5 **PLATE TECTONIC MODEL**

5.1 **INTRODUCTION**

As mentioned earlier the ~1100Ma Namaqua-Natal Metamorphic Province (NNMP) is considered as part of the ~1150Ma to ~950Ma Llano-Namaqua orogen during the Grenvillian assembly of the supercontinent Rodinia (Moores, 1991 and Dalziel et al., 2000). The NNMP straddles the central part of South Africa (Figure 5.1) where it bounds the western, southern and SE margins of the Archean Kaapvaal Craton. From here it can be traced to the Falkland Islands of the SW Atlantic Ocean and Dronning Maud land of east Antarctica (Jacobs et al., 1997, Figure 1.1).

![Figure 5.1: The geographic distribution of the Kaapvaal Craton, Kheis Terrane and Namaqua-Natal Metamorphic Province in South Africa (Modified after Moen, 1999).](image)

In the Kwazulu-Natal region of South Africa the Natal Metamorphic Complex (the eastern extent of the NNMP in South Africa) has been obducted onto the Kaapvaal Craton
(Matthews, 1972) and an integrated structural model for this accretionary event has recently been established (Jacobs et al., 1993 and Jacobs and Thomas, 1994) and refined by Jacobs et al. (1997). According to the refined model the accretion of the Natal Metamorphic Complex province of the NNMP onto the south-eastern margin of the Kaapvaal Craton has been the result of a prolonged phase of arc-continent collision that was initiated at ~1200Ma by the deposition and extrusion of supracrustal rocks (Thomas and Eglington, 1990) which was followed by subduction of the Tugela Ocean at ~1180Ma (Jacobs and Thomas, 1996). This resulted in the formation of the juvenile Mzumbe and Margate arc terranes and hot-spot volcanism in the Tugela Ocean (Wilson, 1991). Obduction of the Tugela Ophiolite took place at around 1135Ma (Jacobs et al., 1997) and was followed by NE directed collision at ~1090Ma (Thomas et al., 1995).

In the Northern Cape Province it is believed that the western margin of the Kheis Terrane is tectonically juxtaposed with the Namaqua Complex of the NNMP. In this area the NNMP is currently subdivided from east to west into the Areachap-, Kakamas- and Bushmanland terranes (Thomas et al., 1994) which is in turn tectonically juxtaposed with the 1.7Ga to 2.0Ga old Richtersveld Terrane (Reid, 1982 and 1997) to its west (Figure 5.2). But in contrast to the Natal Complex of the NNMP, the exact timing of the accretion of the different tectonic terranes remains unclear.

A wide range of models have been proposed for the depositional history as well as plate tectonic setting and evolution of units making up the Kheis Terrane and Namaqua Complex of the NNMP. A summary of these interpretations and models is given in Table 5.1 and the reader is referred Appendix A for a detailed summary of these models. In broad terms the geology and deformational history of the Kheis Terrane and the Namaqua province are explained by various adaptations of a plate tectonic model which ultimately involves the collision of the Kaapvaal Craton with either another craton (Geringer et al., 1988), or cratonic fragment (Stowe, 1986) or a collage of cratonic fragments (Thomas et al., 1994) to its west at ~1000Ma (Table 5.1). Exceptions to this theme are a) the models proposed by Malherbe (1979) who suggested an environment dominated by the subsidence of a granitic basement high (Table 5.1) to explain the deposition of the rocks making up the Kheis Terrane, and b) Botha and Grobler (1979) who invoked a rifting-event due to the presence of a mantle plume followed by crustal contraction to explain the deposition and deformation of lithologies to form the so-called Kheis Terrane (Table 5.1).
Figure 5.2: The distribution of the various lithological units, provinces and terranes that comprise the geology of the Kaapvaal Craton, Kheis Terrane and Namaqua Province in the Northern Cape of South Africa. Note the subdivision of the granitoids of the Keimoes Suite into unfoliated (post-tectonic), medium-foliated (syn to post-tectonic) and well-foliated (pre to syntectonic) granites. Also note that the Areachap Group is least affected by the intrusion of the granitoids of any unit to the west of the Kheis Terrane. [Age data indicated on this map from Cornell et al. (1998); Gutzmer et al. (2000); Reid (1982 and 1997); Geringer et al. (1988); Cornell et al. (1986); Cornell et al. (1990), Geringer et al. (1986) and Geringer and Ludick (1990) and this study].

Apart from the various models for the tectonic evolution, a wide range of explanations as to the timing of formation and depositional environments of lithological units forming part of the Kheis Terrane and Namaqua Province have been given throughout the years (Table 5.1). For example, the lavas and sedimentary rocks of the Koras Group (Table 5.1) have been explained as being a result of rifting due to mantle upwelling (Botha and Grobler, 1979),
Table 5.1: Summary of depositional and tectonic models for various parts of the Kheis and Namaqua terranes as interpreted by various authors.

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<td>Keimoes Suite</td>
<td>Stage 4: Intrusion followed formation of island arc complex due to westward directed subduction.</td>
<td>Stage 2: Intrusion due to development of granitic magma during closure of ocean between Kaapvaal Craton and Namaqua plate.</td>
<td>Stage 2: Intrusion due to development of granitic magma during closure of ocean between Kaapvaal Craton and Namaqua plate.</td>
<td>Stage 8: Intrusion associated with subduction along Cape Conductive Belt.</td>
<td>Stage 2: Deposited as sedimentary rocks in fore-arc basin to the west of island arc.</td>
<td>Stage 2: Deposited as sedimentary rocks in fore-arc basin to the west of island arc.</td>
<td>Stage 4: Post-tectonic continental arc sequence.</td>
<td>Stage 2: Intrusion associated with subduction along Cape Conductive Belt.</td>
<td>Stage 3: Sedimentary rocks intruding into rocks deposited in fore- and back-arc basins.</td>
<td>Stage 5: Intrusion associated with subduction along Cape Conductive Belt.</td>
<td>Stage 5: Intrusion associated with subduction along Cape Conductive Belt.</td>
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<td>Korannaland Group</td>
<td>Stage 3: Deposition controlled by reactivation of zone of crustal weakness (suture zone).</td>
<td>Stage 3: Deposition controlled by reactivation of zone of crustal weakness (suture zone).</td>
<td>Stage 6: Sedimentary rocks deposited on passive continental margin after rifting.</td>
<td>Stage 3: Deposition controlled by reactivation of zone of crustal weakness (suture zone).</td>
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<td>Areachap Group</td>
<td>Stage 2: Calc-alkaline volcanic-arc related sequences lain down on juvenile oceanic crust</td>
<td>Stage 3: Island arc complex developed during late stage of westward directed subduction</td>
<td>Stage 1: Subduction of oceanic crust and calc-alkaline volcanism forming island arc</td>
<td>Stage 4: “Korannaland” microcontinent accreted to southwest of island arc (Jannelsepan Formation)</td>
<td>Stage 2: Emplacement as mantle derived volcanics and calc-alkaline granitoids due to closure and subduction of oceanic lithosphere</td>
<td>Stage 1: Pre-tectonic volcanic arc developed due to eastward directed subduction</td>
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<td>Wilgenhoutsdrif Group</td>
<td>Stage 5: Situated on craton side of island arc</td>
<td>Stage 3: Deposited prior to initiation of folding due to presence of Richtersveld arc-massif to the west</td>
<td>Stage 2: Straddling a westward directed subduction zone in a plate tectonic collision</td>
<td>Stage 4: Formed as a flysch-type deposit due to erosion as a result of deformation at ~1400Ma</td>
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<td>Groblershoop Group</td>
<td>Stage 2: Deltaic and shallow water sediment along passive continental margin</td>
<td>Stage 2: Westward thickening clastic wedges, originally part of Kheis Subprovince but deformed and displaced after 1st deformational event</td>
<td>Stage 1: Sedimentary rocks deposited after rifting</td>
<td>Stage 1: Sediment accumulated on continental margins of Kaapvaal craton</td>
<td>Stage 2: Deposited in back-arc basin east of island arc</td>
<td>Stage 1: Overlies Olifantshoek Group rocks conformably in shallow basin formed by subsidence of granitic basement high</td>
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<td>Olifantshoek Group</td>
<td>Stage 1: From braided rivers on the stable shelf adjacent to Kaapvaal Craton</td>
<td>Stage 1: Deposited as westward thickening clastic wedges after extrusion of Hartley lavas</td>
<td>Stage 1: Sedimentary rocks deposited after rifting</td>
<td>Stage 1: Sediment accumulated on continental margins of Kaapvaal Craton</td>
<td>Stage 1: Deposited in back-arc basin east of island arc</td>
<td>Stage 2: Deposited as alluvial fans and beach deposits in eastern side of shallow basin formed by subsidence of granitic basement high after deposition of Griqualand West Sequence</td>
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<td>Hartley Formation</td>
<td>Stage 1: Early bimodal volcanic activity in a continental rift system</td>
<td>Stage 1: Deposited in back-arc basin east of island arc</td>
<td>Stage 1: Sedimentary rocks deposited after rifting</td>
<td>Stage 1: Deposited in back-arc basin east of island arc</td>
<td>Stage 1: Deposited in back-arc basin east of island arc</td>
<td>Stage 2: Deposited as alluvial fans and beach deposits in eastern side of shallow basin formed by subsidence of granitic basement high after deposition of Griqualand West Sequence</td>
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Table 5.1 (continued): Summary of depositional and tectonic models for various parts of the Kheis and Namqua Terranes as interpreted by various authors.
as a post-continental arc sequence (Botha and Grobler, 1979) and that its deposition was controlled by the reactivation of a zone of crustal weakness [possibly along an old suture zone (Moen, 1980)] or by rifting due to isostatic rebound following a folding event (Van Zyl, 1981).

The fact remains that from the literature compilation of Table 5.1 it is clear that there is no comprehensive model for the origin of the Kheis Terrane and the eastern part of the Namaqua part of the NNMP that incorporates and explains the presence of all the main geological units present. It is also important to note that although the depositional setting of the Keis Supergroup in a trailing margin setting has been considered in very simple terms (Jansen, 1983 and Schlegel, 1993), one major question has never been addressed. This question has to do with the source of the tremendous volume of arenaceous sedimentary rocks comprising the ~20km thick (Olifantshoek Group is ~4km thick, Groblershoop Group is ~12km thick and the Wilgenhoutsdrif Group is ~3.8km in thickness) succession of the Keis Supergroup. This is also one of the issues to be addressed in this chapter.

Thus, in summary, it is the purpose of this chapter to compile data and geological observations derived from basin analysis and heavy mineral provenance studies of the sedimentary rocks comprising the Kheis Terrane, U-Pb SHRIMP zircon geochronology of detrital and magmatic zircon grains from the Kheis Terrane and Namaqua section of the NNMP as well as interpretations of satellite imagery, aerial photography and geophysical data into a comprehensive plate tectonic model. This model should explain a) the origin of the Kheis Terrane and its relationship with the Kaapvaal Craton and Namaqua province of the NNMP and b) the deformational history of the eastern part of the Namaqua province during the assembly of the supercontinent Rodinia.

5.2 NEW FINDINGS OF THIS STUDY THAT NEEDS TO BE EXPLAINED IN A PLATE TECTONIC MODEL

Major findings of this study that needs to be considered in a plate tectonic model (or that needs explanation in a plate tectonic context) are as follows:

a) The boulder conglomerate at the base of the Neylan Formation forms a major sequence boundary separating the Lucknow Formation of the Transvaal Supergroup from the
Plate tectonic model

Olifantschoek Group of the Keis Supergroup. This is a major modification of earlier stratigraphic subdivisions that classified the strata of the Lucknow and Mapedi formations with the Olifantschoek Group.

b) The Keis Supergroup, incorporating the Olifantschoek-, Groblershoop- and Wilgenhoutsdrif groups, is a very thick (in the order of ~20km) succession of fluvial to shallow marine platform sediments deposited on the trailing western margin of the Kaapvaal Craton. It is mostly made up of coarse to fine orthoquartzites and quartzites with very little shale, implying major sedimentary by-pass of argillaceous components during deposition.

c) The Keis Supergroup is characterized by a major detrital zircon age mode of 1.85Ga to 2.1Ga for which there is no real source on the Kaapvaal Craton. There is also very little Paleoarchean detrital zircon grains present in the Keis Supergroup, with the other major population being Neoarchean in age. This is not a typical age spectrum for silicic rocks on the Kaapvaal Craton.

d) There is a marked angular unconformity at the base of the Wilgenhoutsdrif Group which implies that the underlying Groblershoop Group was buckled before deposition of the Wilgenhoutsdrif Group.

e) The Sprigg Formation used to be considered the base of the Areachap Group immediately to the west of the Keis Supergroup and east of the Keimoes Suite (Figure 5.2) of the Namaqua Front. However, this study found that the Sprigg Formation is younger that the Areachap Group and should not be classified with it.

f) On the western side of the Namaqua Front, the Goedehoop-, Sandputs- and Collinskop formations all display similar detrital zircon age populations and can be considered part and parcel of the same lithostratigraphic unit, i.e. the Korannaland Group.

g) The Goedehoop Formation of the Korannaland Group in the Kakamas Terrane of the Namaqua-Natal Metamorphic Province display excellent preservation of sedimentary structures, including cross-bedding, which implies very low strain conditions during deformation. It unconformably overlies the Kokerberg gneiss (one of the typical pink gneisses of the Kakamas Terrane) and contains detrital zircon grains that were derived from this gneiss with an age of 1160Ma.

h) The detrital zircon age populations in the quartzites of the Korannaland Group, to the west of the Namaqua Front, are distinctly different from that of the Keis Supergroup, to the east of this boundary. These detrital zircon grains are characterized by a major age mode of ~1200Ma and with virtually no zircon grains older than 2000Ma. This implies that the zircon grains of the Kheis- and Kakamas Terranes were sourced from different areas and no mixing
of detritus took place until the deposition of the Sprigg Formation which contains evidence of zircon grains derived from both terranes.

i) Ar/Ar age determinations on muscovite developed on cleavage planes in fine-grained quartzites and quartz schists of both the Korannaland and Keis successions gave consistent ages in the order of 1116Ma. This represents the age of shallow crustal deformation after deposition of the sedimentary rocks on both sides of the Namaqua Front.

j) The Areachap Group, which has always been considered a volcanic arc complex between the Kheis and Kakamas Terranes, may in actual fact extend to the north and form the boundary between the Kgalagadi and Kheis Terranes (Figure 5.3).

k) Structural elements in the Kakamas and Kheis Terranes indicate left-lateral movement of crustal blocks during the final stages of the formation of the Namaqua-Natal Metamorphic Belt to the west of the Kaapvaal Craton (Figure 5.3).

5.3 BROAD TECTONIC FRAMEWORK

In order to address all the new findings of this study as outlined above, it is thus necessary to work from a very broad tectonic framework for southern Africa. Several such frameworks have been published (Corner et al., 1990, Thomas et al., 1993 and Dorland 2004). These are adopted for this study, based on the major tectonic terranes that can be identified in the aeromagnetic map of southern Africa (De Beers, 1998) in combination with earlier published tectonic terrane maps for the area (Figure 5.4).

The main tectonic elements recognised on this map with relevance to the present study are as follows:

a) The Kaapvaal Craton.

b) The Zimbabwe Craton and adjacent Limpopo and Magondi Belts. From the detrital zircon provenance studies conducted it was realized that the rocks of the Kaapvaal Craton was most probably not a suitable source for the metasedimentary rocks of the Keis Supergroup. Therefore the Zimbabwe Craton, Magondi- and Limpopo Belts have to be considered in more detail and to understand their possible role as source terranes.

c) The Keis Terrane, incorporating the Olifantshoek-, Groblershoop- and Wilgenhoutsdrif Groups of the Keis Supergroup.
Figure: 5.3: Aeromagnetic map indicating the major structural provinces present in the northern part of South Africa, western Botswana and eastern Namibia. Note the extension of the Areachap Group to the north along the Kalahari line. The magnetically stable Kgalagadi Terrane is also easily discernable to the west of the Kalahari line and is surrounded by the Koras-Sinclair-Ghanzi Rift (Borg, 1988). Also note the position of the Proterozoic Okwa basement complex that appears to be displaced to the east along the Makgadikgadi line.
d) The Kgalagadi tectonic domain separated from the Kheis Terrane by the Kalahari line which may represent the extension of the Areachap Group.

e) The Namaqua Front incorporating the intrusive rocks of the Keimoes Suite.

f) The Namaqua Metamorphic Complex with its subdivision into the Kakamas- and Bushmanland Terranes, with the metasedimentary rocks of the Korannaland Group disconformably overlying the Kokerberg gneiss and other correlative pink gneisses of the Namaqua Complex.

g) The Richtersveld Terrane incorporating the Orange River Group and the Vioolsdrif Suite.

h) The Koras Group and other relicts of rifted basins.

i) Late stage shear zones along the Namaqua Front related to left-lateral displacement and deformation of rocks of the successions on both sides of the Namaqua Front.

Figure 5.4: Interpretation of an aeromagnetic map of southern Africa indicating the geological features important in this study (Modified after De Beers, 1998).
5.4 SUMMARY OF MAJOR FEATURES OF DIFFERENT TERRANES AND ROCK UNITS

5.4.1 KAAPVAAL CRATON

To understand the role of the Kaapvaal Craton and its Archean to Paleoproterozoic cover rocks as source area for the metasedimentary rocks of the Keis Supergroup, the evolution of the Kaapvaal Craton must be investigated. Poujol et al. (2003) subdivided the Kaapvaal Craton (Figures 5.5 and 5.6A) into 4 domains. These domains are important as they contain possible source rocks for the sediments deposited on the western margin of the Kaapvaal Craton. According to Poujol et al. (2003) the oldest rocks in the Kaapvaal Craton are located in the Ancient Gneiss Complex in the eastern part of the craton. The eastern domain experienced plutonic and volcanic activity between ~3550Ma and 3250Ma, during the time of the formation of the Barberton greenstone belt. The central domain contains xenocrysts with an age of 3480Ma to 3425Ma while the oldest xenocryst in the northern domain was dated at 3364Ma. In the western domain the oldest xenocrysts are 3428Ma old. After the development of the Barberton Greenstone Belt between 3550Ma and 3200Ma granitoid emplacement took place between 3200Ma and 3170Ma in the central, northern and western parts of the Kaapvaal Craton. This was also the time of formation of the Pietersburg and Kraaipan greenstone belts. A major event of granitic intrusion took place throughout the entire Kaapvaal Craton at around 3110 to 2970Ma. According to Poujol et al. (2003) this period of igneous activity corresponds to the initiation of volcanism in the 3100Ma old Madibe and Muchison greenstone belts. This was also the time when the central part of the craton experienced the eruption of the Dominion Group volcanics. At the same time as the eruption of the Dominion Group volcanics, granitic plutons intruded into the Murchison and Pietersburg greenstone belts and the central part of the craton. At around 2980Ma to 2970Ma igneous activity was again experienced in the Murchison greenstone belt in the northern part of the Kaapvaal Craton as well as in the Pongola Supergroup, while the rest of the craton experienced nearly no magmatic activity up to ~2800Ma. Numerous plutons were emplaced into the northern and western domains from about 2880Ma to 2820Ma, the time of the deposition of the Central Rand Group of the Witwatersrand Supergroup.

According to Kositcin and Krapez (2004) the detrital zircon grains from the West Rand- and Central Rand Groups of the Witwatersrand Supergroup indicate that felsic rocks in the source area of the Witwatersrand basin formed between ~3090Ma to 3060Ma and ~3000Ma to
2870Ma. According to Nhleko (2003) the detrital zircon grains in the Mozaan Group of the Pongola Supergroup records magmatic events with ages of 3600Ma to 2890Ma.

Magmatic activity also took place to the north of the craton between 2780Ma and 2770Ma (Poujol et al., 2003). The Amalia greenstone belt and the Ventersdorp Supergroup were deposited during a final pulse of magmatic activity on the Kaapvaal Craton in the time period between 2750Ma and 2700Ma (Eglington and Armstrong, 2004, Poujol et al., 2003, and Tinker et al., 2002).

According to Eglington and Armstrong (2004) the protobasins of the Transvaal Supergroup (Figure 5.5) were formed between 2.65Ga and 2.6Ga at the same time as the intrusion of extensive granitoids associated with the Limpopo orogeny along the northern part of the Kaapvaal Craton. The Pretoria Group of the Transvaal Supergroup was deposited between 2.4Ga and 2.1Ga (Eglington and Armstrong, 2004). According to Eglington and Armstrong (2004) the deposition of the Transvaal Supergroup was terminated at ~2.1Ga with the extrusion of the volcanics of the Dullstroom Formation and the Rooiberg Group. This was followed by the intrusion of the Bushveld Complex at ~2.06Ga (Walraven, 1997).

5.4.2 ZIMBABWE CRATON

In very general terms the Zimbabwe Craton is made up of the 3.5Ga old gneisses and Sebakwian granite-greenstone complexes of the Tokwe segment (Figure 5.6B). A second period of granite–greenstone development took place between 3.0Ga and 2.8Ga (Wilson, 1979), forming the Belingwean- and Lower Bulawayan Groups. A third period of granite-greenstone development, represented by the Upper Bulawayan- and Shamvaian Groups took place at ~2.7Ga. Large parts of the Zimbabwe Craton is crosscut by the 2574±7Ma (Armstrong and Wilson, 2000) old Great Dyke over a distance of 530km (Figure 5.6B). It is also important to note that apart from the sedimentary rocks of the 2.67Ga to 2.66Ga old Shamvaian (Huizenga, 1995) Group the Zimbabwe Craton is nearly devoid of any cover sequences.
Figure 5.5: Proterozoic geological framework of the Kaapvaal and Zimbabwe Cratons. Note the location of the Limpopo-, Magondi- and Namaqua-Natal Belts (Modified after Hanson et al., 2004). The Archean make-up of the Kaapvaal and Zimbabwe Cratons is shown in Figure 5.6.
Figure 5.6: The Archean make-up of the Kaapvaal (A) and Zimbabwe (B) Cratons of southern Africa (Modified after Poujol *et al.*, 2003 and Huizenga, 1995).
5.4.3 LIMPOPO AND MAGONDI BELTS

The Limpopo Belt is an east-northeast trending zone of granulite facies rocks that separates the granite-greenstone terranes of the Kaapvaal-and Zimbabwe Cratons. (McCourt and Armstrong, 1998) (Figures 5.5 and 5.6). The Limpopo belt is subdivided into the Southern Marginal Zone, the Central Zone and the Northern Marginal Zone. The Northern Marginal Zone is composed of igneous charnockite and enderbite that intruded ultramafics, mafic, felsic and metasedimentary units between 2710±38Ma and 2637±19Ma (Berger et al., 1995). The Southern Marginal Zone consists of granulite facies supracrustal gneisses that were intruded by 2664Ma to 2671Ma old granitoids (Retief et al., 1990 and Barton and Van Reenen., 1992). Three main high grade events are recognised in the Central Zone of the Limpopo Belt. These occurred at about 3200Ma to 3100Ma, 2650 to 2520Ma and at 2000±5Ma (Boshoff, 2003 and Holzer et al., 1998).

The Magondi Supergroup consists mainly of metasedimentary rocks and minor amounts of mafic and intermediate to felsic metavolcanics. It is subdivided into the Deweras-, Lomagundi- and Piriwiri Groups which were deposited between 2.1.Ga and 2.0Ga ago (Loney, 1969 and Master, 1991) (Figures 5.5 and 5.6). The Magondi Supergroup was deformed and metamorphosed during the Magondi orogeny between 2.0Ga and 1.8Ga. Treolar and Kramers (1989) obtained a Rb-Sr whole rock age of 1890±260Ma for charnockites and 1780±280Ma for migmatites in the northern part of the Magondi Belt. Munyaniwa et al., (1995) obtained U-Pb and Pb-Pb single zircon ages for charno-enderbites in the Magondi belt, which constrains the emplacement of these bodies between 1930Ma and 1960Ma.

5.4.4 KEIS SUPERGROUP

5.4.4.1 Olifantshoek Group

The angular unconformity at the base of the Neylan Formation (Figure 5.7) marks a dramatic change from the environment that prevailed during the deposition of the Lucknow Formation of the Transvaal Supergroup. The stable, shallow marine environment that was subjected to a period of low clastic input and therefore suitable for the development of immature carbonate platforms (as indicated by the presence of stromatolites and carbonate oolites in the Lucknow
Formation) was transformed into a rifted continental margin at ~1928Ma with outflow of the bimodal lavas of the Hartley Formation (Cornell, 1987 and Cornell et al., 1998).

Cornell et al. (1998) argued that the Waterberg-Soutpansberg branch of the rift system failed to develop an oceanic basin while the Matsap branch along the western margin of the
Kaapvaal Craton developed into a rift wherein the sedimentary rocks and lavas of the Hartley Formation were deposited.

**Figure 5.8:** Simplified map indicating the distribution of the lithological units making up the Keis Supergroup, Areachap Group, Korannaland Group and Koras Group in the Northern Cape Province of South Africa. The intrusive rocks belonging to the Keimoes Suite have been omitted from this map. Their distribution can be seen in Figure 5.2.
The transition from the Hartley Formation to the lowermost part of the Volop Formation of the Olifantshoek Group is gradational. The deposition of the sedimentary rocks of the Olifantshoek- and Groblershoop Groups of the Keis Supergroup is related to the development of a stable continental margin (Boilott, 1981) on the western side of the Kaapvaal Craton after the rifting event at 1928Ma.

Figure 5.9: Schematic representation of U-Pb SHRIMP ages obtained from detrital zircon grains from the Keis Supergroup and Korannaland Group. Note that the Sprigg Formation is show on its own, and not anymore associated with the Areachap group. The age of the youngest zircon grains obtained from each unit are indicated which may be an indication of the maximum depositional age from each unit. Note the absence of Archean zircon grains from the sedimentary rocks from the Areachap Group and the Korannaland Group, which is in sharp contrast to the sedimentary rocks derived from the Kaapvaal Craton.

The sedimentary rocks of the Volop Formation were most probably deposited in a braided river type of environment. Paleocurrent directions in the lower Fuller Member of the Volop
Formation were from the northeast and gradually changed into a more westerly direction during the deposition of the Ellies Rus- and Glen Lyon Members. This changeover may be a function of the erosion of the restrictive ridges of the earlier rifted margin allowing sediment to be transported from the rift margin into the developing and widening marine basin.

This change in source areas is also confirmed by changes observed in the concentrations in heavy minerals from the Fuller- to the Ellies Rus and Glen Lyon Members. U-Pb SHRIMP dating of detrital zircon grains from the Glen Lyon Member yielded two main age populations of 1.9 to 2.3Ga and 2.6 to 2.9Ga (Figure 5.9) with the youngest zircon being 1950±27Ma (7% discordant).

The upper part of the Verwater Member becomes coarser grained and gradually gives way to the upward coarsening, grey-coloured, texturally more mature quartzites and orthoquartzites of the Top Dog Formation (Figures 5.7 and 5.8). The Top Dog Formation contains abundant symmetric ripple marks as well as imprints of mud cracks. Facies relationships as well as the bimodal to polimodal westerly-directed transport directions imply that the Top Dog Formation represent a gradual changeover from the fluvial depositional environment that dominated the deposition of the Volop Formation to the shallow marine environment of the overlying Groblershoop Formation. Detrital zircon grains from the Top Dog Formation are very similar in age but has a slightly more uniform distribution than those of the Glen Lyon Member in its age populations, with two age populations present at i.e. ~2Ga to 2.1Ga and 2.6Ga to 2.8Ga (Figure 5.9).

5.4.4.2 Groblershoop Group

Above the Top Dog Formation the upward coarsening sequences of the Groblershoop Group appear, grading from fine-grained quartz-sericite schists to coarser-grained quartzites and orthoquartzites. This change in depositional environment as well as the mineralogy of heavy mineral concentrations present in the sedimentary rocks of the Groblershoop Group, indicate the development of a stable continental shelf that is affected by marine transgressions and regressions. The fine-grained material represent sea level high stands and the coarser-grained quartzite units sea level low stands resulting in progradational deposition of shallow marine material. This succession may thus well represent the final stage in the development of a stable continental margin, the “Atlantic phase” of Boilott (1981). A minimum U-Pb SHRIMP
age of $\sim 1825 \pm 21$ Ma from a detrital zircon from the Viulnek Formation suggests that its deposition probably took place $\sim 100$ Ma after the initiation of rifting at 1927 Ma.

Similarities in the age populations of detrital zircon grains from the Olifantschoek- and Groblershoop Groups may indicate continuous reworking of sediment derived from the Kaapvaal Craton and sediment supply from a common source area, while the younger zircon grains (1.8 to 2.0 Ga) from the Viulnek Formation probably suggest the first acquisition of detritus from another source area. This may be the time that the Magondi Belt became a source for the sedimentary rocks of the Keis basin.

As previously noted the corrected stratigraphic thickness of the sedimentary rocks comprising the Groblershoop Group is between 12 km and 15 km, which can be subdivided into the $\sim 1500$ m thick Faanshoek Formation, $\sim 400$ m thick Faansgeluk Formation, 1300 m thick Maraisdraai Formation, the 1700 m thick Viulnek Formation, the $\sim 4000$ to 7000 m of mostly sedimentary rocks of the Opwag Formation (which can internally be even further subdivided) which is overlain by the 3100 m thick Skurweberg Formation. No major erosional or reactivation surfaces are present between these units which were defined on the basis of their grain size. According to Lemon (1990) and Bond et al. (1995) sediments deposited in Atlantic type continental settings can be between 8 and 13 km thick and are made up of seaward thickening wedges that are separated by undisturbed planar horizons in their shallow water sections (Pitman, 1978). It is suggested that the geological record preserved in the rocks of the Groblershoop Group represent an oblique section through a series of prograding sedimentary wedges, which were each affected by marine transgressions and regressions. It is also important to note the presence of the $\sim 1290$ Ma old (Moen, unpublished data) Kalkwerf granitic gneiss which appears to have been deformed with the sedimentary rocks of the Groblershoop Group.

5.4.4.3 Wilgenhoutsdrif Group

A change in geological conditions is preserved by the sedimentary rocks and interbedded lavas of the Wilgenhoutsdrif Group. The Wilgenhoutsdrif Group (Figures 5.7 and 5.8) is subdivided into the $\sim 900$ m thick basal Grootdrink Formation, composed of light-grey orthoquartzite. Detrital zircon grains from the Grootdrink Formation produced a single age population of 1.8 Ga to 2.1 Ga (Figure 5.9) and one older zircon having an age of 2630 $\pm 12$ Ma.
Plate tectonic model

(7% discordant). The base of the Grootdrink Formation is erosive towards the rocks of the underlying Groblershoop Formation. This erosional surface may also have acted as a décollement surface during later deformational events. The upper-most part of the Skurweberg Formation of the Groblershoop Group, the Witkop Member, is eroded by the Grootdrink Formation in nearly all localities, except in the Kaaian Hills to the southeast of Upington. This probably indicates that the unconformity at the base of the Grootdrink Formation could be angular and may well be the result of crustal warping or buckling as a fore bulge developed due to the encroachment of a volcanic arc (Areachap Group) from the west. The Grootdrink Formation gradually changes into the quartzites, interbedded carbonates and serpentinite lenses of the ~3500m thick Zonderhuis Formation. A hydrothermally altered jasper breccia appears towards the top of the Zonderhuis Formation. The shale preserved in the top of the Zonderhuis Formation may indicate that the basin in which the sedimentary rocks of the Wilgenhoutsdrif Group were deposited continued to deepen as the crust continued to buckle, a result of the encroachment of the Areachap volcanic arc.

The two volcanic cycles comprising the Leerkrans Formation overlie the Zonderhuis Formation with what appears to be a conformable contact (Moen, 1980). The first cycle consists of quartz-feldspar porphyries overlain by metamorphosed basaltic lavas containing pillow lava structures, lapilli and calcite-filled vesicles. The second cycle commences with a felsic pyroclastic deposit that grades into a reworked tuffaceous unit containing quartzite pebbles and thin lenses of quartzite and basic pyroclastic rocks (Moen, 1980). According to Barton and Burger (1983), Moen (1980) and Stowe (1986) the deposition of the sedimentary rocks and lavas of the Wilgenhoutsdrif Group is related to an active continental plate margin, where they were deposited in either a back-arc or fore-arc environment. Stowe (1986) went even further in stating that the rocks of the Wilgenhoutsdrif Group were deposited on the craton side of an island arc. Moen (1999) suggested that these rocks were deposited in an unstable basin that was filled by alternating deep and shallow water siliciclastic deposits, intermittently interrupted by basic volcanism that was followed by large volumes of bimodal volcanism. This probably occurred in a back- or fore-arc basin related to an active continental margin (Cornell et al., 1986, Geringer et al., 1986 and Geringer and Ludick, 1990).

A U-Pb SHRIMP age of ~1290Ma was obtained from the lavas of the Leerkrans Formation that overlay the shales of the Zonderhuis Formation (Moen, unpublished data). These lavas and associated volcanoclastic deposits probably indicate the first direct input from the
volcanic arc that was deposited in the unstable basin as envisaged by Moen (1999). With the
volcanic arc represented by the Areachap Group to its west and the passive continental margin
of the Kaapvaal Craton to its east, the sedimentary rocks, lavas and volcanoclastics of the
Wilgenhoutsdrif Group were therefore most probably deposited in a back-arc basin setting.

The same age of 1290Ma (Moen, unpublished data), as obtained for the lavas of the Leerkrans
Formation, was also determined for the granitic Kalkwerf Gneiss suggesting that these
lithologies could be intrusive and extrusive correlates of each other. The Kalkwerf Gneiss
is intrusive into the rocks of the Groblershoop Group. This may imply a period of easterly
directed subduction, below the Kaapvaal Craton. It is also important to note that the Kalkwerf
Gneiss was deformed with the rocks of the Keis Supergroup, and therefore predates the
defformation of the Kheis Terrane, a observation supported by $^{40}$Ar/$^{39}$Ar dates obtained from
muscovite in the Kheis Terrane indicating fabric developed at shallow depth around 1131Ma
to 1112Ma.

5.4.5 AREACHAP GROUP

The meta-igneous rocks of the Jannelsepan Formation and meta-pelitic rocks of the Bethesda
Formation (Geringer and Ludick, 1990) as well as the apparently cross-cutting kinzigites of
the Rateldraai Formation, make up the Areachap Group that occurs to the west of the
Dagbreek Formation and the so-called Triorolspan shear (Figures 5.2 and 5.8). The poorly
outcropping Areachap Group is believed to extend from the north of Upington in a
southeasterly direction (Figure 5.7) all the way down to Prieska in the south (Cornell et al.,
1990) where it hosts the Prieska massive sulphide deposits. The study and interpretation of
gravity and magnetic data of the area occupied by the Kheis Province, Areachap Group,
Keimoes Suite, Kakamas- and Richtersveld Terranes allows for the extrapolation of the rocks
comprising the Areachap Group from its present outcrop area around a larger scale
manifestation of the Karos anticline and the Orange River syncline (as indicated on
(Figures 5.3 and 5.11). Here it can be correlated with the Kalahari line that continues into
western Zambia (McMullan, 1998; Meixner et al., 1985 and Reeves 1979). In Botswana the
Kalahari line is characterised by a prominent magnetic and gravity lineament and has been
interpreted to be a mid-Proterozoic rift which is separated from the Kaapvaal Craton in the
east by a belt of north-south trending magnetic anomalies, thought to be the folded rocks of
the Kheis Terrane (McMullan, 1998 and Reeves, 1979). Very little is known about the geology of the Kalahari line itself apart from the fact that it is intruded by the mafic dykes and sills of the Mesoproterozoic Tshane Complex (McMullan, 1998).

Two distinctly different amphibolite units occur within the Jannelsepan Formation in the Upington area, with the lower Donkerhoekspruit Member being a massive hornblende and biotite amphibolite and the upper Swartkop Member being pyroxene bearing (Geringer and Ludick, 1990). Not much is known about the relationship between the amphibolites of the Jannelsepan Formation and the rocks of the Bethesda- and Rateldraai Formations, mainly due to the intrusion of the late-tectonic granites of the Keimoes Suite (Geringer et al., 1988 and 1990). The same can be said about the origin of the metamorphosed sedimentary rocks of the Bethesda Formation and the kinzigites of the Rateldraai Formation. Trace element concentrations as well as inter-element ratios from the Jannelsepan Formation indicate that the massive amphibolites represent a low K-tholeiite and calc-alkaline basalt while the pyroxene amphibolite represents metamorphosed shoshonite and high-K volcanics that formed as a part of a volcanic arc system (Geringer and Ludick, 1990), either in a back-arc environment or a mature arc setting (Table 5.1).

The amphibolites of the Jannelsepan Formation give Rb-Sr age data that yielded errorchron ages of 1115±228 Ma and 1268±246 Ma (Barton and Burger, 1983). Cilliers (1987) obtained a Pb-Pb isochron age of 1665±133 Ma for the amphibolites of the Jannelsepan Formation in the Boksputs area, while Theart (1985) reported an age of 1599±220 Ma. This data led Cornell et al. (1986) and Geringer and Ludick (1990) to suggest that the precursors to the amphibolites of the Jannelsepan Formation originated somewhere between 1600 and 1350Ma. A whole-grain thermal evaporation $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1285±14Ma on zircon grains from the Smouspan Gneiss and Copperton Formation indicates that the Copperton Formation must have existed as an island arc complex for more than 100Ma before it was involved in the tectonometamorphic event that led to its intense deformation (Cornell et al., 1990).

5.4.6 KGALAGADI TERRANE

The northward extension of the Areachap Group, and in particular the amphibolites of the Jannelsepan Formation, hold fundamental implications as to the tectonic evolution of the western margin of the Kaapvaal Craton. If it is correct to assume that the amphibolites of the
Jannelsepan Formation represent the remnants of a volcanic arc, this arc terrane may extend to the north along the so-called Kalahari line, separating the Kgalagadi- and Kheis Terranes.

It is clear from magnetic data that the Kgalagadi Terrane (not to be confused with the Kgalagadi Province as defined by Stowe, 1984), situated to the west of the northward extension of the Areachap Group in the northern part of South Africa, eastern Botswana and western Namibia (Figures 5.3, 5.4 and 5.11), has a cratonic character (Eberle et al., 2002 and Reeves, 1979). Reeves (1979) modelled that the cratonic material are buried underneath ~15km of cover to the west of the Kalahari line. This cover represented by the sedimentary rocks of the Nama Group, Dwyka Formation of the Karoo Supergroup and the Kalahari Formation.

Figure 5.10: Schematic interpretation of aeromagnetic and gravity data indicating the inferred distribution of the Areachap Group towards the north were it becomes the Kalahari line in Botswana.
The only indications of older crustal material that could be part of the Kgalagadi Terrane is the Okwa Basement Complex in Botswana and the Kgalagadi Basement Inlier in Namibia. The felsites and sericitic quartzites of the Okwa Basement Complex were intruded by granites between 2.1 and 2.0Ga ago (Aldiss and Carney, 1992). It may have been displaced to the east along the Makgadikgadi line as this line also terminates the Kalahari line to the south of the Okwa Complex (Figure 5.3). According to Ziegler and Stoessel (1993), the granitoids and amphibolites of the Kgalagadi Basement Inlier were formed between 2.6 and 1.8Ga as an accretionary belt on the northwest side of the Kaapvaal-Zimbabwe Craton. It was intruded by the Weener and Piksteel Intrusive Suites at around 2.0 and 1.8Ga, the Alberta Mafic Complex at ~1.4Ga and the Gamsberg Granite Suite between 1.25 and 1.0Ga.

The stable geophysical nature of the Kgalagadi Terrane (Figures 5.3 and 5.4) containing the >2.1Ga Okwa Basement Complex and the 2.6 to 2.0Ga Kgalagadi Basement Inlier (Ziegler and Stoessel, 1993) as well as the extension of the Areachap Group island arc to its east suggest that it may represent a separate cratonic fragment that were accreted onto the western margin of the Kaapvaal Craton soon after 1.29Ga (age of the Kalkwerf gneiss and Leerkrans lava, unpublished data of Moen). The idea that the Kgalagadi Terrane is a separate cratonic fragment is supported by the interpretation of the regional magnetic data set of Namibia whereby the western rim of the Kaapvaal-Zimbabwe Craton is clearly marked by the northwesterly-southeasterly orientated Sinclair domain (Eberle et al., 2002) which forms part of the Koras-Sinclair-Ghanzi Rift (Borg, 1988).

5.4.7 RICHTERSVELD AND KAKAMAS TERRANES

The metamorphosed and tectonized rocks of the Richtersveld Terrane (Figure 5.2) represent the oldest known rocks in the NNMP of the Northern Cape Province in South Africa and southern Namibia. It is composed of the volcanic rocks of the 2.0Ga Orange River Group (Reid, 1997) which are intruded by the 1.7 to 1.9Ga old granitoid batholith of the Vioolsdrif Suite (Reid, 1982). The Vioolsdrif batholith exhibits a two-stage history with the intrusion of differentiated basic complexes at 1900±30Ma that was followed by the emplacement of adamelite and leuco-granite at 1730±20Ma (Reid, 1982). The volcanic and igneous rocks of the Orange River Group and the Vioolsdrif Suite are interpreted to have developed in a mature arc environment that was incorporated into the NNMP as part of an accretionary complex of separate terranes (Reid, 1997).
The 1.7 to 2.0Ga Richtersveld Terrane is interpreted to have been intruded by the younger pink granitic gneiss, the Kokerberg gneiss (also referred to as the Riemvasmaak Formation by Geringer and Botha, 1976a, 1976b and as indicated on Geological Map 2820 Upington, Geological Survey of South Africa, 1988) of the Kakamas Terrane (Figures 2 and 4). Frick and Wheelock (1983), Kröner (1971), Moore (1977) and Von Backström (1964) provided evidence for a sedimentary origin for the pink gneiss of the Kokerberg gneiss, but Lipson (1980) concluded that it represents a granite with geochemical affinities that are different from the neighbouring augen gneiss. This matter is complicated by the fact that the pink granitic gneisses outcropping throughout the Bushmanland and Kakamas Terranes may not be correlatives of each other and may even have different pre-metamorphic precursors (Frick and Wheelock, 1983). In this study the Kokerberg gneiss have been dated at 1166±15Ma. It contains only one population of euhedral zircon grains strongly supporting an magmatic origin.

The Kokerberg gneiss is directly and nonconformably overlain by at least three upward coarsening cycles of the Puntsit Formation of the Korannaland Group. These cycles are composed of either flat-laminated amphibolites or phyllites overlain by calc-silicates that grade upward into poorly sorted quartzites to arkosic quartzites. The topmost part of the Puntsit Formation is in turn overlain by a thick succession of medium to poorly sorted arkosic to orthoquartzites of the Goedehoop Formation. The quartzites of the Puntsit- and Goedehoop Formations have not experienced high strain deformation and display well developed planar and trough cross-bedding, although they occur in an area that forms part of the NNMP.

As mentioned earlier the Kokerberg gneiss yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1166±15Ma Detrital zircon grains from the Goedehoop Formation produced two distinct age populations of 1.1 to 1.4Ga and ~1.8Ga (Figure 5.9) with the youngest zircon having a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1114±57Ma. The youngest zircon age from this formation is within error of the age of the pink granitic gneiss of the Kokerberg gneiss. This observation combined with the similarities in the zircon morphologies from the Kokerberg gneiss and Goedehoop Formations, together with the presence of a nonconformable contact between the Kokerberg gneiss and the Puntsit- and Goedehoop Formations provide clear evidence that the Puntsit and Goedehoop Formations were derived from the Kokerberg gneiss.
The calc-silicates and sedimentary rocks of the Sandputs Formation appear to have the same relationship with the Kokerberg gneiss as the Puntsit- and Goedehoop Formations, except that where the latter occur to the east of the Neusspruit shear zone, the outcrop area of the Sandputs Formation is restricted to a narrow region to its west (Figure 5.7). In this study it suggested that this is merely an artefact of previous geological interpretations and that the Sandputs Formation may be a correlative of the Puntsit- and Goedehoop Formations as noted by Geringer (1973). Detrital zircon grains from the Sandputs Formation yielded two age populations, i.e. 1.1 to 1.6Ga and 1.7 to 2.0Ga (Figure 5.9) with the youngest zircon having an age of 1176±77Ma (-5% discordant) that is within error the age of the zircon grains from the pink granitic gneiss of the Kokerberg gneiss.

The garnet-sillimanite-cordierite-biotite gneiss and quartz-feldspar gneiss of the Collinskop Formation (Geringer and Botha, 1973a) outcrop sporadically from Namaqualand in the west (Joubert, 1971) up to the Keimoes-Kakamas area in the east (Von Backström, 1964) where they have been described as kinzigites. In the area to the north of Keimoes these rocks appear to have an intimate relationship with granitoids that seem to belong to the Keimoes Suite and occur as circular basin-like structures with the outer limits defined by the inward dipping kinzigite (indicated as the Melkbosrant dome, Geological Map 2820 Upington, Geological Survey of South Africa, 1988) that form metamorphic aureoles around the granites. According to Geringer and Botha (1973b) the precursor of this rock may have been a carbonate-free pelitic rock and its sedimentary nature was confirmed by the presence of two zircon age populations during U-Pb SHRIMP dating, i.e. 1.1 to 1.5Ga and 1.7 to 1.9Ga (Figure 5.9) with the youngest zircon having an age of 1148±11Ma (9% discordant) which is again within error the same age as the zircon grains from the pink granitic gneiss of the Kokerberg gneiss. It is interesting to note that more kinzigites occur to the south of the Orange River to the southwest of Upington, but here they have been included into the Areachap Group as the Rateldraai Formation.

Authigenic muscovite defines a well developed foliation that occur semi-parallel to the sedimentary layering in the quartzites of the Goedehoop Formation of the Korannaland Group and yields well-defined $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages ranging of between 1111±3Ma to 1121±3Ma with a weighted mean average of 1113±3Ma.
5.4.8 KEIMOES SUITE

The Keimoes Suite (Figure 5.2) consists of at least 13 individual granitic plutons and appears to be bimodal in its nature with compositions ranging from granodiorite to alkali-granite with a small number of gabbroic intrusions also being present (Geringer et al., 1988) and occurs along the eastern marginal zone of the NNMP (as defined by Blignaut et al., 1983).

According to Geringer et al. (1988) outcrops of the Keimoes Suite strike in a northwesterly direction for more than 200km from Putsonderwater in the south to the northwest of Upington. The Keimoes Suite is bordered by the Waterval Thrust in the west and the Brakbosch Fault in the east (Figure 5.2). The granitoids are intrusive into the metasedimentary rocks, calc-silicates and granitic gneisses of the Korannaland Group as well as the rocks of the Areachap Group with the exception of the Sprigg Formation to the east. The intrusion of the Keimoes Suite took place during a considerable time period in the development of the Namaqua event. It started with the intrusion of metagabbros, followed by syntectonic- late-tectonic- and late to post tectonic granites and charnockites (Geringer et al., 1988; Slabbert, 1985; Stowe, 1983 and Van Zyl, 1981) at around 1100Ma ago (Geringer et al., 1988).

According to Geringer and Ludick (1990) the Keimoes Suite granitoids intruded as a result of the development of large amounts of calc-alkaline granite magma during the closure of an ocean. This event followed on the subduction of oceanic crust that led to the formation of an island arc sequence, represented by the amphibolites of the Jannelsepan Formation. These magmas then intruded into the highly metamorphosed and deformed rocks of the Areachap Group and Korannaland Group prior to the formation of the transcurrent Neusspruit-, Strausheim-, Boven Rugzeer and Trioolspan shears.

An examination of the granitoids comprising the Keimoes Suite indicates that a large amount of pre- to syntectonic bodies are concentrated along a northwest-southeast directed axis to the west of Upington (Figure 5.2). To their west undeformed post-tectonic bodies (mainly charnockitic in composition) occupy a position parallel to that of the pre- to syntectonic granitoids. It is also clear that the bulk of the granitoids intrude into the gneisses of the Richtersveld- and Kakamas Terranes and the sedimentary rocks of the Korannaland Group (Figure 5.2).
Regarding the timing of their intrusion it must be noted that the well-foliated granitoids of the Keimoes Suite intrude into the sedimentary rocks of the Goedehoop Formation, which was derived from the ~1166Ma old Kokerberg granitic gneiss, and resulted in the formation of muscovites dated at ~1113Ma. Therefore the intrusion of the well-foliated, syntectonic granitoids of the Keimoes Suite postdate the initial intrusion of the Kokerberg gneiss by ~50Ma.

Charnockitic rocks from the NNMP in Namaqualand, Natal, and Antarctica have “within plate” chemical signatures and were probably derived from fractionalised basaltic magmas that intruded into lower to middle crustal levels in an extensional environment (Grantham et al., 2001). The Friersdale charnockites, forming part of the Keimoes Suite, exhibit an I-type signature and an age Rb/Sr age of ~1087Ma (Barton and Burger, 1983). They were probably intruded along or close to a zone of crustal instability (suture zone) between continental blocks amalgamated as a result of the formation of the Namaqua Province of the NNMP.

The kinzigites of the Rateldraai Formation of the Areachap Group share a similar relationship with the granitoids of the Keimoes Suite (as indicated on Geological Map 2820 Upington, Geological Survey of South Africa, 1988) as the kinzigites of the Collinskop Formation of the Goedehoop Formation (Figure 5.7). This relationship is probably very similar to the link observed between the kinzigites of the Collinskop Formation and the granitoids of the Keimoes Suite in the Melkbosrants dome, and in this case the Rateldraai Formation appears to occur as a metamorphic aureole around strongly foliated syntectonic granitoids of the Keimoes Suite that intrudes into the rocks of the Bethesda Formation. It is therefore suggested that the kinzigites of the Rateldraai Formation probably also represent metamorphosed pelitic rocks but unfortunately no age data on detrital zircon grains exist to validate its inferred sedimentary nature or true stratigraphic position.

5.4.9 KORAS GROUP

The ~7000m thick bimodal succession of basalt, quartz-porphyries and immature sediment of the Koras Group (Gutzmer et al., 2000, Grobler et al., 1977, Moen, 1987 and SACS, 1980) unconformably overlies the tectonically folded and foliated sedimentary rocks and lavas of the Wilgenhoutsdrif Group (Figure 5.2) as well as the highly foliated sedimentary rocks of the Bethesda Formation of the Areachap Group (Gutzmer et al., 2000). Geographically the
outcrop area of this unit can be subdivided into a northern, central and southern domain (Moen, 1987) and stratigraphically into the quartzites and conglomerates of the basal Christiana Formation that is overlain by the andesites and basalts of the Boom Rivier Formation. The Bossienek conglomerate and quartzite follows on the Boom Rivier Formation and are in turn overlain by the 1500m thick Swartkopsleegte Quartz Porphyry Formation from which Gutzmer et al. (2000) obtained a U-Pb zircon SHRIMP age of 1172 ± 7 Ma.

Grobler et al. (1977) suggested that the deposition of the sedimentary rocks of the Koras Group occurred as alluvial fans along the flanks of fault scarps in narrow graben-like basins that also controlled the extrusion of the basic and acid lavas. According to Borg (1988) the Koras Group forms part of a series of basins which formed along two branches of a propagating rift system (the Koras-Sinclair-Ghanzi Rift), with rifting being initiated in the south (Koras Group) and propagating to the north over time.

The depositional environment and plate tectonic setting of the sedimentary rocks and volcanics of the Koras Group represents holds major implications for the development of a plate tectonic model for the Kheis- and Kakamas Terranes.

First of all, weak to poorly deformed rocks of the Koras Group seem to overly the folded strata of the Wilgenhoutsdrif Group as well as the possible extension of the amphibolites of the Jannelsepan Formation to the northeast of Upington. Its present day outcrop patterns seem to be controlled by the much younger Brakbosch and Blauwboschpan faults giving the impression that its depositional basins were orientated in a northwesterly to southeasterly direction.

Secondly, the ages obtained from the granitic gneiss of the Kokerberg gneiss and from muscovites from the Goedehoop Formation have been interpreted as to indicate that the final collisional event between the Kaapvaal Craton and the Richtersveld cratonic fragment occurred after ~1166 Ma and possibly at ~1113 Ma. Therefore Gutzmer et al.’s (2000) well constrained age of ~1172 Ma for the extrusion of the quartz porphyries of the Swartkopsleegte Formation indicates that the sedimentary rocks and volcanics had to be deposited prior to the formation of the NNMP and the deformation of the rocks of the Keis Supergroup into the Kheis Terrane.
One possibility that may explain the undeformed nature of the Koras Group, is that it was deposited in an aulacogen, probably at a high angle to an oceanic basin or in a failed rift setting. It is therefore acceptable to retain the proposed model of Borg (1988) whereby the Koras Group forms part of the Koras-Sinclair-Ghanzi Rift which developed as two rifted basins aligned parallel to the margins of the Kaapvaal Craton. If this is the case, and if the age obtained from the Koras Group (~1172Ma, Gutzmer et al., 2000) is taken into account, it provides clear evidence for a major rifting event affecting the western margin of the Kaapvaal Craton ~60Ma before the final collision between the Kaapvaal Craton and the Richtersveld cratonic fragment.

5.4.10 SPRIGG FORMATION

Detrital zircon grains extracted from the Sprigg Formation are as young as 1150±13Ma and up to 1400Ma in age. These ages are comparable to the ages of detrital zircon grains from extracted from the Goedehoop Formation that was ultimately derived from the pink granitic gneiss of the underlying Kokerberg gneiss. The presence of these young zircon grains may indicate that the sedimentary rocks of the Sprigg Formation had to be deposited after the intrusion of the Kokerberg gneiss or at least after the deposition of the sedimentary rocks comprising the Goedehoop-, Sandputs- and Collinskop Formations (Figure 5.9). An older zircon population observed in the Sprigg Formation of ~1.8 to 2.2Ga indicates that the Sprigg Formation most probably also sourced the sedimentary rocks deposited on the passive margin of the Kaapvaal Craton. Therefore it is believed that the Sprigg Formation was deposited in basins formed due to thermal relaxation after the final collisional event at ~1113Ma and sourced rocks from both the Kheis- and Kakamas Terranes.

5.4.11 OTHER CONSIDERATIONS

It is clear from the available aeromagnetic data that the structural grain in the Richtersveld- and Kakamas Terranes is noticeably different from that of the Kheis Terrane to the east and the Kgalagadi terrane to the north (Figures 5.3, 5.4 and 5.11). The tectonic history of this part of South Africa, Botswana and Namibia is thus based on the presence of three separate and easily distinguishable cratonic blocks or fragments affected by a complex interplay of plate
tectonic processes. These are the Kheis Terrane with the Kaapvaal Craton to its east, the Kgalagadi Terrane to its northwest and the Richtersveld and Kakamas Terranes to the west of the Namaqua Front.

5.5 PLATE TECTONIC MODEL

5.5.1 KEIS PASSIVE MARGIN BASIN AND PROVENANCE

As previously noted a detailed model for the deposition of the mainly sedimentary rocks of the Olifantshoek- and Groblershoop Groups of the Keis Supergroup since the initiation of rifting on the western margin of the Kaapvaal Craton at ~1927Ma is presented in Chapter 2. According to this model the Volop Formation of the Olifantshoek Group was deposited in a rifted basin margin that became a passive continental margin during the deposition of the Groblershoop Group.

An important question that needs to be addressed is the origin of the large volume of siliciclastics incorporated into the Keis Supergroup. The possible provenance terranes that may have acted as source for the sedimentary rocks of the Keis Supergroup was discussed in detail in Chapter 3. From this it was clear that the rocks of the Kaapvaal Craton were not a likely source area. Instead preliminary results indicated a significant amount of these metasedimentary rocks may have been derived from the 2.1Ga to 1.8Ga old Magondi Belt, to the northwest of Zimbabwe (Figures 5.5, 5.6 and 5.12). This is also supported by the presence of southwesterly directed paleocurrents in the Volop Formation of the Olifantshoek Supergroup.

Therefore other geological terranes to the northeast of the Kheis Terrane and in the vicinity of the Magondi Belt may also have acted as provenance terranes to the metasedimentary rocks of the Keis Supergroup.

The Zimbabwe craton is made up of three major time-geological units with ages of 3.5Ga, 3Ga to 2.8Ga and ~2.7Ga (Figure 5.6B). The Limpopo Belt also contains three major age populations of 3.2Ga to 3.1Ga, 2.65Ga to 2.5Ga and ~2.0Ga. These ages correspond very well with the main detrital zircon populations of the Keis Supergroup of 2.85Ga to ~2.6Ga and 2.15Ga to 1.8Ga, for which there is no real source areas of sufficient size and extent on
the Kaapvaal Craton. Therefore it is very likely that rifting on the western margin of the Kaapvaal Craton occurred at the same time as uplift and erosion of the Zimbabwe Craton and its surrounding mobile belts during the ~2.0Ga Limpopo and 2.0Ga to 1.8Ga Magondi orogeny (Figure 5.11).

**Figure 5.11:** The metasedimentary rocks of the Keis Supergroup were mainly derived from the Magondi- and Limpopo Belts as well as the Zimbabwe Craton with some input from the Kaapvaal Craton while those of the Korannaland Group mainly come from the Richtersveld- and Kakamas Terranes in the west (Modified after De Beers, 1998).

### 5.5.2 COLLISIONAL TECTONICS

The available age data, current spatial configuration of the relevant geological units and the geophysical data led to the differentiation between two possible collisional events that occurred after deposition of the Keis Supergroup along a passive margin on the western side of the Kaapvaal Craton.
Evidence for the first collisional event is provided by the northward extension of the Areachap Group along the Kalahari line with the Kgalagadi Terrane to its west and the Kheis Terrane to the east (Figure 5.3). To the south, the Areachap Group and the sedimentary rocks of the Keis Supergroup and older rocks of the Transvaal Supergroup, is obliquely terminated along a zone of strike-slip faults and shear zones including the Brakbosch fault and Triooolsan- and Boven Rugzeer shear zones (Figures 5.2 and 5.8).

The second collisional event is identified on the basis of the available zircon and $^{40}\text{Ar}/^{39}\text{Ar}$ age data which firstly indicates that the rifting event preserved by the sedimentary rocks and bimodal volcanics of the Koras Group predates the closure of an oceanic basin at $\sim$1113Ma. Secondly, the data show that the intrusion of granitic plutons of the Kokerberg gneiss was initiated at $\sim$1166Ma due to westward subduction as a result of the closure of the ocean between the Kaapvaal-Zimbabwe-Kgalagadi craton and the Richtersveld-Namaqua fragment at $\sim$1113Ma. Zircon provenance data indicate that the sedimentary rocks of the Korannaland Group were seperated from the Kaapvaal-Zimbabwe-Kgalagadi craton at time of their deposition by either an arc environment or an oceanic basin. In addition the regional deformational pattern in the area, i.e. the large scale folding of the sedimentary rocks of the Keis Supergroup and the Areachap Group along large scale fold such as the Orange River synclines, Gariep anticline and syncline and Karos anticline, suggests that the second collisional event had to occur oblique to the rifted margin of the Kaapvaal-Zimbabwe-Kgalagadi craton and resulted in a sinistral deformational pattern.

The angular unconformity at the base of the Wilgenhoutsdrif Group is probably the result of the development of a fore bulge as an island arc (represented by the amphibolites of the Jannelsepan Formation of the Areachap Group) started to collide with the Kaapvaal Craton (Figure 5.12A) at $\sim$1290Ma. During this time the Grootdrink- and the Zonderhuis Formations of the Wilgenhoutsdrif Group were deposited in a back-arc environment. Detritus were derived from erosion of the passive margin sediments of the Skurweberg Formation. As the island arc approached the Kaapvaal Craton the back arc-basin deepened while subduction was directed to the east (Figure 5.12B) resulting in the intrusion of the precursor to the Kalkwerf granite gneiss into the passive margin sedimentary rocks of the Groblershoop Formation.
The exact timing of the accretion of the island arc onto the western margin of the Kaapvaal Craton (Figure 5.12B) remains uncertain, but the unpublished ages of Moen on the deformed Leerkrans lava and Kalkwerf granite gneiss suggest that it must have been after 1290Ma (evidence for this is the deformation of the Kalkwerf granitic gneiss with the sedimentary rocks of the Groblershoop Group) but before 1172Ma (age of the Koras Group, Gutzmer et al., 2000). This arc-continent collision resulted in some deformation of the sedimentary rocks of the Keis Supergroup, but the most significant deformation occurred as the Kaapvaal Craton and its accreted island arc collided with the Kgalagadi terrane to its west (Figure 5.12C). The timing of continent-continent collision is bracketed by the age of the Leerkrans lava and Kalkwerf granite gneiss and the extrusion of the porphyries of the Swartkopsleegte Formation of the Koras Group to between 1290Ma and 1172Ma (Figure 5.12C).

The next stage in the tectonic evolution is the rifting of the amalgamated Kaapvaal-Zimbabwe-Kgalagadi craton along the Koras-Sinclair-Ghanzi rift that commenced immediately prior to the extrusion of the quartz porphyries of the Swartkopsleegte Formation at ~1172Ma with the deposition of the conglomerates of the Christiana Formation of the Koras Group (Figure 5.12D).

Soon after rifting and deposition of the Koras Group, a different cratonic fragment (the Richtersveld cratonic fragment) approached from the west at an oblique angle to the rifted margin of the Kaapvaal-Zimbabwe-Kgalagadi craton (Figure 5.12E). Subduction was to the west and lead to the intrusion of the Kokerberg granitic gneisses at ~1166Ma to form a continental arc on the Richtersveld cratonic fragment that is represented by the Kakamas Terrane. Erosion of the older Richtersveld volcanic rocks and granites as well as the newly intruded granites of the Kokerberg gneiss resulted in the deposition of the sedimentary rocks of the more proximal Goedehoop- and Sandputs Formations and the more distal Collinskop Formation of the Korannaland Group as well as the Sprigg Formation of the Areachap Group.

Just prior to the closure of the ocean between the Richtersveld cratonic fragment and the Kaapvaal-Zimbabwe-Kgalagadi craton, the intrusion of the Keimoes Suite granitoids began (Figure 5.12F). Where the granitoids intruded into the sedimentary rocks of the Korannaland Group it resulted in the formation of metamorphic aureoles of kinzigites i.e. the Collinskop Formation of the Korannaland Group and the Rateldraai Formation of the Areachap Group.
Figure 5.12: Diagram of the plate tectonic model explaining the deformation of the Keis Supergroup into the Kheis Terrane and the formation of the eastern part of the Namaqua Complex of the NNMP. Note the plan view indicating the position of the diagrams in a plate tectonic environment. See text for full explanation of each phase. A) Development of the Areachap arc to the west of the Kaapvaal Craton. B) Accretion of the volcanic island arc onto the western margin of the Kaapvaal Craton. C) Continent-continent collision between the Kaapvaal Craton and the Kgalagadi Terrane resulting in the formation of the so-called Kheis Terrane sometime between 1290Ma and 1172Ma. D) Rifting of the amalgamated Kaapvaal-Zimbabwe-Kgalagadi craton along the Koras-Sinclair-Ghanzi rift.
Figure 5.12 (continued): E) Approach of Richtersveld cratonic fragment at an oblique angle to the rifted margin and the intrusion of the Kokerberg granitic gneiss at 1166Ma. F) Intrusion of the Keimoes Suite granitoids just prior to the closure of the ocean between the Kaapvaal-Zimbabwe-Kgalagadi Craton and the Richtersveld cratonic fragment at 1113Ma. G) Present day exposures as a result of deformation due to an oblique continent-continent collision after ~1113Ma

The closure of the ocean between the Kaapvaal-Zimbabwe-Kgalagadi craton and the Richtersveld cratonic fragment occurred sometime between ~1113Ma and 1087Ma when the Friersdale charnockites intruded along a zone of crustal instability close to the suture zone between the two amalgamated continental blocks (Figure 5.12G). The event resulting in the formation of the combined the Kaapvaal-Zimbabwe-Kgalagadi-Richtersveld craton was an oblique collision and resulted in the large-scale deformational patterns that can be observed in the Kheis- and Kakamas Terranes and the eastern margin of the Richtersveld cratonic fragment (Figure 5.12F and G). This continent-continent collision resulted in the folding and deformation of the Jannelsepan island arc and the Keis Supergroup along the large-scale folds of the Orange River-, Gariep and Karos anticlines and synclines. This continent-continent
collision predates the formation of the transcurrent Brakbosch-, Dagbreek-, Blauwboschpan and Dabep faults and the Neusspruit-, Strausheim-, Boven Rugzeer- and Triooolspan shears that all seem to be expressions at different crustal levels of the same listric horse-tail fault system.

5.6 CONCLUSIONS

The tectonic history of the western margin of the Kaapvaal Craton since the extrusion of the 2.22Ga Ongeluk Andesite Formation (Armstrong, 1987 and Cornell et al., 1996) up to the formation of the NNMP during the Grenvillian assembly of Rodinia is characterised by the presence of at least two separate Wilson cycles. These Wilson cycles differ from each other in their duration, with the first lasting more than 600Ma while the second only took 60Ma to complete.

The first cycle was initiated by rifting on the Kaapvaal Craton at ~1927Ma and was followed by the accretion of the volcanic island arc (represented by the amphibolites of the Jannelsepan Formation) onto the western passive margin of the Kaapvaal Craton. This cycle reached its end when the Kaapvaal-Zimbabwe Craton collided with the Kgalagadi terrane sometime between ~1290 and ~1172Ma resulting in the deformation of the Kheis Terrane as a result of east-west directed compressional forces. Therefore it is proposed that the deformation of the Kheis Terrane occurred during the late Mesoproterozoic instead of at ~1906Ma and ~1750Ma as suggested by Cornell et al. (1998).

The start of the second Wilson cycle was signalled by rifting along the Koras-Sinclair-Ghanzi rift on the Kaapvaal-Zimbabwe-Kgalagadi craton just before ~1172Ma. It was soon followed by subduction underneath the Richtersveld cratonic fragment at 1166Ma. The closure of the oceanic basin between the Kaapvaal-Zimbabwe-Kgalagadi craton and the Richtersveld cratonic fragment occurred about 50Ma later (~1113Ma) and this oblique collision resulted in the large-scale fold represented by the Orange River-, Gariep- and Karos anticlines and synclines.

The extension of the Areachap Group towards the north and into Botswana along the Kalahari line opens up new exploration prospects for Copperton-type massive sulphide deposits underneath the Kalahari sand of northern South Africa.
The termination of the rocks of the Transvaal- and Keis Supergroups and Areachap Group to the east of Marydale is probably the result of two independent processes. The first was oblique termination of the geology as a result of rifting along the Koras-Sinclair-Ghanzi rift at ~1172Ma which was followed by deformation due to the oblique collision between the Kaapvaal-Zimbabwe-Kgalagadi craton with the Richtersveld cratonic fragment at ~1113Ma (Figure 5.12 F and G).
5.7 REFERENCES


Corner, B, Durrheim, R.J. and Nicolaysen, L.O. (1990): Relationship between the Vredefort Structure and Witwatersrand basin within the tectonic framework of the Kaapvaal Craton as interpreted from regional gravity and aeromagnetic data. *Tectonophysics*, **171**, 49-61


