3 GENERAL GEOLOGY OF THE LIMPOPO BELT

3.1 Shear Zones bounding the Central Zone

The **Triangle Shear Zone** that bounds the CZ in the north represents a 35 km wide belt of transpressive *dextral* deformation that dips to the south at a shallow angle (Kamber et al., 1995a). Protomylonites and mylonites dominate and are clearly derived from NMZ lithologies. However, Rollinson & Blenkinsop (1995) have observed quartzitic rocks and Ridley (1992) reports calc-silicates and pelitic gneisses, which suggest that some CZ lithologies may also be entrained in the Triangle Shear Zone. Van Breemen & Hawkesworth (1980) and Kamber et al. (1995a) have dated *dextral* strike slip movement between 2040 and 1990 Ma, using Pb-Pb garnet and Sm-Nd garnet whole-rock ages respectively. This movement occurred under granulite grade conditions (750^o to 800^{o}C at 0.8 GPa) (Kamber et al., 1995a).

The 10 km wide, **Palala Shear Zone** (also called the "Palala-Zoetfontein lineament", Schaller et al., 1999), dips steeply to the north and field observations and microstructural work indicate both, *sinistral* and *dextral* shearing, but on the whole, *dextral* movements are predominant. In the Koedoesrand Window (Figure 2, and for the locality, see Figure 1), the Palala Shear Zone marks the boundary between the granulite facies Limpopo CZ and the low-grade rocks of the adjacent KC. Schaller et al. (1999) studied the tectonic evolution of the Palala Shear Zone in the Koedoesrand Window and presented evidence that strike slip shearing started at high temperatures, accompanied by the exhumation of granulite rocks, and continued during a fall in metamorphic grade to greenshcist facies conditions. This event is bracket between a Pb-Pb titanite age of 2020 Ma at peak granulite conditions and an Rb-Sr biotite age of 1971 ± 26 Ma from a greenshcist facies mylonite (Schaller et al., 1999).

3.2 Shear Zones bounding the marginal zones

The two marginal zones of the LB (NMZ and SMZ) are bounded by major reverse shear zones with down dip mineral stretching lineations, the Hout River Shear Zone (Smit et al., 1992) in the south and the North Marginal Thrust Zone (Blenkinsop et al., 1995) in the north (Figure 1). The **Hout River Shear Zone** consists of E-W trending, steeply northward-dipping thrusts and several near vertical NE trending strike-slip faults (Smit et al., 1992). De

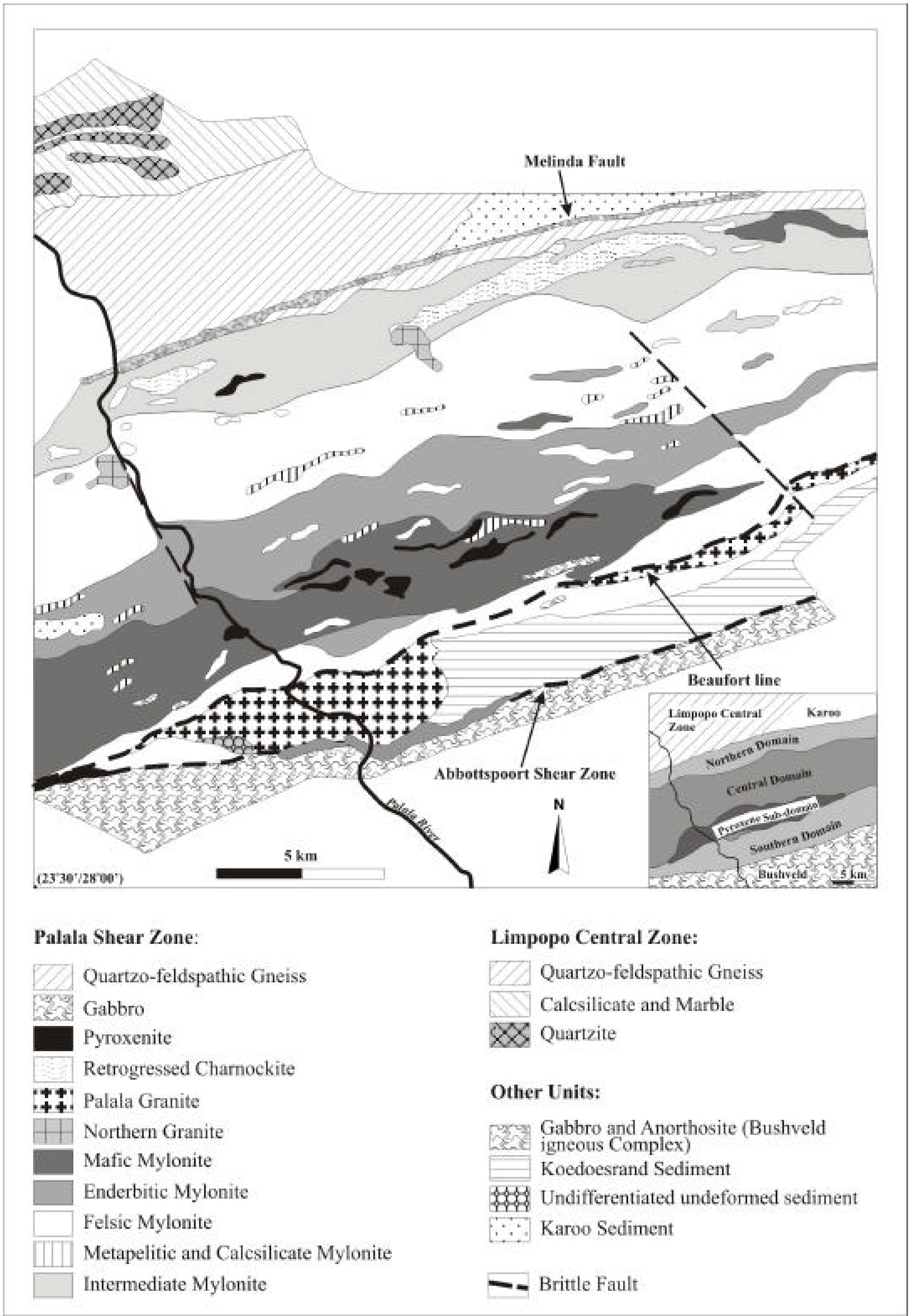

Figure 2: Geological map of the Palala Shear Zone in the Koedoesrand Window. The small inset shows the different tectonic units: Limpopo Central Zone, Palala Northern, Central and Southern Domains, and also the Bushveld Igneous Complex (Modified after Schaller et al., 1999).

Beer & Stettler (1992) confirmed that the high-grade rocks of the SMZ were thrusted southward over the low-grade cratonic rocks along this shear zone, based on their detailed geoelectrical and seismic reflection data. Geothermobarometric studies likewise showed that the vertical displacement across the Hout River Shear Zone was of the order of 15 km (Miyano et al., 1992; Perchuk et al., 1996). The geometry of the Hout River Shear Zone changes from strike-slip faulting in the eastern section into a system of frontal and lateral ramps in the western section (Smit et al., 1992) (Figure 3). The timing of thrusting in the SMZ has been constrained by the 2671 ± 2 Ma zircon date for the late-tectonic Matok pluton (Barton et al., 1992a) and by Pb-Pb step leaching dates on synkinematic metamorphic minerals in rocks from the adjacent KC forming the footwall of the Hout River Shear Zone (Kreissig et al., 2000). Garnet and kyanite in such schists gave ages of 2691 ± 20 Ma and 2672 ± 51 Ma respectively.

The shallow southerly dipping **Northern Marginal Thrust Zone** has a strike length of 320 km and consists of several mylonitic zones a few tens of meters wide separated by protomylonites (Rollinson & Blenkinsop, 1995). These workers also pointed out that reverse shear sense is demonstrated by several shear sense indicators and the latest stage of this reverse shearing occurred under greenshcist facies conditions as indicated by syn-tectonic growth of chlorite and epidote. Intrusive granites occur as elongated batholiths, aligned along the regional NNE-SSW trend of the shear zone, and are referred to as the Razi Suite (Robertson, 1973) (Figure 4). The Razi Suite granites are often porphyroclastic and can be found as mylonites, protomylonites and undeformed intrusions in the shear zone, and therefore interpreted to intrude the shear zone syn- (throughout the deformation) to post-tectonically (Rollinson & Blenkinsop, 1995). Zircon U-Pb ages range from ± 2627 to ± 2547 Ma for intrusions of this suite (Mkweli et al., 1995; Frei et al., 1999). Together with the age for a charnoenderbite from the NMZ (2637 ± 0.019 Ma, Berger et al., 1995) these data were interpreted to constrain the timing of the main thrusting to between approximately 2640 and 2600 Ma (Kramers et al., 2001). A reactivation of this shear zone, under greenshcist facies conditions, occurred at approximately 2000 Ma (Kamber et al., 1996)

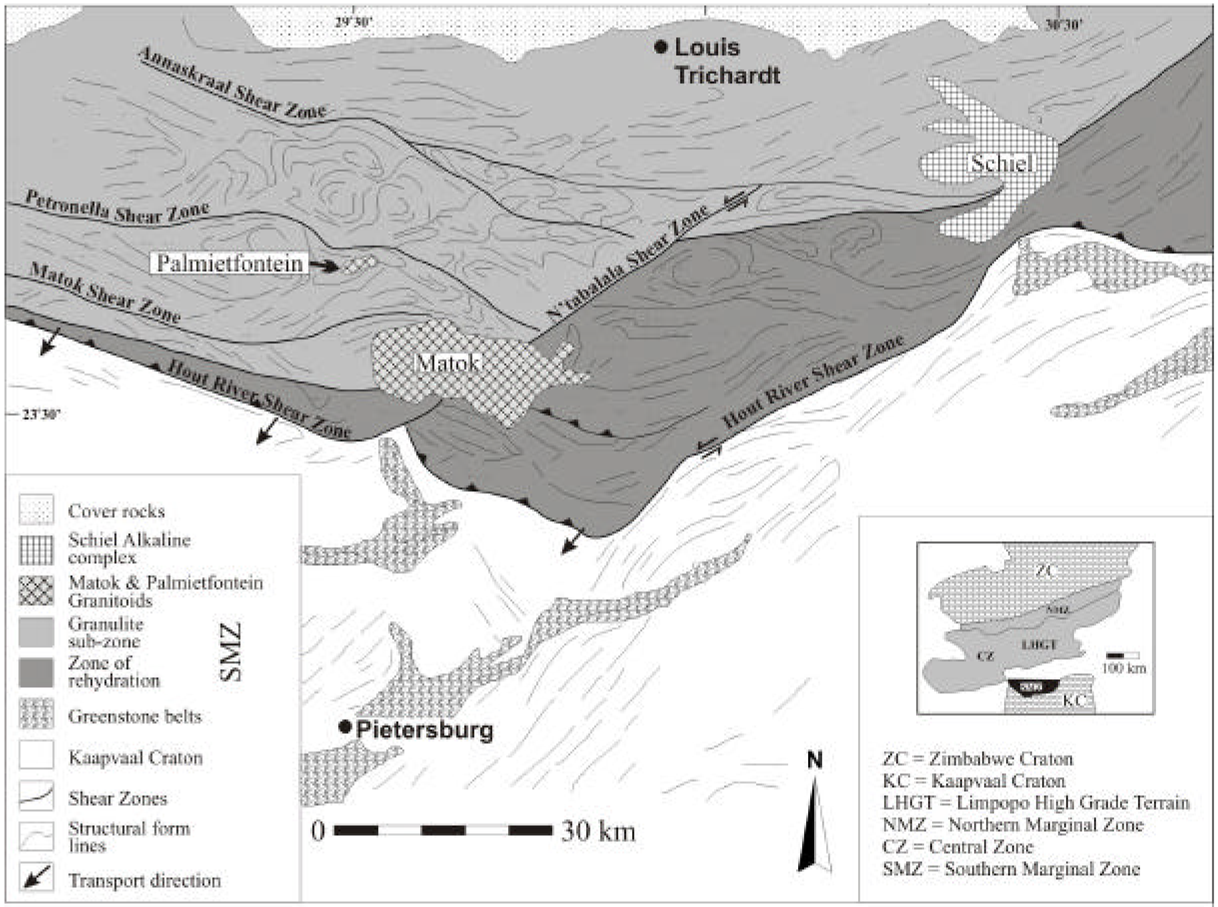

Figure 3: Structural map of the Southern Marginal Zone of the Limpopo Belt (Modified after Smit and Van Reenen, 1997), showing deep crutal shear zones, the Hout River Shear Zone, and the Matok Pluton.

3.3 The Northern Marginal-, Southern Marginal- and Central Zone

The two marginal zones are considered as lower crustal equivalents of the adjacent granite greenstone terranes (Du Toit et al., 1983; Kreissig et al., 2000). The **NMZ** is exposed in Botswana and Zimbabwe and is about 60 km wide and 450 km long (e.g. Rollinson & Blenkinsop, 1995; Kamber & Biino, 1995) (Figure 4). This marginal zone is predominantly

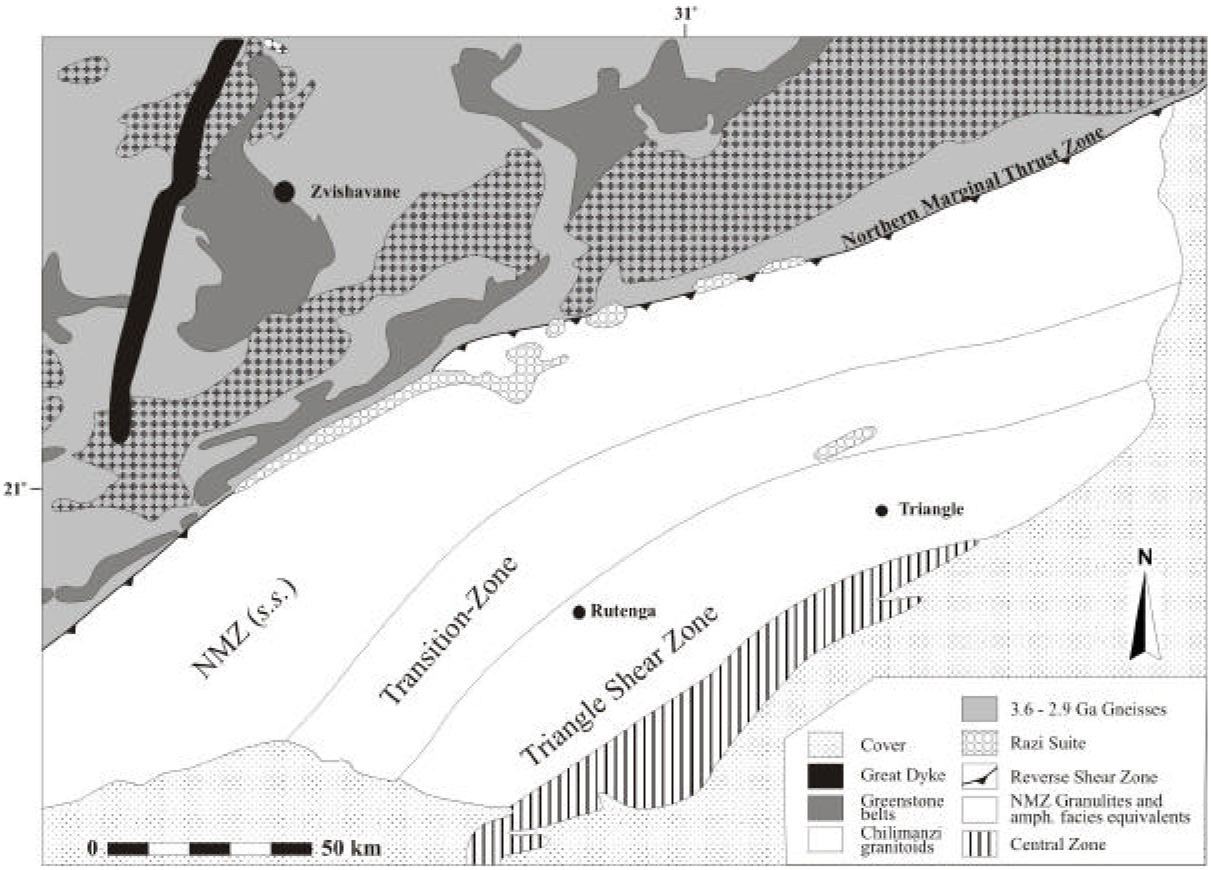

Figure 4: Geological sketch map of the Northern Marginal Zone of the Limpopo Belt (Modified after Berger et al., 1995)

composed of igneous charnockite and enderbite that intruded ultramafic, mafic, and subordinate felsic and metasedimentary units between 2710 ± 38 and 2637 ± 19 Ma (Berger et al., 1995). The charnockite and enderbite gneisses are regarded as orthogranulites *senso stricto* i.e. the products of a melt that crystallized under granulite facies conditions (Ridley, 1992; Berger et al., 1995; Rollinson & Blenkinsop, 1995). These orthogranulites account for over 90% of the present surface area of the NMZ (Ridley, 1992). Porphyritic granite of the Razi suite (Robertson, 1973) outcrops as a series of elongate bodies along the northern boundary of the granulite zone and forms a small but important component of the NMZ (Figure 4). The western part of the NMZ was later intruded by the N-S trending Great Dyke around 2575 ± 5 Ma, and its satellites around 2579 Ma (Armstrong & Wilson, 2000). Peak metamorphic conditions for the NMZ were recorded to have temperatures up to 800^{o}C and pressures up to 8.5 Kbar, using frozen mineral reaction textures and thermobarometry (Kamber & Biino, 1995; Rollinson 1989). These authors proposed an anti-clockwise *P-T* path for the granulites in which the peak temperature was reached synchronously with dominant SSE-NNW compression and syn-tectonic porphyritic charnockite and granite emplacement (Razi Granites), which occurred between 2650 and 2500 Ma (Rollinson & Blenkinsop, 1995; Blenkinsop & Frei, 1997). It should be noted that the proposed anticlockwise *P-T* evolution, based on a study of mafic granulites, is not in agreement with *P-T* paths derived from metapelitic gneisses (Tsunogae et al., 1992) that show a common clockwise *P-T* evolution similar to that of the SMZ. Structurally the NMZ is characterized by homogeneous NE trending, steeply southward dipping foliations with down-dip lineations and sub-horizontal fold axes.

The approximately 60 km wide ENE-WSW trending **SMZ** consists of granulite facies supracrustal gneisses of mafic, ultramafic and metapelitic composition, known as the Bandelierkop Formation, intermixed with volumetrically dominant tonalitic grey gneiss referred to as the Baviaanskloof Gneiss (Du Toit et al., 1983). Most lithologies are migmatitic and a moderate to strong steeply north dipping, ENE-WSW trending foliation, in which highly attenuated greenstone remnants occur in gneissic and migmatitic granitoids. The most prominent structural feature of the SMZ is the presence of large crustal sheets (Smit et al., 1992), composed of migmatitic granulitic gneiss, which show a complex fold pattern and evidence for early, regional-scale shear deformation. Intrusive granitoids include the composite late- to post-kinematic Matok pluton, for which zircon-ages between 2664 and 2671 Ma were determined (Retief et al., 1990; Barton et al., 1992) followed by the post-kinematic Palmietfontein Granite with a whole rock Rb-Sr isochron age of 2456 ± 78 Ma (Barton et al., 1983), and the 1850 Ma old Schiel Alkaline Complex (Barton et al., 1996) (Figure 3). The granulites in the SMZ are characterized by both decompression-cooling (DC), and isobaric-cooling (IC) *P-T* paths (Perchuk et al., 2000; Smit et al., 2001). IC paths characterize the contact of the SMZ with the KC, while IC paths are common *P-T* trajectories for rocks located far (> 40 km) away from this contact. Mineral equilibria from the juxtaposed KC in the footwall sections of the HRSZ record *P-T* loops: peak *P-T* conditions were reached at the lowest *P-T* conditions recorded by the adjacent granulites. The identical isotopic ages suggests synchronous movement of the KC during exhumation of the adjacent high-grade terrain, while granulites move up to the surface, cool metabasalts and metakomatiites move down, cooling the granulites along the contact zone. The existence of both DC and IC paths in the same granulite complex can thus be explained by differences in the movement of different granulite "blocks" during exhumation. In contact with the KC rocks, these ascending blocks have been cooled isobarically for about 100^{o}C because of the *T* gradient between the hot high-grade terrain and the cooler subducting KC plate. Granulites further away from this border were exhumed along the DC paths up to the mid-crust (Perchuk et al., 2000). A U-Pb age on monazite of 2691 ± 0.007 Ma from the Bandelierkop locality, were interpreted as the age of peak metamorphism (Kreissig & Holzer, 1997; Kreissig et al., 2001).

The CZ is the topic of this study and is discussed in detail in the next section.