected the area. This had the effect of overturning the Lower Member of the Blouberg Formation, locally thrusting basement over the lower Blouberg strata, and uplifting the basement source area to the north, leading to vigorous erosion and sedimentation of the Upper Member of the Blouberg Formation, in small localised depositories.

A corollary of southwards-vergent thrusting is that the crust to the north is likely to have been uplifted and thickened, so it is probable that this event produced a highland area north of the Palala Shear Zone in the area occupied by rocks of the Central Zone ('Limpopo mountains' in Figure 8.6c). The evidence for the presence of such a highland area is reflected in the sedimentary record; the Setlaole and Makgabeng Formations do not appear to have been deposited to the north, and southerly directed fluvial palaeocurrent directions in these strata suggest that rivers flowed off these



mountains. The palaeowind direction recorded from cross-bedding in the Makgabeng palaeo-erg generally blew towards the S.W. and W.S.W., suggesting that winds may have been funnelled along the Limpopo mountain ridge. Of the Waterberg strata, only the upper part of the Mogalakwena Formation encroached northwards over the denuding Limpopo mountains. This onlap relationship is exhibited by thin basal conglomerates which are generally restricted to north-south trending palaeovalleys, and which are overlain by distal coarse sandstone and granulestone, indicating that the source area of the Limpopo mountains had retreated northwards and eastwards as Mogalakwena sedimentation progressed. The lack of coarse clastic material in the upper and distal parts of the Mogalakwena strata is indicative of peneplanation of the source area to the north and east. It is difficult to reconcile the unimodal westwards and southwestwards palaeocurrent directions recorded in the Mogalakwena Formation with the apparent northwards onlap of the Mogalakwena strata over the southern strand of the Melinda Fault, and the rarely inferred presence of north-south trending palaeovalleys. It seems likely that the Mogalakwena rivers may have flowed obliquely (and sub-parallel) to the edge of the mountains to the north, with the edge of the mountain land defined by the southern strand of the Melinda Fault (Figure 8.6c).

Together, the Blouberg and Mogalakwena Formation can be thought of as representing a flysch (syn-tectonic) and molasse (post-tectonic) sequence, respectively, developed as a result of orogenic reactivation within the Limpopo Mobile Belt, which created the Limpopo mountains (Figure 8.6c). Such a model, however, must be reconciled with the fact that no basement-derived gneiss clasts have been reported in the conglomerates of the Mogalakwena Formation. In contrast, however, clasts of gneiss have been reported in the basal Tshifhefhe Formation of the Soutpansberg Group (Brandl, 1987). This can be explained by considering different sediment maturity between these two stratigraphic units. Clasts within the Waterberg Group are likely to have been transported a considerable distance, so that foliated gneiss clasts derived from a Limpopo source are unlikely to have survived transport. Clasts within the basal Soutpansberg strata are likely to have been transported only a short distance, possibly from proximal fault scarps bounding a half-graben, and so remain intact within conglomeratic beds.



The Soutpansberg Group can be interpreted to have been deposited as a response to renewed tectonism, though in a generally extensional setting, in contrast to the earlier, syn-Blouberg, compressional regime (Figure 8.6d). The orientation of feeder dykes to the Sibasa Formation suggests a generally north-south orientated extensional regime, and it is likely that the previously imposed, northward-dipping, syn-Blouberg reverse faults located along the southern strand of the Melinda Fault, were reactivated as normal faults. These normal faults locally juxtaposed the upper, distal strata of the Mogalakwena Formation in the hanging wall against the basal conglomeratic strata in the footwall of the southern strand of the Melinda Fault (Figure 8.6d).

In order to accomplish this juxtaposition of the contrasting facies of the Mogalakwena strata, a considerable displacement is inferred along the Melinda Fault. Downward displacement of the hanging wall is thought to have created a depository above, which was fault bound only to the south (i.e. a half-graben). It is this depository in which the Soutpansberg Group is thought to have been laid down, initially with the only locally developed lava of the Sibasa Formation, and subsequently with the sandstones of the Wyllies Poort Formation (Figure 8.6d). The general maturity of the Wyllies Poort Formation may be a reflection of the considerable distance of transport across the low-angled palaeotopography in the hanging wall. The changes in facies and palaeocurrent directions recorded in the Wyllies Poort Formation are thought to represent transverse and longitudinal fluvial regimes within the half-graben environment.

Evidence for the most recent reactivation above the Palala Shear Zone was recorded from the post-Wyllies Poort northern strand of the Melinda Fault, which may be a dextral strike-slip fault (Figure 8.6e). It proved impossible to further constrain the timing of this reactivation, though in other areas the Melinda Fault is thought to cut the strata of the Karoo Supergroup (Brandl, 1986). Therefore a Phanerozoic timing for this reactivation may be appropriate.

Overall, the sedimentary basins in the area around Blouberg mountain can be considered to have been created as a response to a three-stage tectonic sequence, of



north-south orientated compression (Blouberg Formation), tectonic quiescence and denudation (Waterberg Group), and north-south orientated extension (Soutpansberg Group). Age constraints available from the literature which have bearing on this proposed tectonic history are not very precise. Cheney *et al.* (1991) suggest an age of 1.8 Ga - 1.9 Ga for the Sibasa basalt (whole rock analysis). A corollary of this is that all the Waterberg-aged rocks, generally regarded as being about 1.6 Ga-1.7 Ga (Callaghan *et al.*, 1991), may be several hundred million years older, as they have been shown to pre-date the Soutpansberg rocks. The fact that all sedimentary strata discussed in this work are red beds, indicates that they were all deposited and lithified after c. 1.9 - 2.0 Ga (Eriksson and Cheney, 1992).

The minimum age of ductile movement along the Palala Shear Zone was gained from the age of a cross-cutting pseudotachylite vein, which showed that movement ceased at c. 2042 Ma (Brandl and Reimold, 1990). It therefore seems that the events proposed in this chapter may have occurred between about 2.0 Ga and 1.8 Ga. Such a time constraint is difficult to reconcile with the age of ~1700Ma for the middle Waterberg Group (Makgabeng Formation) proposed by Jansen (1976), which was based on the fact that the Abbotspoort Fault in the Palala area, which does not displace the Makgabeng Formation, *does* displace Palala granite, dated at 1770 ± 30 Ma. A similar age of 1725 ± 245 Ma for the Palala granite was reported by Barton and McCourt (1983), though they indicate that this may be a reset age. Walraven *et al.* (1990) suggest that the Palala granite is most likely of Bushveld age (~2050 Ma), thus only constraining the deposition of the Waterberg strata to a post 2050 Ma age.

The outcome of current opposing opinions regarding the age of the Limpopo collisional event (whether it occurred at 2.65 Ga or 2.0 Ga) is of obvious relevance to this work. If a c. 2.0 Ga date for collision along the Palala Shear Zone is considered (Kröner *et al.*, 1999), there appears to be little time for exhumation of the banded gneiss, the subsequent deposition of the various sedimentary units, and the intervening tectonic events. However, an older (2.65 Ga) age for the collision allows sufficient time for exhumation, and on this basis it may be postulated that the brittle, syn-Blouberg basin inversion and flysch deposition may be related to the c. 2.0 Ga



(Limpopo reactivation) event itself. Thus the history proposed here would tend to favour a model with 2.65 Ga collision, and a 2.0 Ga reactivation (such as that proposed by McCourt and Armstrong, 1995), where reactivation at 2.0 Ga occurs possibly as a result of the Magondi Orogeny (part of the Africa-wide Eburnean event), or the Vredefort structure, or the emplacement of the Bushveld Complex (Mccourt and Armstrong, 1998).

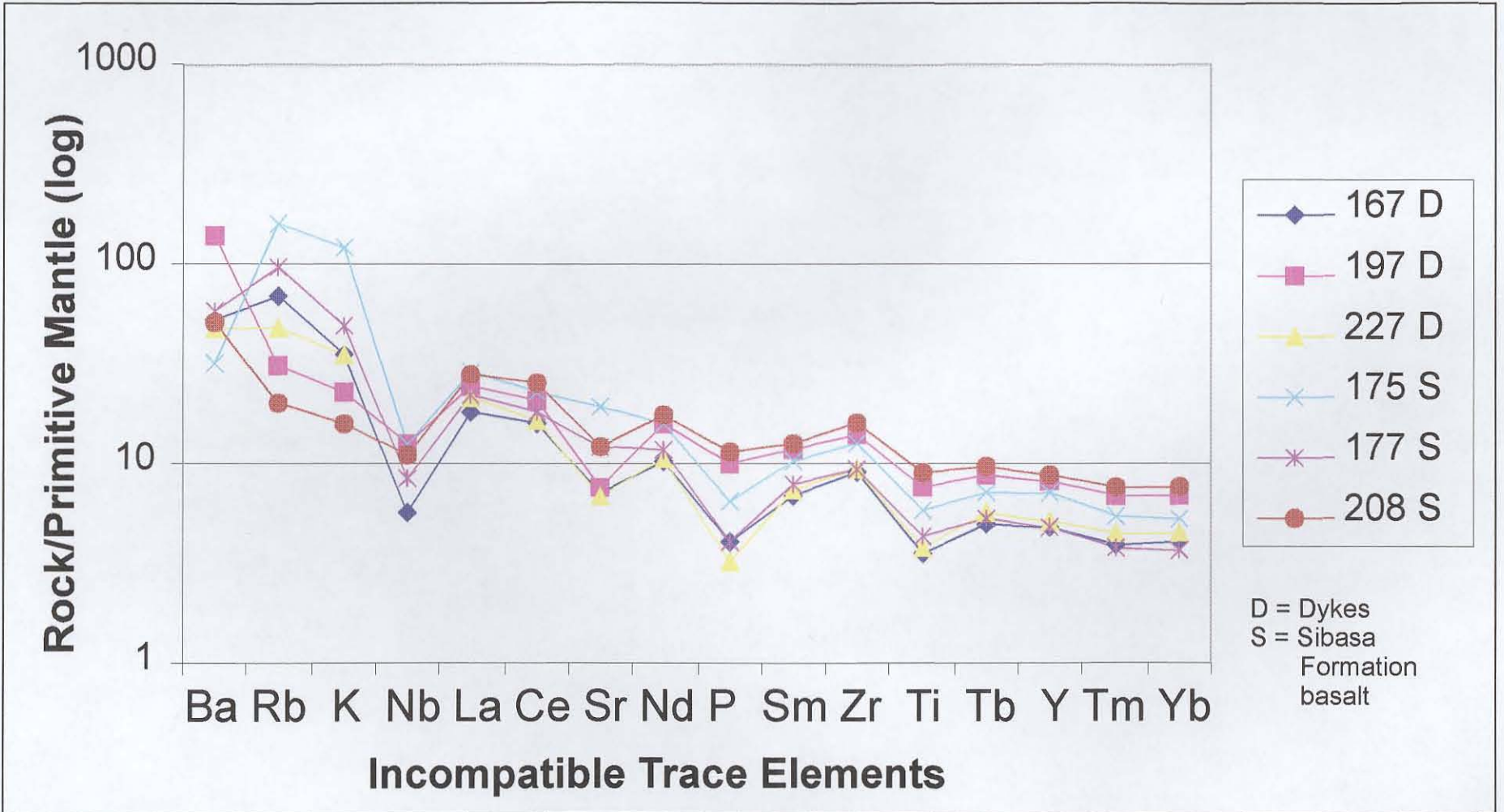


Figure 8.1: Spidergram to compare values of normalised incompatible trace elements for dykes (D) and the Sibasa Formation (S).

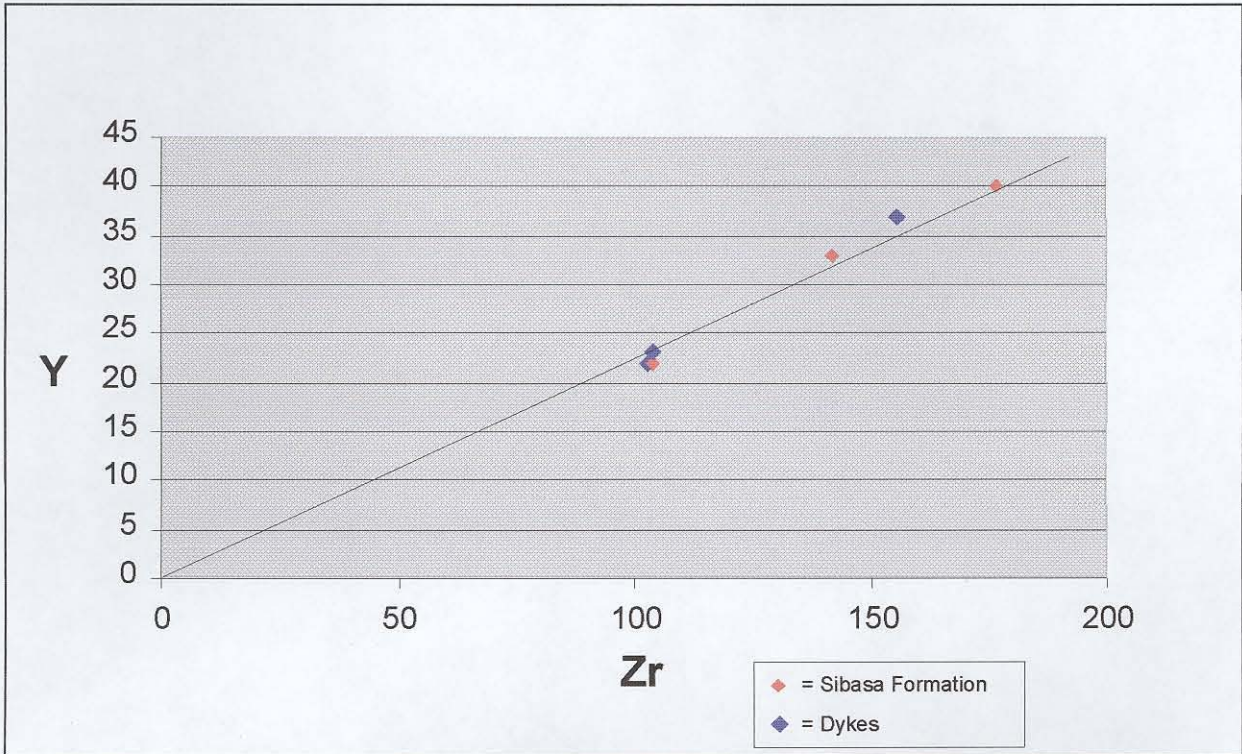


Figure 8.2: Variation diagram between Zr and Y for the Sibasa basalts and dykes intruding the study area, suggesting a common parental magma.

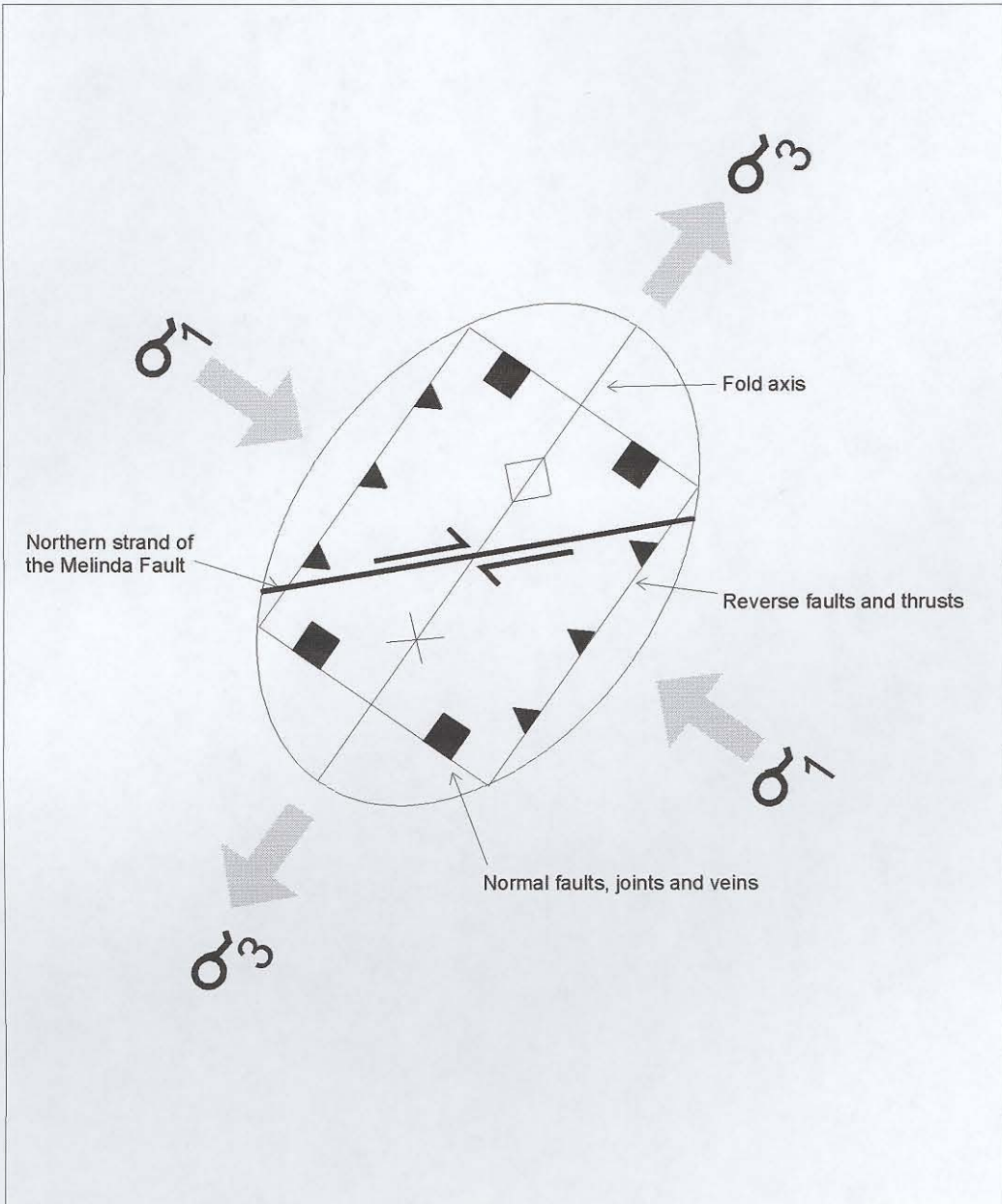


Figure 8.3: Strain ellipse which accounts for the structures recorded in the Wyllies Poort Formation adjacent to the northern strand of the Melinda Fault and its splays.

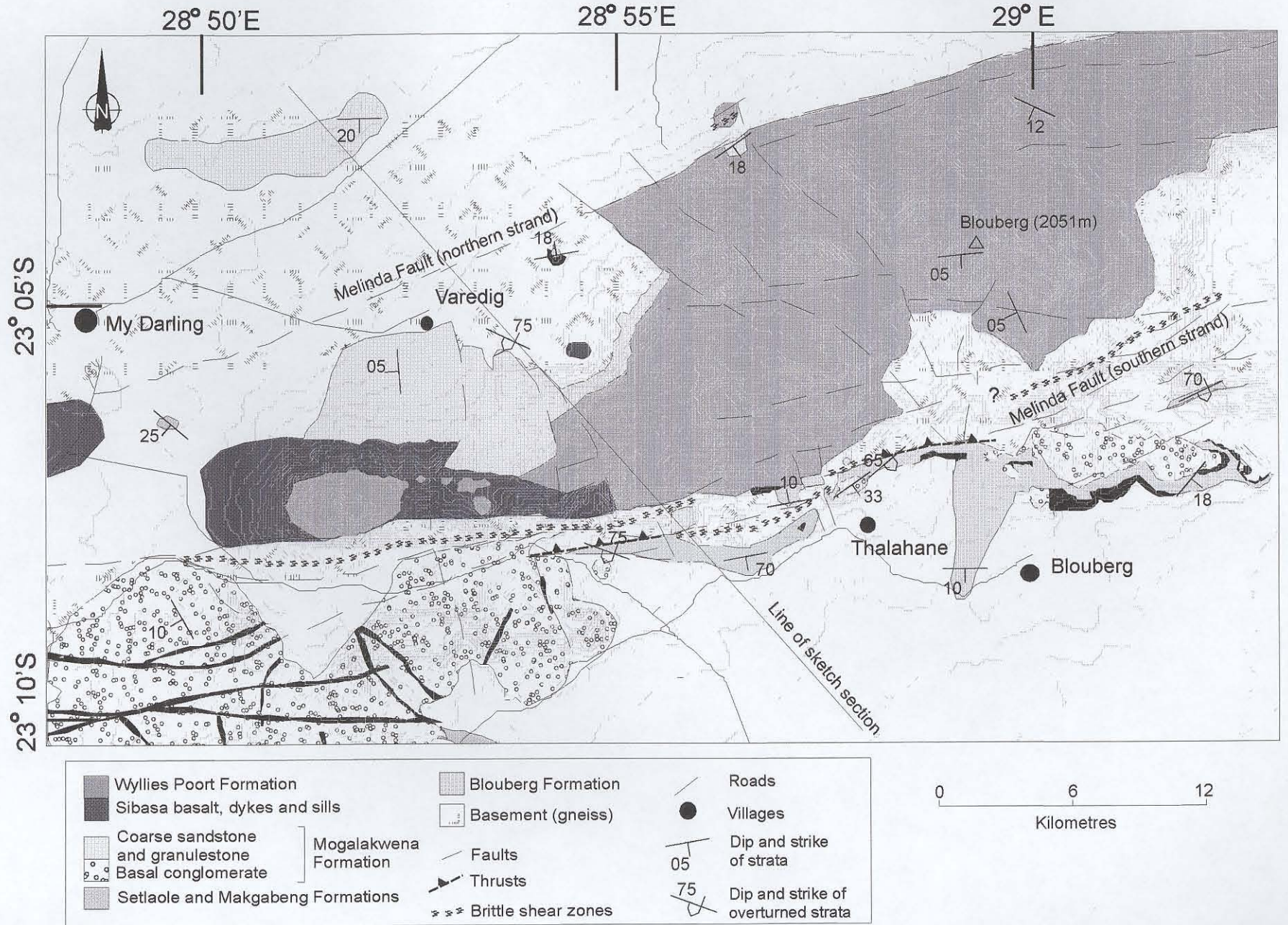


Figure 8.4: Map showing approximate line of idealised cross-section shown in Figure 8.5.

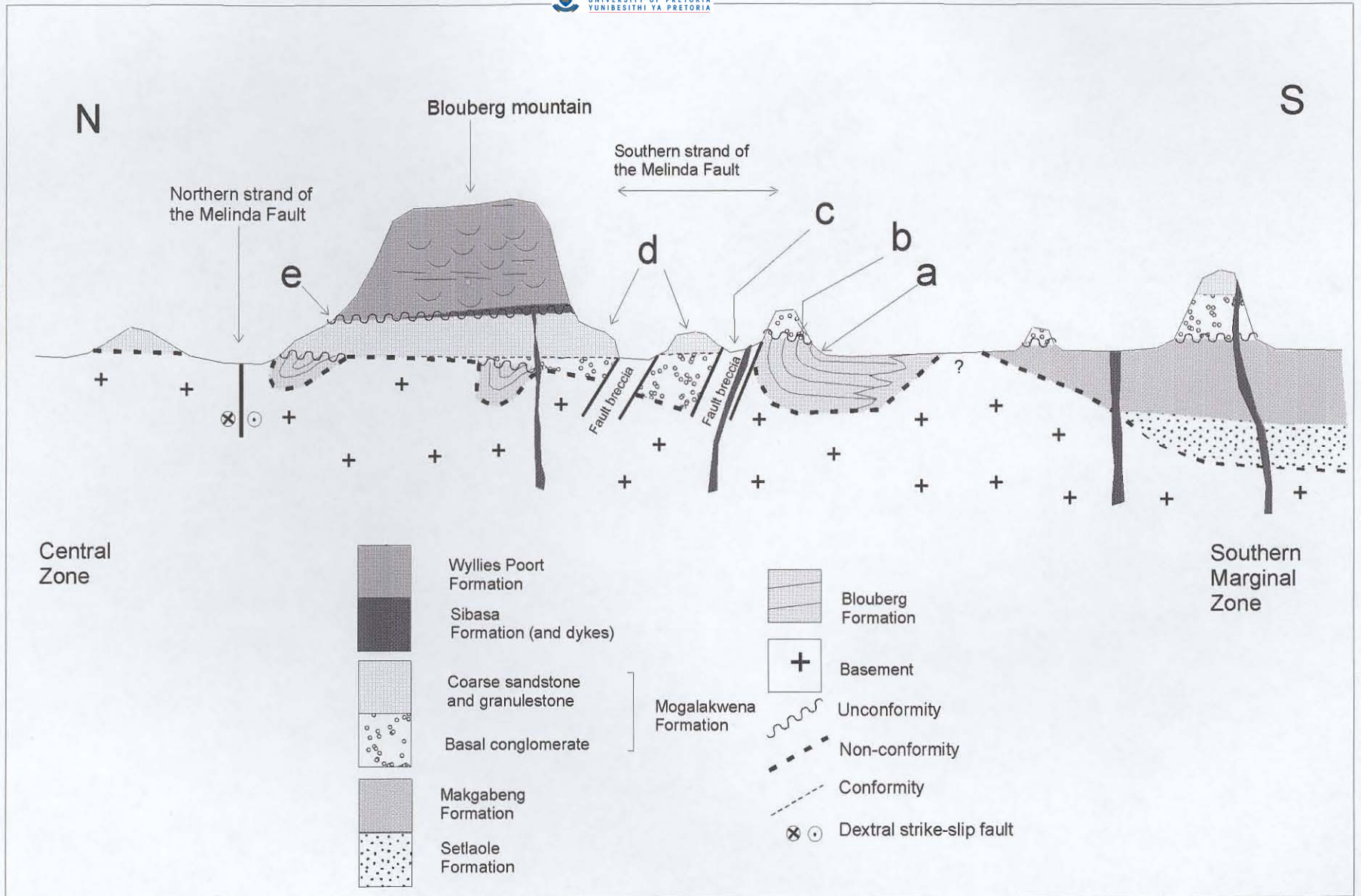
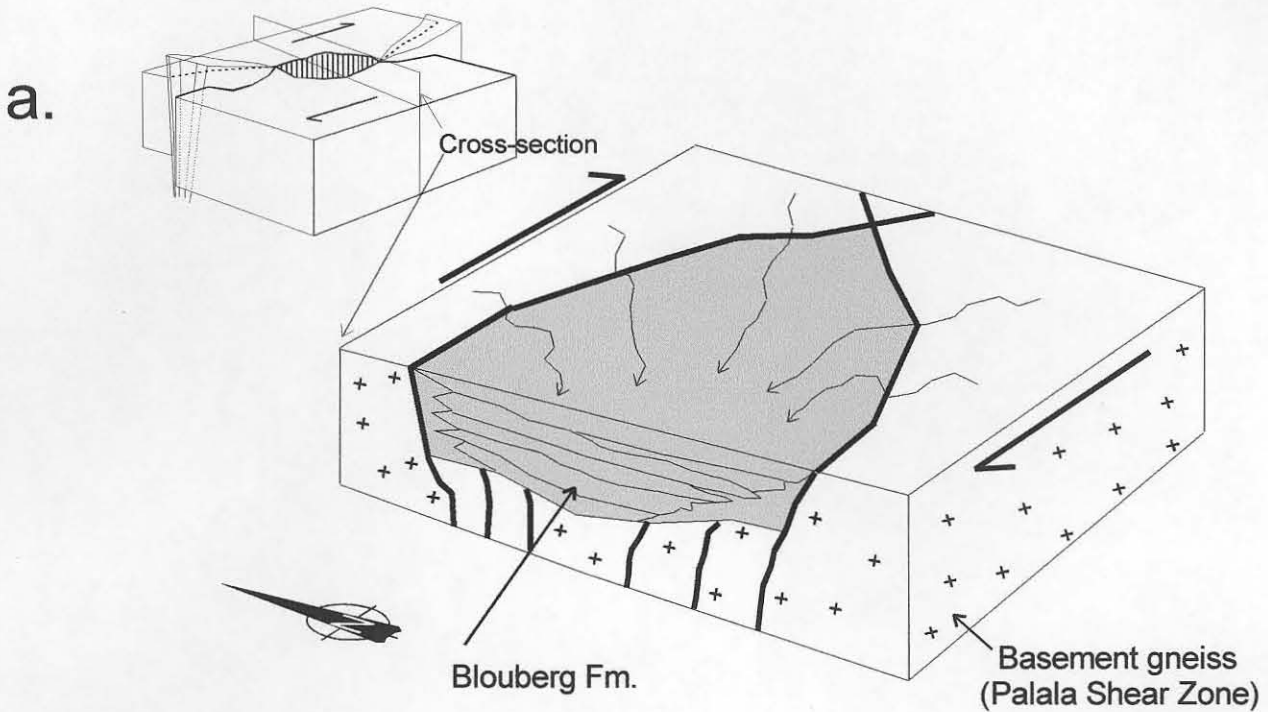
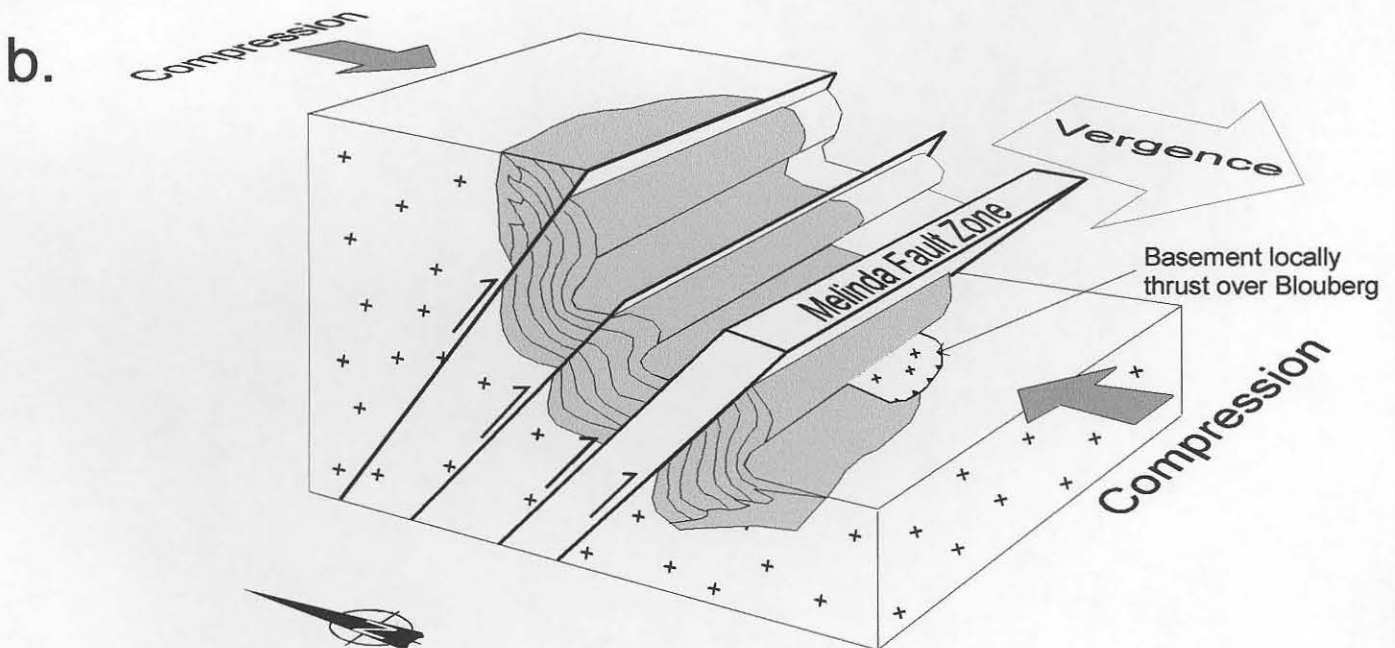


Figure 8.5: Idealised cross-section through Blouberg mountain, showing the structural relationships proposed in Chapter 8.

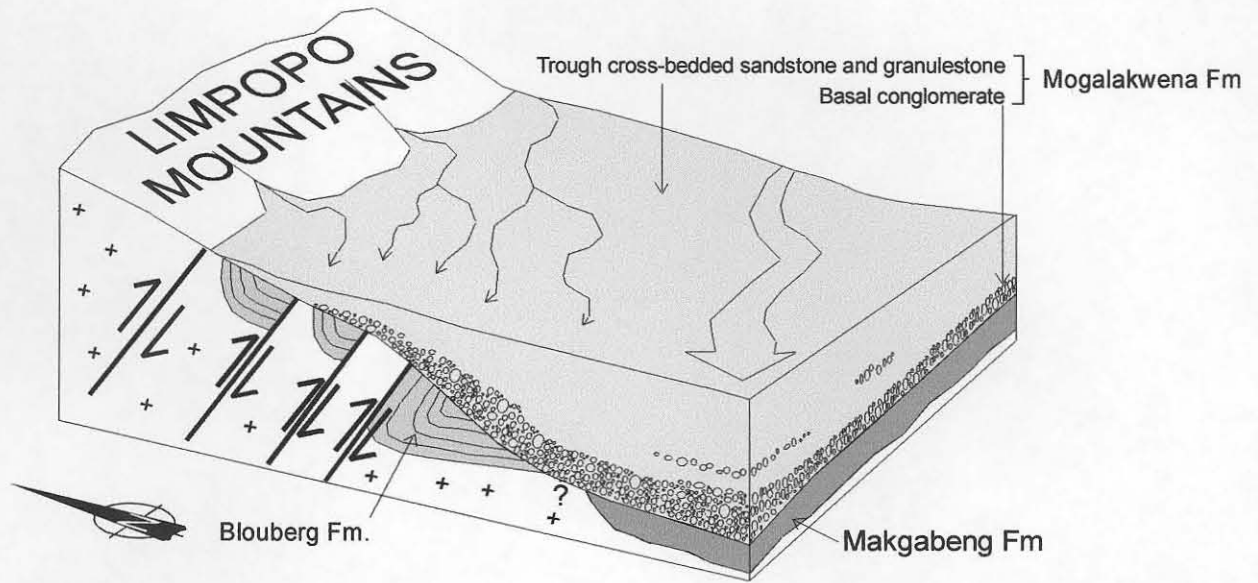


Blouberg Formation (Lower Member) sedimentation in deep, localised depositories, possibly reflecting pull-apart basins, caused by strike-slip reactivation along the Palala Shear Zone.



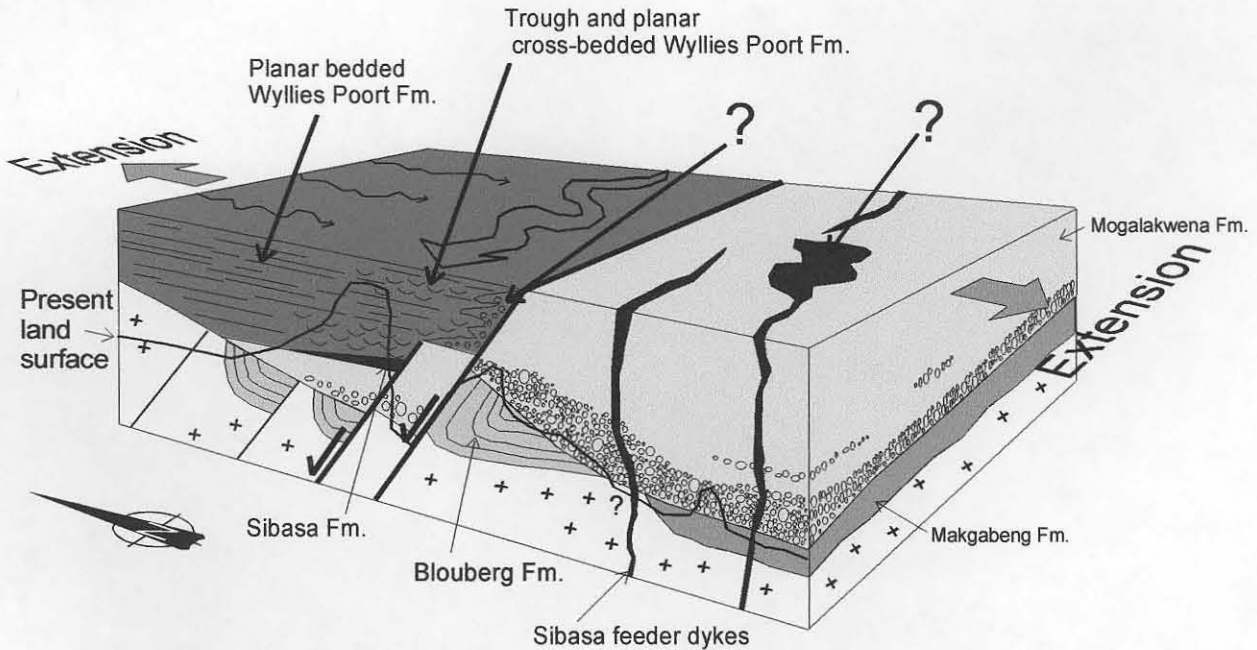
Lower Member of Blouberg Formation is folded, locally overturned and faulted during southwards-vergent, N-S orientated compression. Reverse faults dip to the north. Upper Member is deposited locally thereafter.

C.



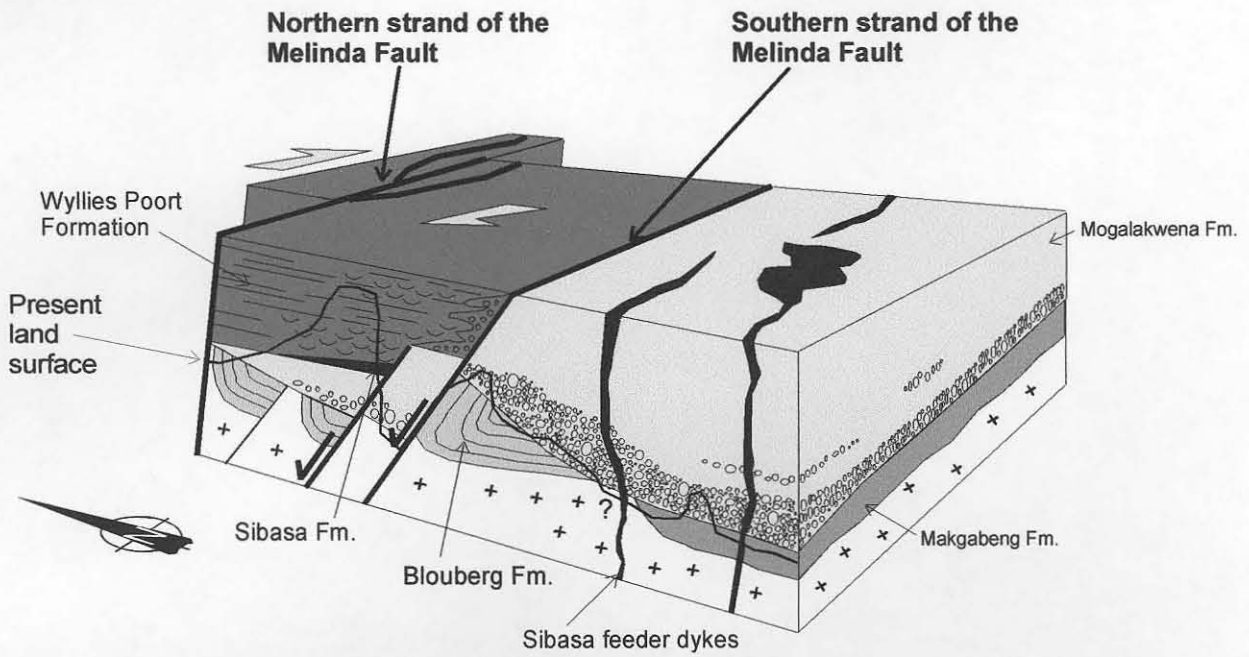
Mogalakwena conglomerates, granulestones and sandstones lain unconformably on the Blouberg Formation and nonconformably on the basement.

d.



Melinda Fault reactivated as normal faults in an extensional regime, creating a half-graben for the Soutpansberg Group depository. The laterally extensive Wyllies Poort Formation is deposited unconformably over the Mogalakwena Formation, and nonconformably over the basement.

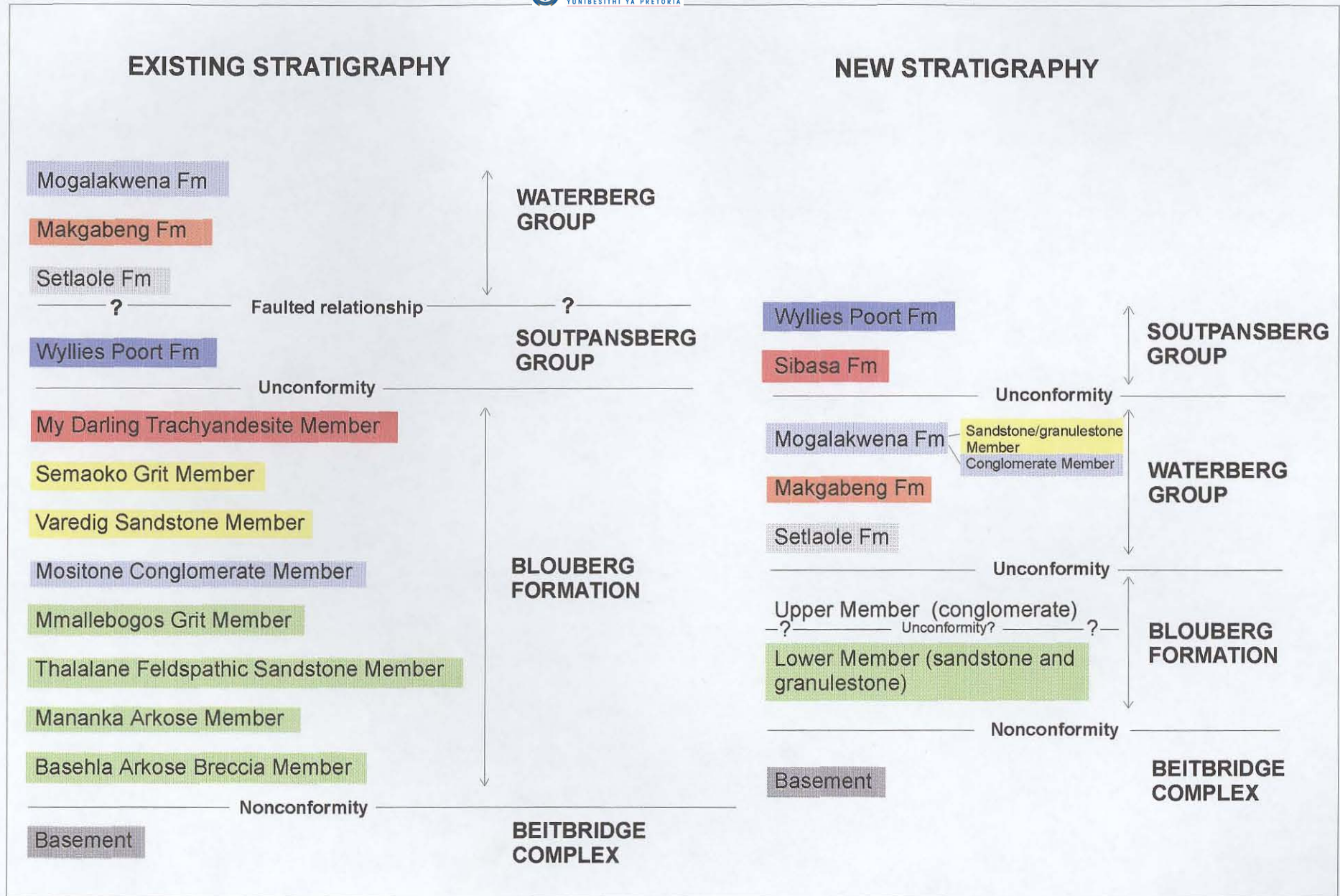
e.



Northern strand of the Melinda Fault activates as a dextral strike-slip fault, and displaces the Soutpansberg strata, probably during the Phanerozoic.

Figure 8.6: Block diagrams illustrating the proposed model for the tectonic and sedimentary evolution of the Blouberg area.

Table 8.1: Proposed new stratigraphic subdivision of the study area. Coloured boxes illustrate correspondence between the existing (Jansen, 1976; Meinster, 1977) stratigraphy and the proposed new stratigraphy.





a.	Facies Code	Facies	Sedimentary structures	Interpretation
	Gmm	Matrix-supported, massive gravel	Weak grading	Plastic debris-flow (High strength, viscous)
	Gmg	Matrix-supported gravel	Inverse to normal grading	Pseudoplastic debris flow (Low strength, viscous)
	Gci	Clast-supported gravel	Inverse grading	Clast-rich debris flow (high strength) or pseudoplastic debris flow (low strength)
	Gcm	Clast-supported massive gravel	-	Pseudoplastic debris flow (inertial bedload, turbulent flow)
	Gh	Clast-supported, crudely bedded gravel	Horizontal bedding, imbrication	Longitudinal bedforms, lag deposits, sieve deposits
	Gt	Gravel, stratified	Trough cross-beds	Minor channel fills
	Gp	Gravel, stratified	Planar cross-beds	Transverse bedforms, deltaic growths from older bar remnants
	St	Sand, fine to very coarse	Solitary or grouped trough cross-beds	Sinuuous-crested and linguoid (3D) dunes
	Sp	Sand, fine to very coarse may be pebbly	Solitary or grouped planar cross-beds	Transverse and linguoid bedforms (2D dunes)
	Sr	Sand, fine to very coarse	Ripple cross-lamination	Ripples (Lower flow regime)
	Sh	Sand, fine to very coarse, may be pebbly	Horizontal lamination parting or streaming lineation	Plane-bed flow (Critical flow)
	Sl	Sand, fine to very coarse, may be pebbly	Low-angle (<15°) cross-beds	Scour-fills, humpback or washed-out dunes, antidunes
	Ss	Sand, fine to very coarse, may be pebbly	Broad, shallow scours	Scour fill
	Sm	Sand, fine to coarse	Massive or faint lamination	Sediment gravity flow deposits
	Fl	Sand, silt, mud	Fine lamination very small ripples	Overbank, abandoned channel, or waning flood deposits
	Fsm	Silt, mud	Massive	Backswamp or abandoned channel deposits
	Fm	Mud, silt	Massive desiccation cracks	Overbank, abandoned channel or drape deposits
	Fr	Mud, silt	Massive, roots, bioturbation	Rootbed, incipient soil
	C	Coal, carbonaceous mud	Plant, mud films	Vegetated swamp deposit
	P	Palaeosol carbonate (Calcite, siderite)	Pedogenic features; nodules, filaments	Soil with chemical precipitation

b.	Element	Symbol	Lithology	Interpretation
	Major sandstone sheet	CHS	Gm, Se, St, Sh, Sl, rare Sp	Braided fluvial channel
	Major sandstone ribbon	CHR	Gm, Se, St, Sh, Sl, rare Sp	Anastomosed fluvial channel
	Minor sandstone sheet	CS	Se, Sh, Sl, rare St, Sr	Sheet splay
	Minor sandstone lens	CRS	Se, St, Sl, Sh, Sr	Minor crevasse channel

Table 8.2a.: Interpretation of facies classifications (after Miall, 1996). b.: Interpretation of architectural elements (based on those of the Cutler Formation, New Mexico) (after Miall, 1996).