4. REGIONAL TECTONIC AND STRUCTURAL SETTING OF THE BUSHVELD COMPLEX

The structural history of the Bushveld Complex directly relates to tectonic processes affecting the Kaapvaal Craton. It is therefore essential to understand these processes and how the regional structures of the Craton influenced the formation and subsequent deformation of the Bushveld Complex. The first part of the chapter deals with the regional tectonic and structural framework of the Kaapvaal Craton. A brief overview is given of all the major structural features found on the Craton, their origin and structural characteristics are discussed. The second part of the chapter focuses on the tectonic setting and structural characteristics of the Bushveld Complex. Only a broad overview is given of the tectonic and structural setting of the Bushveld Complex while a more detailed discussion of the specific structures follow in the next chapter.

4.1 TECTONIC AND STRUCTURAL FRAMEWORK OF THE KAAPVAAL CRATON

The Kaapvaal Craton represents an Archaean continental fragment of a once much larger continent (de Wit et al., 1988; Groenewald et al., 1991). The Kaapvaal Craton's boundaries are somewhat obscured by younger sediments but are reasonably well established (McCourt, 1995) (Figure 4.1). To the north it is bounded by the Limpopo Belt which represents a collisional zone between the Zimbabwe Craton and the Kaapvaal Craton. The eastern boundary is the north-south trending Lebombo monocline which was formed as a response to the break-up of Gondwana around 150 Ma. In the west and south the Kaapvaal Craton is delineated by the Natal-Namaqua belt. This belt mainly represents an accretionary tectonic event during Kibaran (1.2 Ga) times. The formation of the Kaapvaal Craton can be divided into two main periods (de Wit et al., 1992; de Wit and Hart, 1993). The first period is characterized by the formation of the mid-Archaean Kaapvaal Shield (Figure 4.1), during which major periods of intraoceanic tectonics were active between 3.64 and 3.08 Ga. This period marks the development of granitoid-greenstone terrains, and ended with a major pulse of accretionary tectonics around 3.2 Ga (Brandl and de Wit, 1997). The second period is characterized by intra-cratonic tectonics as well as continental-edge processes between 3.1 and 2.65 Ga (Brandl and de Wit, 1997). Major basin development on the Craton as well as

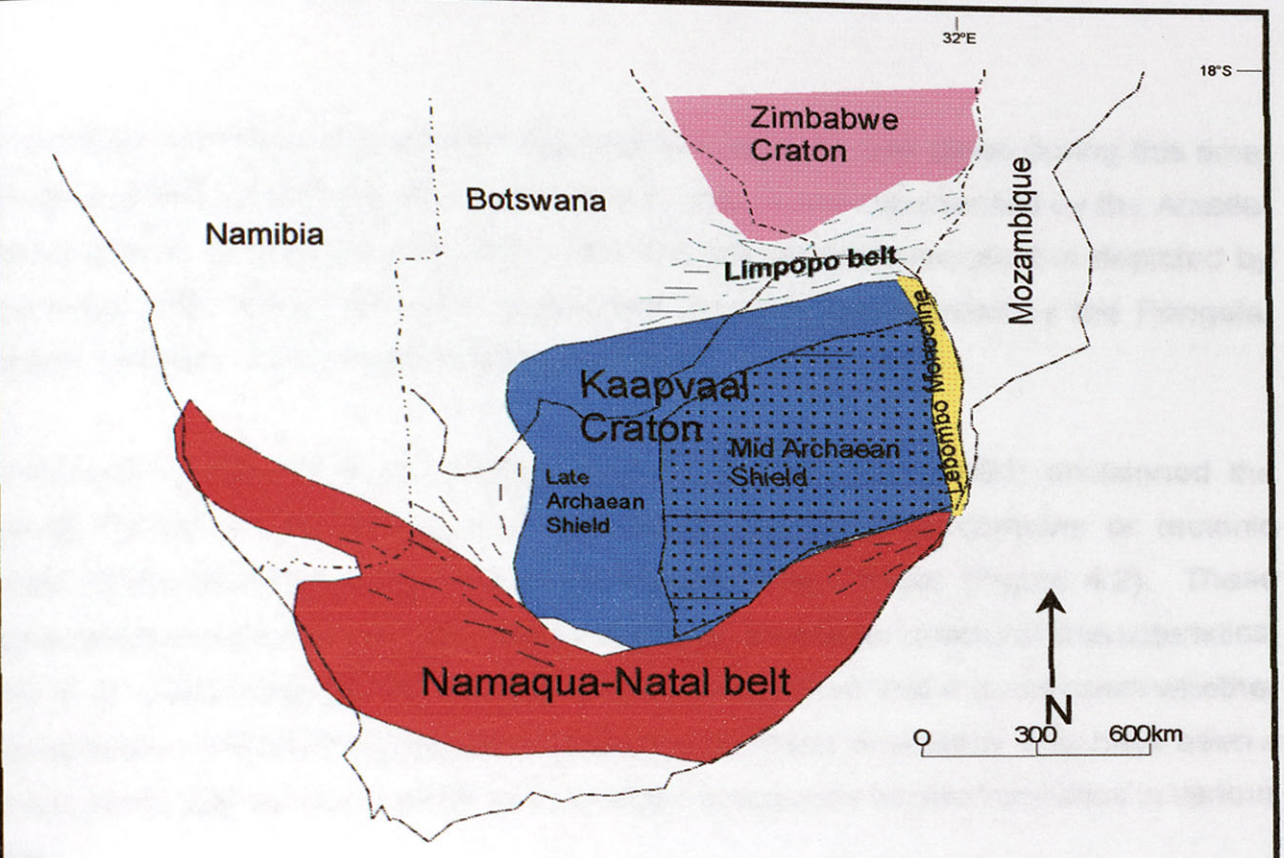

Figure 4.1. Boundaries of the Kaapvaal Craton: Limpopo belt, Lebombo Monocline and Namaqua-Natal belt. (Modified after Thomas et al., 1993)

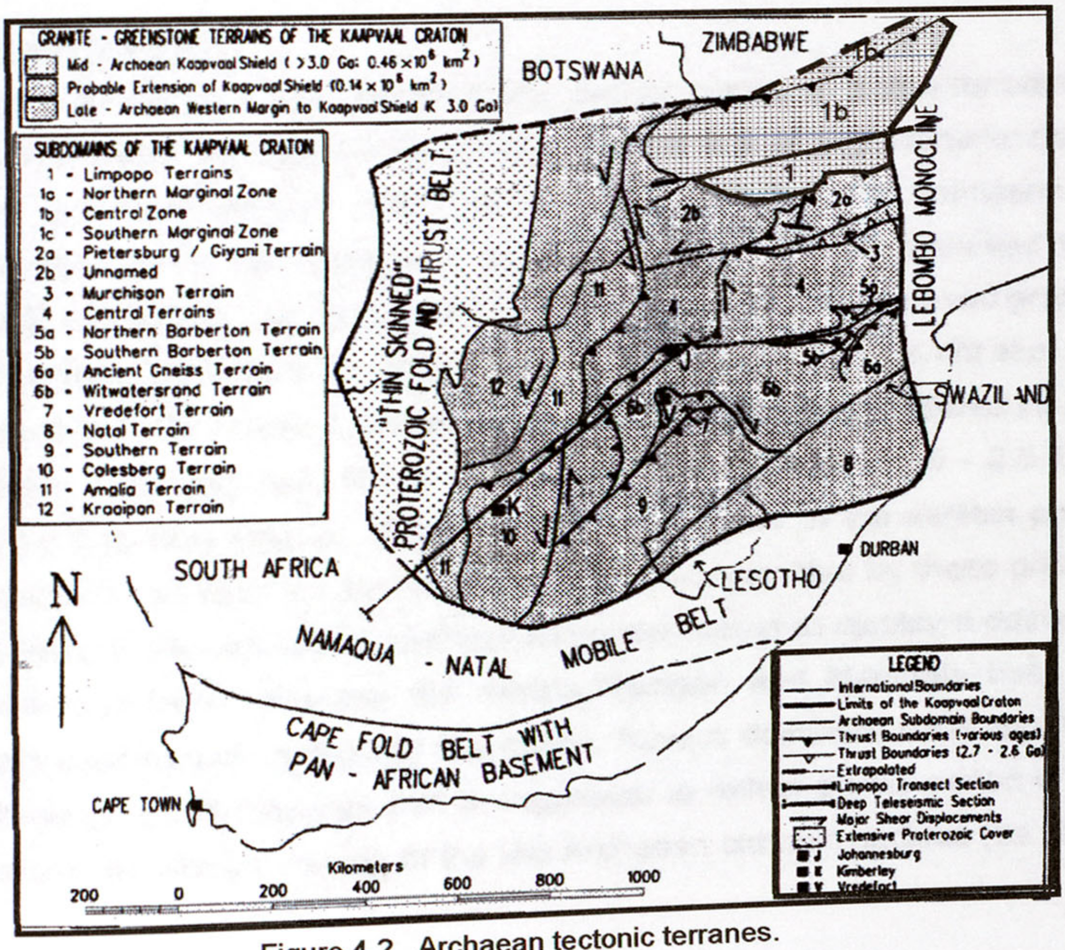

Figure 4.2. Archaean tectonic terranes. (After de Wit et al., 1992)

continental growth along the western and northern margins took place during this time. Continental growth processes along the western margin are represented by the Amalia-Kraaipan granite-greenstone belt, whereas along the northern margin it is depicted by the Limpopo belt. Basin formation during this period is represented by the Pongola, Dominion, Witwatersrand, and Ventersdorp basins.

Authors such as de Wit et al. (1992a) and de Wit and Hart (1993) envisioned the Kaapvaal Craton as consisting of a number of Archaean subdomains or tectonic terranes, which accreted along large ENE trending shear zones (Figure 4.2). These domains are defined based on age, lithostratigraphy as well as structural characteristics. de Wit et al. (1992) and de Wit and Hart (1993) pointed out that it is unknown whether these terranes are allochthonous with respect to each other or whether they have been a coherent geological province which has undergone separate tectonic activities in various regions.

4.1.1 Early Archaean architecture

4.1.1.1 Greenstone belts.

The Archaean greenstone belts represent the earliest period of craton formation. The volcano-sedimentary packages are believed to be the result of "intra-oceanic obduction of hydrated Archaean oceanic crust" (de Wit et al., 1992). The formation of the greenstone belts, along with extensive granitoid emplacement, represented the first continental lithosphere (de Wit et al., 1992). The oldest and best preserved greenstone belt is the Barberton belt which formed between 3.7 and 3.2 Ga ago (de Wit et al., 1992). Other greenstone belts include in the north, the Giyani/Sutherland, Pietersburg (2.8Ga) and Murchison (3.09 Ga) belt, in the west the Amalia-Kraaipan (2.95 - 2.5 Ga) and Marydale (3.0 Ga) belts (Visser, 1998), (Figure 4.3). Some of the earliest prominent structural trends to be seen on the Kaapvaal Craton are depicted by these greenstone belts. The belts in the central and northern Kaapvaal Craton all display a dominant NE to ENE structural trend, whereas the Amalia-Kraaipan and Marydale belts on the western and southwestern margin of the craton, have a dominant NNW to NS trend. The NS trending Amalia-Kraaipan belt is suggested to reflect a later period of terrane accretion along the western margin of the Mid Archaean cratonic nucleus (de Wit et al., 1992). All the greenstone belts have characteristic internal deformation such as steeply dipping schistosities, folding, thrusting, and in some places transcurrent faulting. These internal structural directions generally follow the same regional trends as exhibited by the individual belts.

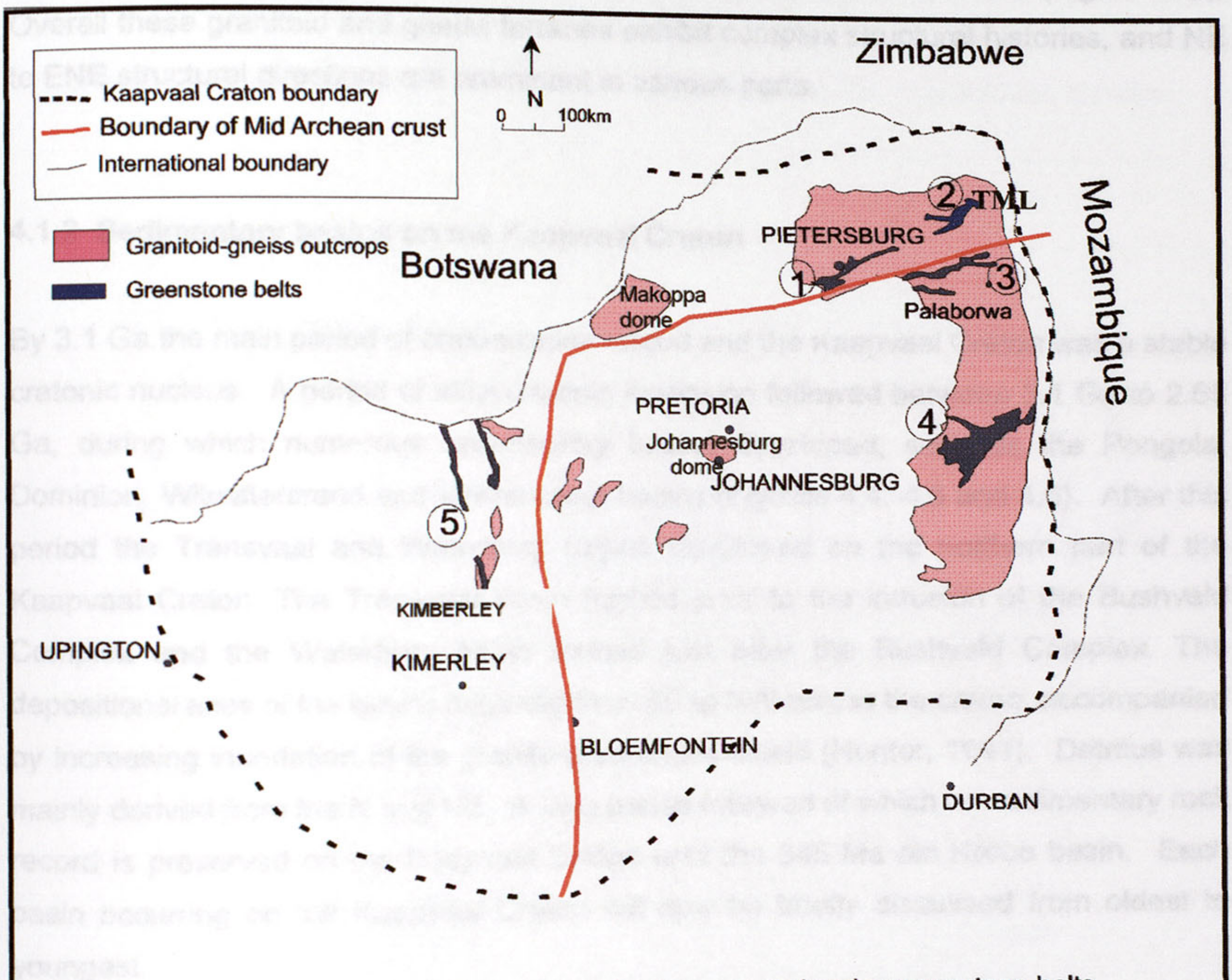

Figure 4.3. Distribution of the granitoid-gneiss terranes and major greenstone belts on the Kaapvaal Craton.
1. Pietersburg, 2. Sutherland, 3. Murchison, 4. Barberton, 5. Amalia-Kraaipan belts.

4.1.1.2 Granitoid terranes

The extensive granitoid terranes with subordinate gneisses exposed in the eastern and northeastern part of the Craton are suggested to represent the early Archaean basement of the Kaapvaal Craton (Tankard et al., 1982). This episode of granitoid emplacement together with greenstone development characterize the first period in the evolutionary history of continental crust and craton formation. Two different opinions exist regarding the formation of these granitoid terranes. Some authors believe that the granitoid terranes developed before the formation of the greenstone belts whereas some authors believe they are intrusive into the greenstone belts (Hunter, 1981). De Wit and Roering (1990) have suggested that the gneisses of the northern Kaapvaal Craton are significantly younger than those of the southern Kaapvaal Craton, with the Thabazimbi-Murchison-Lineament (TML) forming the boundary between the two (Figure 4.3). Overall these granitoid and gneiss terranes exhibit complex structural histories, and NE to ENE structural directions are prominent in various parts.

4.1.2 Sedimentary basins on the Kaapvaal Craton

By 3.1 Ga the main period of cratonization ended and the Kaapvaal Craton was a stable cratonic nucleus. A period of intra-cratonic extension followed between 3.1 Ga to 2.65 Ga, during which numerous sedimentary basins developed, such as the Pongola, Dominion, Witwatersrand and Ventersdorp basins (Figures 4.4. 4.5 and 4.6). After this period the Transvaal and Waterberg basins developed on the northern part of the Kaapvaal Craton. The Transvaal basin formed prior to the intrusion of the Bushveld Complex and the Waterberg basin formed just after the Bushveld Complex. The depositional axes of the basins migrated from SE to NW across the craton, accompanied by increasing inundation of the granite-greenstone shield (Hunter, 1981). Detritus was mainly derived from the N and NE. A long period followed of which no sedimentary rock record is preserved on the Kaapvaal Craton until the 345 Ma old Karoo basin. Each basin occurring on the Kaapvaal Craton will now be briefly discussed from oldest to youngest.

4.1.2.1 Pongola basin.

The Pongola sediments and volcanics are mainly preserved along the southeastern margin of the Kaapvaal Craton. The Pongola basin is estimated to have formed around 3.1 – 2.9 Ga ago (Tankard et al., 1982), and is interpreted to have been an epicontinental basin which was open to the southeast. The structures of the Pongola Supergroup have been complicated by early northwest-directed thrusting, followed by NS dextral and NW-SE sinistral shearing (Gold, 1983). It has been suggested by Matthews (1990) that the Pongola basin might have been deformed in a region which was part of a major transform boundary located at the southern margin of the Kaapvaal Craton. The main structural trend is depicted by the NW-SE orientated, and southeast plunging Amsterdam Syncline in the northern part of the Pongola Supergroup (Figure 4.4). The 2.87 Ga layered Usushwana Complex intruded along the flanks of the Amsterdam syncline and is aligned in the same direction. In addition a large open syncline which plunges to the SE is situated to the south of the Amsterdam syncline.

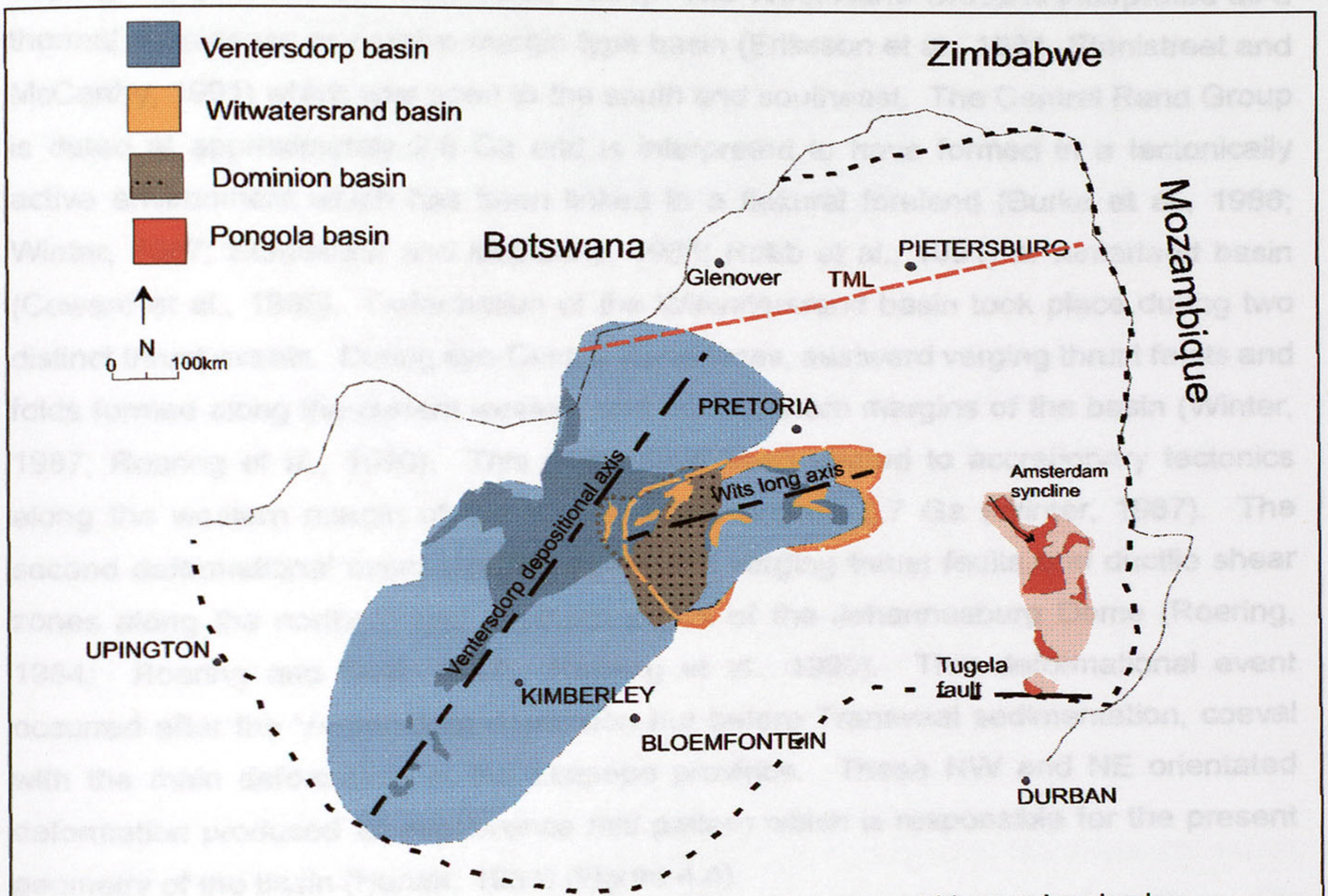

Figure 4.4 The distribution of the Pongola, Dominion, Witwatersrand and Ventersdorp basins. Lighter colours reflect inferred original outcrops, (after Tankard et al., 1982)

4.1.2.2 Dominion basin.

The Dominion Group situated in the center of the Kaapvaal Craton consists mainly of volcanics with minor sediments and has been dated between 3.09 - 3.07 Ga (Armstrong et al., 1990). The sediments are believed to have accumulated in an Andean-type back-arc basin (Burke et al., 1985) or in a failed rift basin (Bickle and Eriksson, 1982; Tankard et al., 1982). The basin is elongated with a NE-SW direction (Figure 4.4). The prominent structural features include northward verging ductile thrusting and folding (Van der Merwe, 1994).

4.1.2.3 Witwatersrand basin

The Witwatersrand basin is elongated in an ENE-WSW direction, and is divided into the older West Rand Group and a younger Central Rand Group (S.A.C.S., 1980). The West Rand Group is dated at approximately 3.0 Ga and is correlated with the Pongola Supergroup (Beukes and Cairncross, 1991). The West Rand Group is interpreted as a thermal subsidence or passive margin type basin (Eriksson et al., 1981, Stanistreet and McCarthy, 1991) which was open to the south and southeast. The Central Rand Group is dated at approximately 2.8 Ga and is interpreted to have formed in a tectonically active environment which has been linked to a flexural foreland (Burke et al., 1986; Winter, 1987; Stanistreet and McCarthy, 1991; Robb et al., 1991) or hinterland basin (Coward et al., 1995). Deformation of the Witwatersrand basin took place during two distinct thrust events. During syn-Central Rand times, eastward verging thrust faults and folds formed along the current western and northwestern margins of the basin (Winter, 1987; Roering et al., 1990). This might have been related to accretionary tectonics along the western margin of the central shield around 2.7 Ga (Winter, 1987). The second deformational event includes northward verging thrust faults and ductile shear zones along the northern and southern edges of the Johannesburg Dome (Roering, 1984; Roering and Smit, 1987; Roering et al., 1990). This deformational event occurred after the Ventersdorp deposition but before Transvaal sedimentation, coeval with the main deformation in the Limpopo province. These NW and NE orientated deformation produced an interference fold pattern which is responsible for the present geometry of the basin (Hunter, 1981) (Figure 4.4).

4.1.2.4 Ventersdorp basin

The Ventersdorp basin formed during a period of large scale crustal extension in the central Kaapvaal Craton. The basin is believed to have developed in a rift setting where extensional faults exploited the earlier formed NE-SW orientated thrust faults of the Witwatersrand basin. Subsequently, the Ventersdorp sediments were deposited in large NE-SW orientated grabens (Figure 4.4). The Ventersdorp basin is dated at approximately 2714 Ma for the base (Klipriviersberg Group) and 2709 Ma for the Platberg Group (Armstrong et al., 1991). The extrusion of the Ventersdorp lavas might have been a response to orogenic activity along the northern margin of the Kaapvaal Craton in the Limpopo Belt (Burke et al., 1985). Button (1981) described the Ventersdorp basin as being gently deformed.

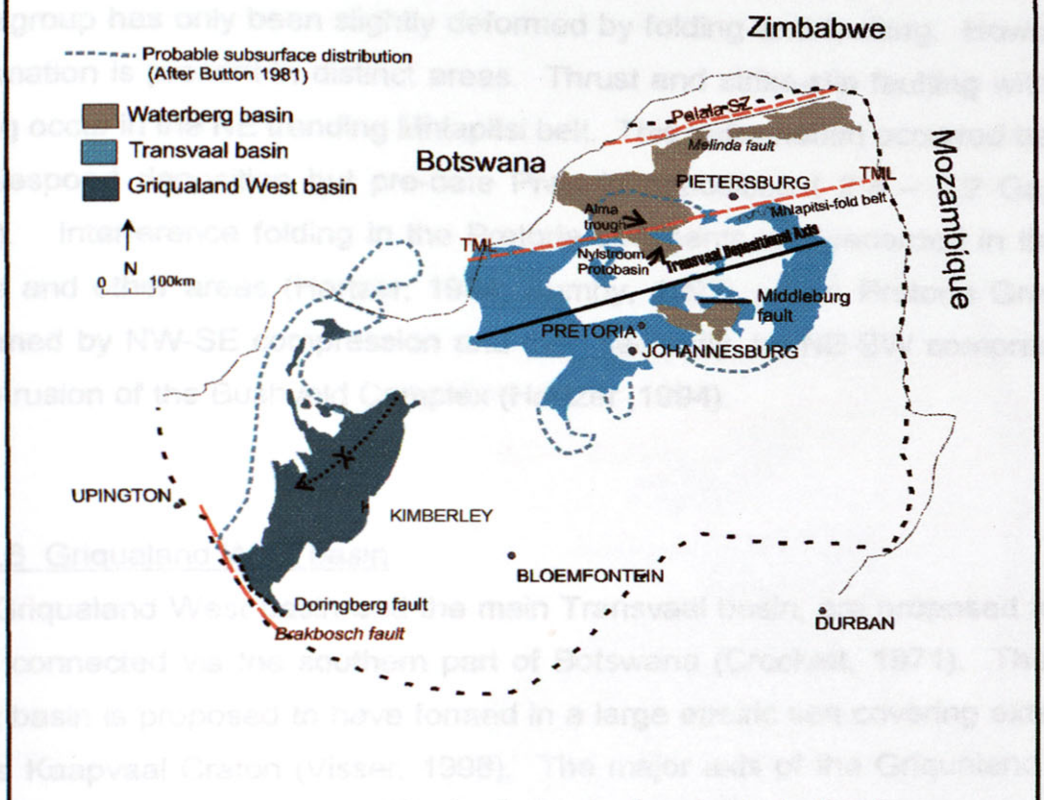

Figure 4.5. Transvaal and Waterberg basins, showing major structural trends.

4.1.2.5 Transvaal basin

The Transvaal basin is an important sedimentary basin influencing the structural setting of the Bushveld Complex since it acts as the floor and roof to the Complex. The Transvaal basin can be subdivided into three distinct depositional periods. The proto basin represented by the Wolkberg Group is dated at approximately 2600Ma (Eriksson et al., 1996) . The basin is believed to be rift related, with the Thabazimbi-Murchison fault zone at its northern margin strongly influencing deposition (Eriksson et al., 1996). Protobasinal rocks, characterized by rift-related volcanic rocks and immature sediments, are preserved along the northern and eastern margin of the Transvaal basin. The chemical sediments of the Chuniespoort Group followed which represents a thermal subsidence basin dated at approximately 2550 Ma (Clendenin, 1989). The sedimentation of the Pretoria Group commenced in an extensional tectonic setting, either within half-grabens, controlled by the TML (Eriksson et al., 1991), or the beginning of a continental rift (Schreiber et al., 1992). The main depositional axis of the Transvaal basin trends roughly ENE, and generally strata dip towards the center of the basin and towards the Bushveld Complex (Figure 4.5). Visser (1998) stated that the Transvaal Supergroup has only been slightly deformed by folding and faulting. However, intense deformation is present in distinct areas. Thrust and strike-slip faulting with associated folding occur in the NE trending Mhlapitsi belt. This deformation occurred between post-Chuniespoort deposition but pre-date Pretoria deposition (2.4 – 2.2 Ga) (Potgieter, 1992). Interference folding in the Pretoria sediments was recorded in the Transvaal inliers and other areas (Hartzer, 1994; Bumby, 1997). The Pretoria Group was first deformed by NW-SE compression and then secondly, by NE-SW compression before the intrusion of the Bushveld Complex (Hartzer ,1994).

4.1.2.6 Griqualand West basin

The Griqualand West basin and the main Transvaal basin, are proposed to have once been connected via the southern part of Botswana (Crockett, 1971). The Griqualand West basin is proposed to have formed in a large epeiric sea covering extensive areas of the Kaapvaal Craton (Visser, 1998). The major axis of the Griqualand West Basin trends roughly NS, (Figure 4.5). Overall the Supergroup has only been moderately deformed. However, deformation along the southwestern margin of the basin is believed to be in response to deformation during the Kheis orogeny at 2.0 Ga (Visser, 1998). Other structures include a large gentle syncline which trends SW, situated in the center of the basin. Fold orientations include NW-SE, NE-SW and NS (Visser, 1998), and faulting follows NNE, NW and some NS orientations.

4.1.2.7 Waterberg basin

The Waterberg basin formed in a half graben setting with the Thabazimbi-Murchison fault zone forming the southern boundary (Callaghan et al., 1991) The 1.8 Ga Waterberg basin came into being during a period where tensional conditions existed on the Kaapvaal Craton due to cooling of the crust after the intrusion of the Bushveld Complex (Jansen, 1982). The Nylstroom protobasin developed to the south of the Thabazimbi-Murchison fault while a deep trough, known as the Alma trough, developed on the northern side of the Thabazimbi-Murchison fault zone (Figure 4.5). ENE to EW trending structures, such as thrust faults and folds, dominate the southern margin of the main Waterberg basin. A southward extension of the Steelpoort fault forms the northwestern margin of the Middelburg-Cullinan basin. Characteristic structures of the Cullinan Middelburg basin also include ENE to EW orientated faults and folds (van der Neut and van der Merwe, 2000).

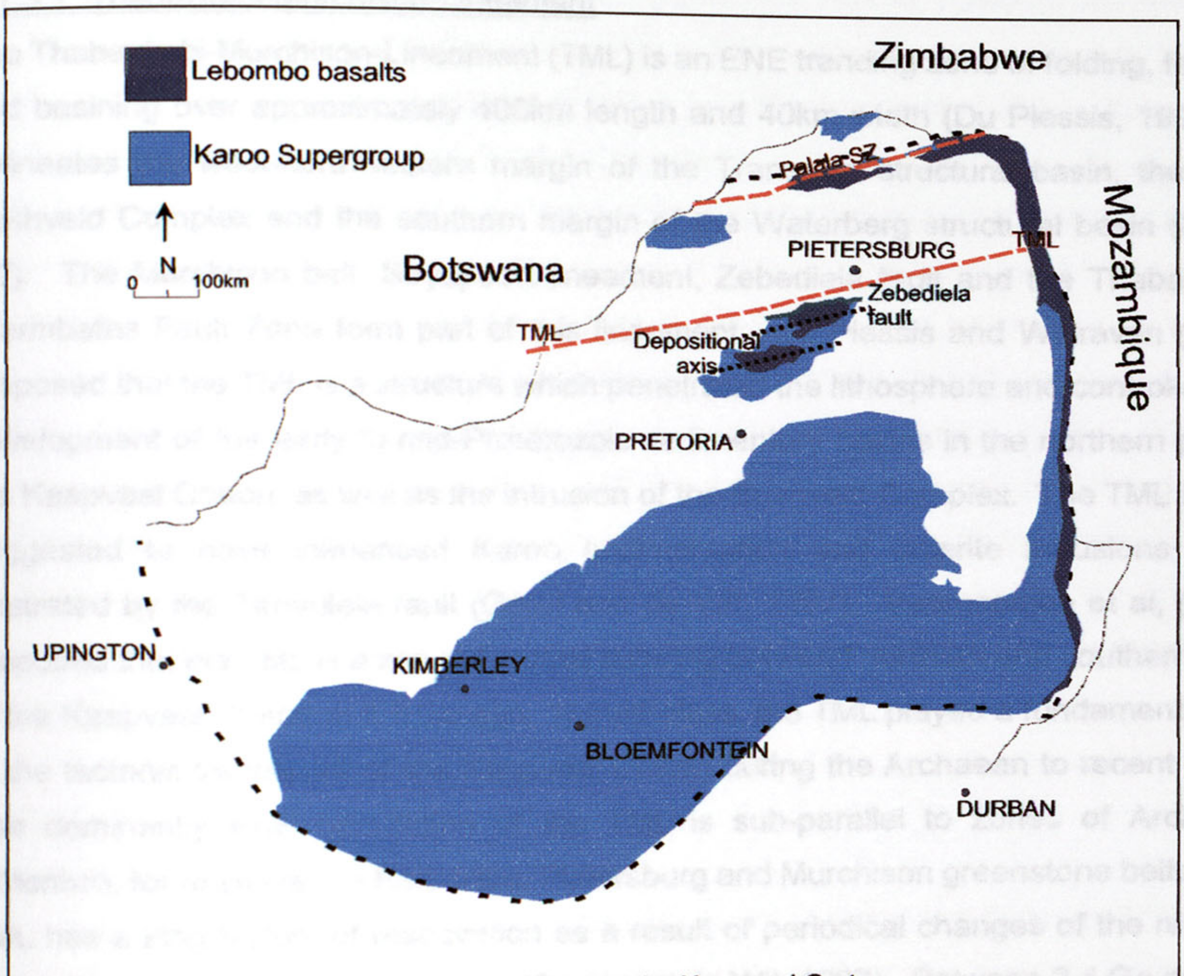

Figure 4.6. Outcrops of Karoo rocks on the Kaapvaal Craton.

4.1.2.8 Karoo basin

The main Karoo basin is interpreted to be a foreland basin of roughly 345 Ma old covering a large portion of southern Africa (Cole, 1992). Located in the center of the Bushveld Complex is what is known to be a preserved remnant of the main Karoo basin. Here the Karoo rocks occur in a basin with an ENE trending axis, roughly parallel to that of the Bushveld Complex and Transvaal basin (Figure 4.6). The basin is bounded on the northwest side by the roughly NE striking Zebediela fault. Karoo rocks also occur in the northern part of the Kaapvaal Craton against the Palala shear zone and the Central Zone of the Limpopo mobile belt (Figure 4.6).

4.1.3 Major Structural lineaments on the Kaapvaal Craton

4.1.3.1 Thabazimbi-Murchison-Lineament

The Thabazimbi-Murchison-Lineament (TML) is an ENE trending zone of folding, faulting and basining over approximately 400km length and 40km width (Du Plessis, 1991). It delineates the west-northwestern margin of the Transvaal structural basin, the main Bushveld Complex and the southern margin of the Waterberg structural basin (Figure 4.7). The Murchison belt, Strydpoort lineament, Zebediela fault and the Thabazimbi-Warmbaths Fault Zone form part of this lineament. Du Plessis and Walraven (1990) proposed that the TML is a structure which penetrated the lithosphere and controlled the development of the early to mid-Proterozoic sedimentary basins in the northern part of the Kaapvaal Craton, as well as the intrusion of the Bushveld Complex. The TML is also suggested to have influenced Karoo sedimentation and dolerite intrusions as is illustrated by the Zebediela fault (Good and de Wit, 1997). Vearncombe et al, (1992) proposed that the TML is a zone of crustal suturing between northern and southern parts of the Kaapvaal Craton at 2.8 Ga ago. Nonetheless, the TML played a fundamental role in the tectonic framework of the Kaapvaal Craton during the Archaean to recent times. The dominantly ENE orientation of the TML is sub-parallel to zones of Archaean tectonism, for example the Barberton, Pietersburg and Murchison greenstone belts. The TML has a long history of reactivation as a result of periodical changes of the regional stress fields on the Kaapvaal Craton (Good and de Wit, 1997). Between 2.4 Ga and 2.2 Ga ago the TML shows evidence of positive inversion (Potgieter, 1992). At 2.02 Ga the TML was reactivated during the intrusion of the Bushveld Complex as left-lateral strike-slip faults (Du Plessis and Walraven, 1990). The final reactivation was normal movement in post-Karoo times, along structures such as the Zebediela Fault (Good and de Wit, 1997).

4.1.3.2 Limpopo Mobile Belt

The Limpopo Belt is a major ENE trending orogenic belt which represents a suture zone between the Zimbabwe Craton to the north and the Kaapvaal Craton in the south.

These two cratons are believed to have collided at around 2.65Ga ago involving oblique collision, with the Kaapvaal Craton moving northwestwards (Roering et al., 1992). However, many authors are of the opinion that the two cratons collided by NS directed collision (Coward and Fairhead, 1980; Light, 1982, Van Reenen et al., 1988). Also, it has been recently proposed (Barton et al., 1994; Holzer et al., 1998) that the collision may be as young as early Proterozoic, the time of the 2.0 Ga tectonothermal event (Van Breemen and Dodson, 1972). This tectonothermal event could, however, be related to the intrusion of the Bushveld Complex. Three zones can be recognized in the belt based on their distinct geological characteristics (van Reenen et al., 1992). They include the Southern Marginal Zone, the Central Zone and the Northern Marginal Zone (Figure 4.8). Detailed work on the lithological character of each zone has been done by authors such as van Reenen et al. (1992). The boundary between the Southern Marginal Zone and the Kaapvaal Craton is a northward-dipping shear zone. The boundary between the Southern Marginal Zone and the central zone is marked by the Palala shear zone, a left-lateral strike-slip fault zone, which was active after the intrusion of the Bushveld Complex (Brandl and Reimold, 1990). While the Northern Marginal Zone and the Central Zone are separated by the Tuli-Sabi shear zone, a gently-dipping dextral strike-slip fault (McCourt and Vearncombe, 1987). The boundary between the Northern Marginal Zone and the Zimbabwe Craton is marked by a southward-dipping thrust fault, (Figure 4.8).

4.1.3.3 Other proposed lineaments

Trompsburg-Great Dyke lineament

It has been noted that layered mafic intrusives, such as the Great Dyke of Zimbabwe, the Losberg Complex, the northern lobe of the Bushveld Complex and the Trompsburg intrusive Complex all lie along a NS trending lineament (Hall, 1932). All these layered intrusives have similar geochemistries and they become progressively younger towards the south. The Great Dyke of Zimbabwe is dated at around 2.5 Ga, and the Losberg Complex at 2.04 Ga. The Bushveld Complex has an age around 2.05 Ga and the Trompsburg Complex is believed to be 1.3 Ga (Visser, 1998). Hall (1932), Willemse (1969), and Hunter (1975) suggested that these intrusions might depict a north-northeast trending abyssal fracture (Tankard et al., 1982). Intermittent injection of magma occurred over a period of 1 400 Ma along this fracture. The Bushveld Complex is believed to have developed over a wide zone across this fracture (Visser, 1998).

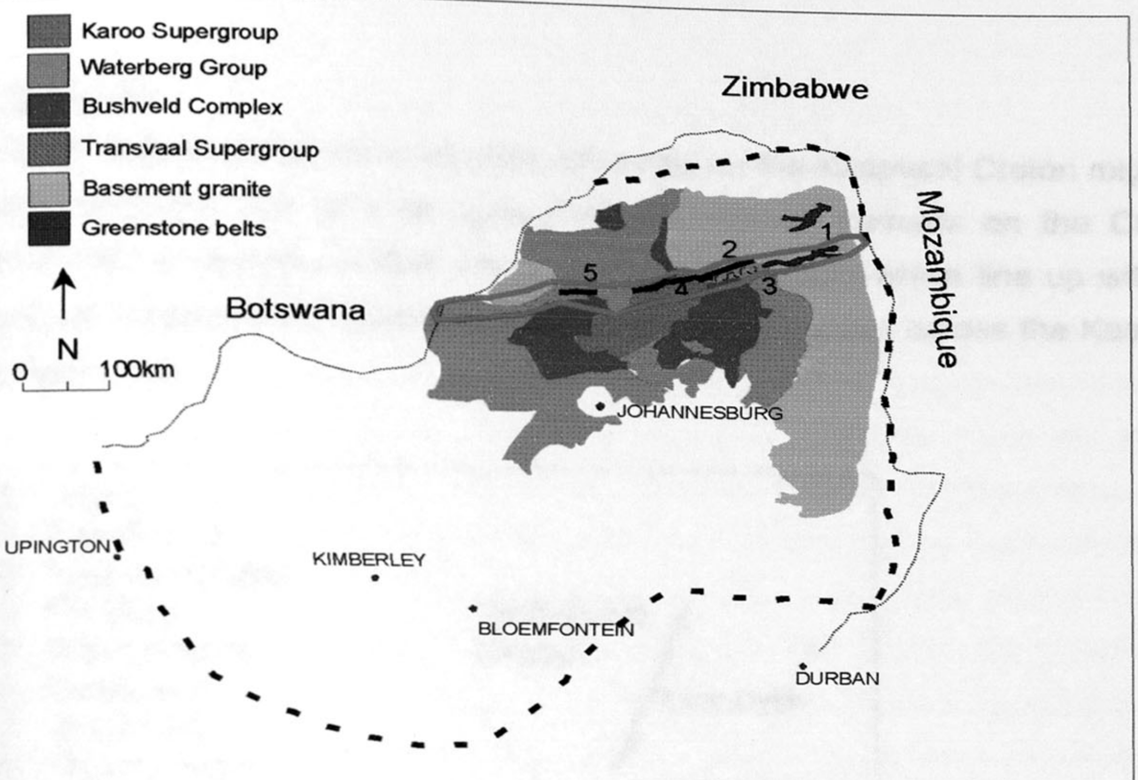

Figure 4.7 Location and geology of the Thabazimbi Murchison Lineament on the Kaapvaal Craton
Structures defining the Thabazimbi Murchison Lineament.
1. Murchison greenstonebelt, 2. Strydpoort lineament,
3. Mhlapitsi fold belt, 4. Zebediela fault, 5. Thabazimbi fault zone
(modified from Good and de Wit, 1997)

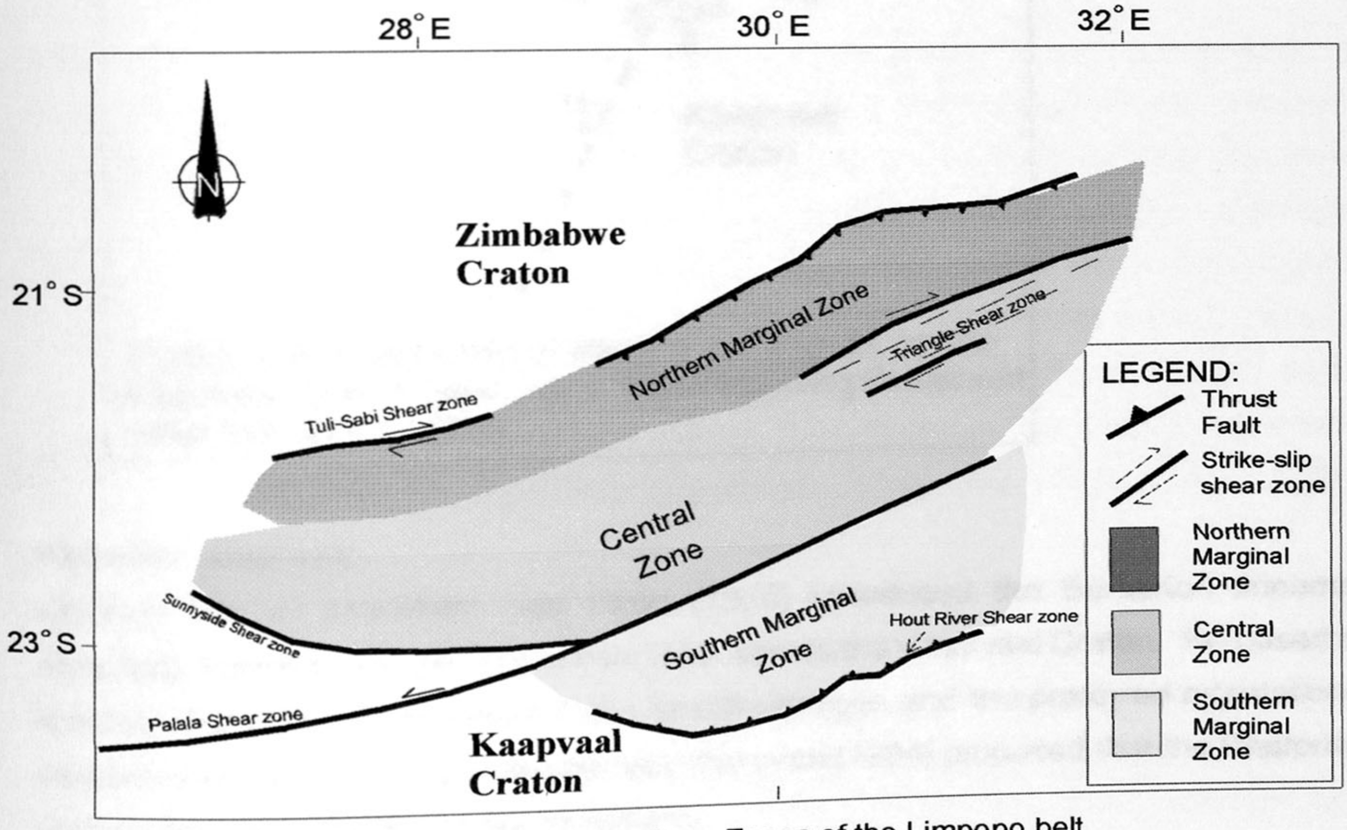

Figure 4.8. Southern, Central and Northern Zones of the Limpopo belt
(modified from Bumby, 2000)

Alkaline Intrusives

Visser (1998) proposed that some alkaline intrusives on the Kaapvaal Craton might be structurally controlled due to their ages and distributional patterns on the Craton. Verwoerd (1993) envisioned a NNE trending structural feature which line up with the Great Dyke of Zimbabwe for alkaline complexes and carbonatites across the Kaapvaal Craton, (Figure 4.9).

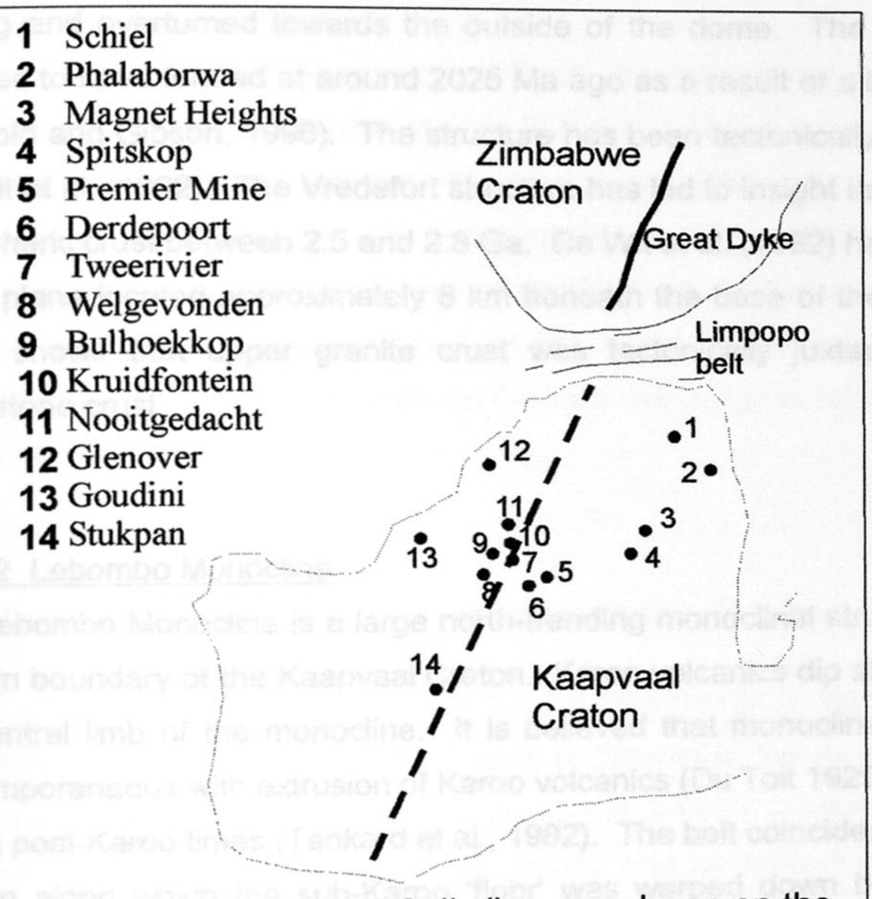

Figure 4.9 Distribution of alkaline complexes on the Kaapvaal Craton depicting a NNW trending lineament (after Verwoerd, 1993).

Barberton lineament

Although not so prominent, van Biljon (1976) introduced the Barberton lineament stretching from the Barberton Mountain land across the Kaapvaal Craton. He based the location of this lineament form ERTS-I satellite images and the preferred orientation of structures along this line. Du Plessis and Walravan (1994) proposed that the Rietfontein fault system is connected to the lineament.

4.1.4 Other important structures affecting the Kaapvaal Craton

4.1.4.1 Vredefort dome

The Vredefort dome, located in the center of the Craton is made up of a granitic core, approximately 50 km wide, with a roughly circular shaped rim 15-20 km wide (Reimold and Gibson, 1996). The rim or collar is made up of rocks belonging to the Dominion, Witwatersrand, Ventersdorp and the Transvaal sequences. These rocks are steeply dipping and overturned towards the outside of the dome. The Vredefort structure is believed to have formed at around 2025 Ma ago as a result of a large meteoritic impact (Reimold and Gibson, 1996). The structure has been tectonically tilted after the impact (de Wit et al., 1992). The Vredefort structure has led to insight into the formation of the lithospheric crust between 2.5 and 2.8 Ga. De Wit et al. (1992) have suggested a major thrust plane located approximately 8 km beneath the base of the Witwatersrand basin which shows that upper granite crust was tectonically juxtaposed above granite-greenstone crust.

4.1.4.2 Lebombo Monocline

The Lebombo Monocline is a large north-trending monoclinal structure which forms the eastern boundary of the Kaapvaal Craton. Karoo volcanics dip steeply east-ward along the central limb of the monocline. It is believed that monoclinal warping was in part contemporaneous with extrusion of Karoo volcanics (Du Toit 1929) but was accentuated during post-Karoo times (Tankard et al., 1982). The belt coincides with a zone of crustal tension along which the sub-Karoo 'floor' was warped down by at least 9 km (Cox, 1970). The Lebombo flexure probably marks the eastward transition from normal continental crust to attenuated crust. It may also be related to the junction between the Archaean Kaapvaal Craton and the Late Proterozoic Mozambique Belt (Visser, 1998).

4.1.4.3 Vryburg Arch

A broad belt of Archaean domal structures define what is known to be the Vryburg Arch (Hunter and Hamilton, 1978). The belt flanks the western margin of the Bushveld Complex, and Transvaal rocks attenuate across the Arch into Botswana, thus separating the Griqualand West and Transvaal basins (Tankard et al., 1982). All of the domes are composed of granitic basement and include the Vredefort, Johannesburg, Makoppa, Gaberone, Molopo, Mafikeng and Klerksdorp domes. Thick Mesozoic and Proterozoic sedimentary sequences occur on either side of the Arch (Tankard et al., 1982). It is believed that the Arch has been active over a long period of time stretching from Archaean times to Pilanesberg times at 1300Ma (Visser, 1998).

4.1.4.4 Mafic dyke swarms

Uken and Watkeys (1997a) recognized three orientations of dyke swarms during Pre-Karoo times. These orientations include NW, NE and EW dyke events. They propose that the NW dykes are associated with Pongola and Witwatersrand rifting, whereas the NE dykes are associated with Ventersdorp rifting. They further proposed that Craton wide compression during the intrusion of the Bushveld Complex was responsible for the emplacement of EW orientated dykes. Karoo-age dyking is characterized by NNW, NE and NS trending dykes associated with the Mesozoic fragmentation of Gondwana (Uken and Watkeys 1997b).

4.1.5 Other marginal tectonic events affecting the Kaapvaal Craton

The Eburnian orogeny prevailed along the western margin of the Craton during the early Proterozoic. The convergent margin type sediments of the Richtersveld Province and the Kheis orogeny at 1.75 Ga both represent this event (Thomas et al., 1993). From mid-late Proterozoic major continental growth and accretionary processes dominated along the southern and western margins of the Craton. This is known as the approximately 1.2 Ga Kibaran event which was responsible for the formation of the Namaqua-Natal Metamorphic Province. Subsequent to this orogenesis was the Late Paleozoic Pan-African event (Thomas et al., 1993). This event is marked by a period of extensive continental fragmentation with geosynclinal deposition seen in the Gariep and Saldanian provinces (Thomas et al., 1993). By the early Palaeozoic the Kaapvaal Craton was situated in the center of the super-continent Gondwana. During this time the Cape Supergroup was deposited along the southern margin of the south African continent in an aborted rift type setting (Tankard et al., 1982). The Cape Supergroup was followed by the formation of the large intracratonic Karoo foreland basin. The Cape and Karoo successions were deformed along the Cape fold belt, due to what is believed to be shallow subduction of oceanic crust at the supercontinental margin. The break-up of Gondwana then commenced in two stages around – 180 and 160 Ma and 135 Ma along lines of Proterozoic and Archaean crustal weakness (Dingle et al., 1983: Martin and Hartnady, 1986; Watkeys and Sweeney, 1988).

4.2 TECTONIC SETTING OF THE BUSHVELD COMPLEX

The Bushveld Complex is situated in the center of the main Transvaal basin (Figure 4.10). The main axis of the Complex trends roughly ENE, parallel to the long axis of the Transvaal basin. Transvaal rocks form the floor and roof of the Complex except towards the northern lobe where Bushveld rocks overlie granitoid-gneiss basement. The mafic phase of the Complex has been dated at 2061 $\pm$ 27 Ga (Walraven, 1990) and the acid phase at 2.05 Ga (Harmer and Von Gruenewaldt, 1991). Several models have been proposed for the distribution of the Complex. Du Plessis and Walraven (1990) summarized the models as follows:

- a complex basin-shaped continuous body (Willemse, 1969)
- a complex comprising four separate bodies (Cousins, 1959)
- a cruciform body comprising four separate lobes, each with its own center of intrusion (Hunter, 1975)
- a major body with interconnected compartments fed from seven centers of intrusion. (Von Gruenewaldt, 1979)

A regional gravity survey done by Smit et al. in 1962 has shown that the mafic rocks are not continuous beneath the younger cover rocks and therefore suggested that the Complex does not have a simple lopolithic form. Smit et al. (1962) proposed a four leaf clover shape for the Complex which represents four separate intrusions. The four lobes include (Figure 1.5);

1. The western lobe, extending from near Pretoria westward to Rustenburg and around the Pilanesberg Alkaline Complex to and along the southern flank of the Makopa Dome.
2. The southeastern lobe which is mostly covered by Mesozoic rocks.

3. The eastern lobe which continues northward from the southeastern lobe to about 50 km east of Potgietersrus.
4. The northern lobe which extends from south of Potgietersrus to Villa Nora.

Overall the Bushveld Complex is relatively undeformed. Dips of the layered sequence of basic and ultrabasic rocks in each of the lobes are toward the center, usually at low angles between 10° and 25°, except for in the northern lobe where it averages 60°. The current structural pattern of the Complex is characterized by three large and prominent faults. These faults include: in the west, the NW striking Rustenburg fault and Brits Graben, in the center, the NNE striking Wonderkop fault, and in the east the NE striking Steelpoort fault (Figure 1.4). Furthermore, the Bushveld Complex is intensely deformed along the Palala shear zone.

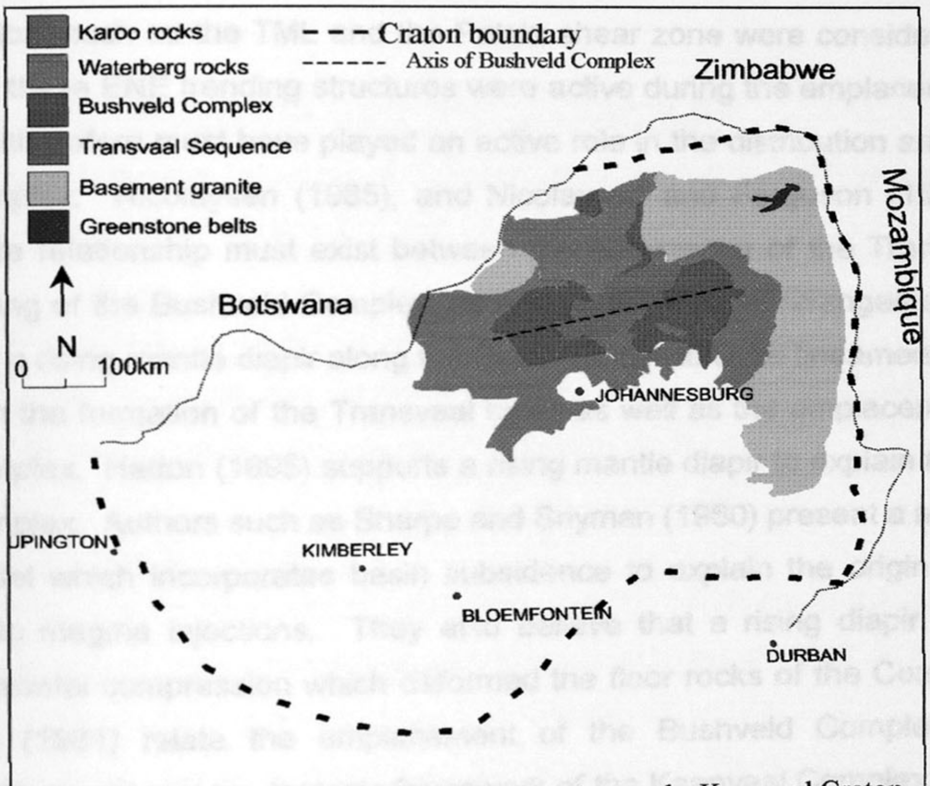

Figure 4.10. Location of the Bushveld Complex on the Kaapvaal Craton

Various models for the tectonic setting and emplacement of the Bushveld Complex have been proposed in the past. These models vary from stable cratonic settings to plate tectonic linked orogenies, as well as major crustal features controlling the emplacement of the Complex (Du Plessis and Walraven 1990). The first author to speculate about the tectonic setting of the Bushveld Complex was probably Hall (1932). He suggested that a deep crustal fracture, the Bushveld-Great Dyke line of intrusion, was responsible for the emplacement of the Complex. Some authors have also speculated about the siting of the Complex at the intersection of the TML and the Bushveld-Great Dyke mega-fracture, (Visser, 1998). On the other hand, Lee and Sharpe (1986) have concluded by means of LANDSAT imaging that no deep seated crustal fractures were responsible for the emplacement of the Complex. Van Biljon (1976) relates the intrusion of the Complex to an active spreading center with major lineaments such as the TML and Barberton lineaments acting as major transform faults. Furthermore, Du Plessis and Walraven (1990) have suggested that the emplacement of the Complex was indeed structurally controlled. During their investigations major structural lineaments of the Kaapvaal Craton, such as the TML and the Palala shear zone were considered. They proposed that these ENE trending structures were active during the emplacement of the Complex and therefore must have played an active role in the distribution and deformation of the Complex. Nicolaysen (1985), and Nicolaysen and Ferguson (1980), suggested that some relationship must exist between the occurrence of the Transvaal basin and the setting of the Bushveld Complex, as well as its elliptical arrangement. They concluded that a rising mantle diapir along the Bushveld-Great Dyke lineament was responsible for both the formation of the Transvaal basin as well as the emplacement of the Bushveld Complex. Hatton (1995) supports a rising mantle diapir to explain the distribution of the Complex. Authors such as Sharpe and Snyman (1980) present a mathematically based model which incorporates basin subsidence to explain the origin of sill intrusion and mafic magma injections. They also believe that a rising diapir was responsible for horizontal compression which deformed the floor rocks of the Complex. Vermaak and Lee (1981) relate the emplacement of the Bushveld Complex to the geological development and the tectonic framework of the Kaapvaal Complex. They conclude that the emplacement may be viewed as a response to superposition of NNW and ENE crustal warps (in Visser, 1998). Tankard et al. (1982) suggests that the emplacement of the Bushveld Complex is probably due to a combination of the above mentioned mechanisms.