2 Regional Geological Setting

2.1 Introduction

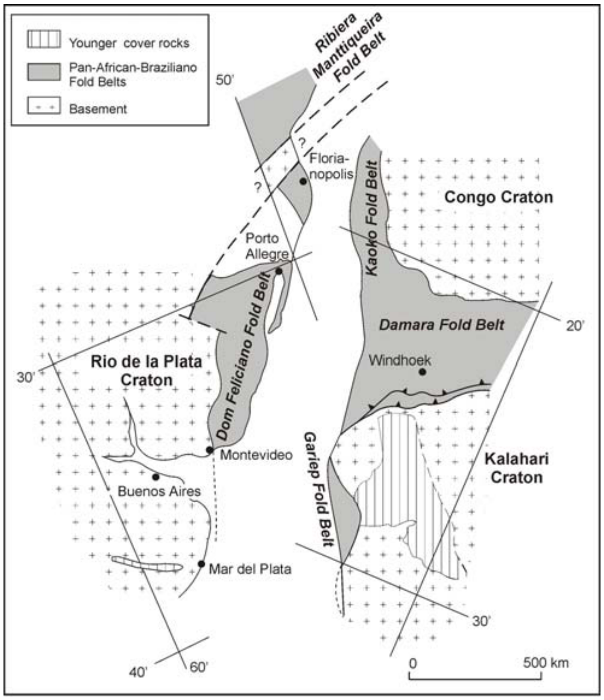

Fig. 5: Regional location of the Neoproterozoic Gariep Belt in south-western Africa, illustrated within the framework of the Pan-African-Brasiliano fold belts (modified after Porada 1989, Tankard et al. 1982, Trompette 1994).

A network of Late Proterozoic to Early Palaeozoic orogenic belts is the prominent regional tectonic fabric of western Gondwana (Frimmel et al. 1996).

In south-western Africa, the Damara Belt separates the Congo and Angola Cratons from the Kalahari Craton, whereas the Gariep Belt forms the southern coastal branch of the orogen and extends also northward into the Kaoko Belt (Fig. 5).

The Gariep and Kaoko Belts originated as a Late Proterozoic suture between the South American Craton and the cratons of southern Africa. The rifting between the Kalahari and Rio de la Plata plates was initiated around 781 Ma ago and lasted some 40 Ma. It was accompanied by the formation of the Adamastor Ocean and oceanic crust (e.g. Frimmel et al. 1996, Frimmel & Frank 1996). The inversion from extension to compression led to a successive closure of first the northern Adamastor ocean (Kaoko Belt), followed by the Khomas sea (intracontinental Damara Belt), and finally the southern Adamastor ocean (Gariep Belt) (Fig. 6). Continent-continent collision and thrusting of the internal onto external zones of the tectonic Gariep Belt culminated between 547 – 543 Ma (Frimmel & Frank 1998).

The Late Proterozoic Pan-African orogeny was followed by a long period of tectonic stability. The break-up of Gondwana and thus, rifting in the South Atlantic was initiated in the Permo-Triassic and continued until the end of the Early Cretaceous. Rifting overlapped with the Cape Orogeny and the development of the Karoo foreland (Gilchrist et al. 1994). Igneous activity was widespread and spanned the period of rifting and into the Tertiary. A series of intrusive complexes in north-western Namibia is related to the Etendeka lavas, which were extruded during rifting. Suites of kimberlite pipes and alkaline plugs also intruded from the Late Cretaceous to the Early Tertiary. Terrestrial sedimentation was extensive in the continental interior and along the coast, represented by the Kalahari and Namib sediments respectively (Gilchrist et al. 1994).

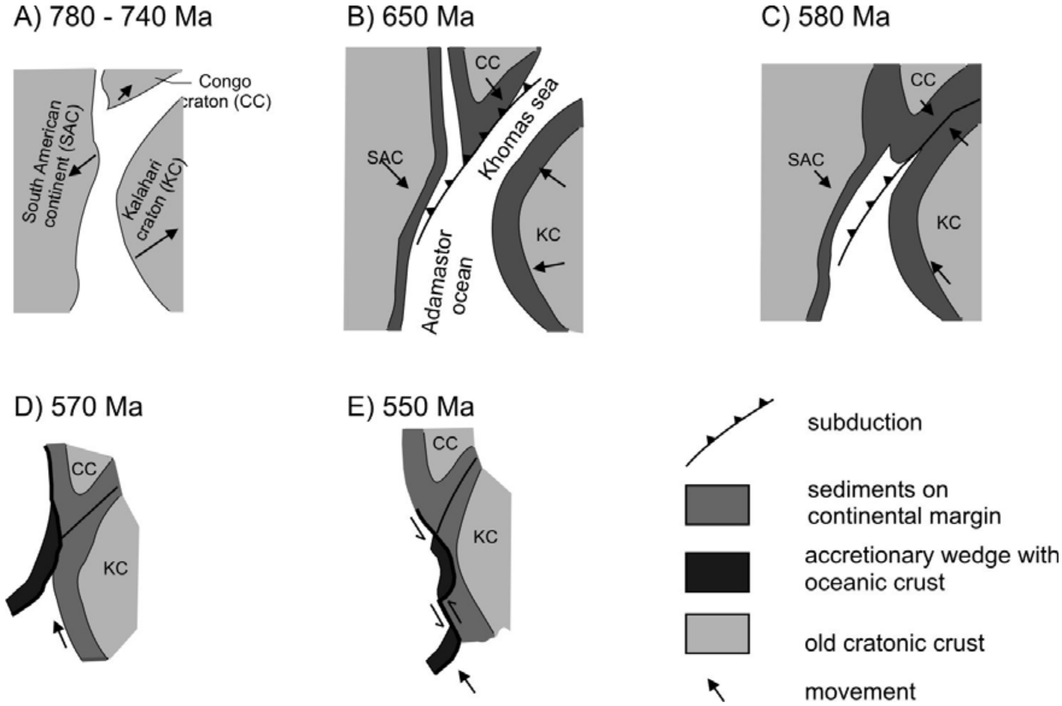

Fig. 6: Schematic diagram showing successive stages of the tectonic evolution that led to the development of the Gariep Belt: (A) rifting, followed by the opening of the Adamastor ocean and the Khomas sea; (B) closure of the Khomas sea and north-westward subduction; (C) continent-continent collision in the Damara Belt and formation of an accretionary wedge further southwest (Chameis Complex); (D) obduction of accretionary wedge; and (E) oblique collision in the Gariep Belt (Frimmel & Frank 1998).

2.2 Gariep Belt

2.2.1 Geological Framework

The Late Proterozoic Gariep Belt is regarded as the southern extension of the Damara orogenic front of central and northern Namibia (Davies & Coward 1982, Reid et al. 1991, Stanistreet et al. 1991, Gresse 1994, Frimmel 2000b, Jasper et al. 2000).

It is subdivided into an eastern para-autochthonous zone, the so-called Port Nolloth Zone (PNZ), which evolved from an intracontinental rift to a passive continental margin on the western edge of the Kalahari Craton (Jasper et al. 2000), and a western allochthonous zone, the Marmora Terrane (Fig. 7).

The Skorpion deposit is situated within the Port Nolloth Zone, which comprises a variety of siliciclastic and chemical sediments and bi-modal volcanic rocks, which have been strongly folded, faulted and overprinted by lower amphibolite facies metamorphism (e.g. Frimmel et al. 1995).

The rocks of the Port Nolloth zone contain stratiform Zn-Pb-Cu-Ag-(± Ba)-sulphide mineralisation in syn-rift sediments and felsic metavolcanic rocks, e.g. Rosh Pinah Pb-Zn mine, and Skorpion Zn mine (e.g. van Vuuren 1986, Alchin & Moore 2005). These base metal mineralisations were formed in an extensional environment during a phase of increased volcanogenic-hydrothermal activity between 740 and 754 Ma. The age of volcanogenic-hydrothermal activity is supported by isotopic SHRIMP age of zircons from a Skorpion meta-rhyolitic flow that gave 751.9 ± 5.5 Ma (Borg & Armstrong 2002).

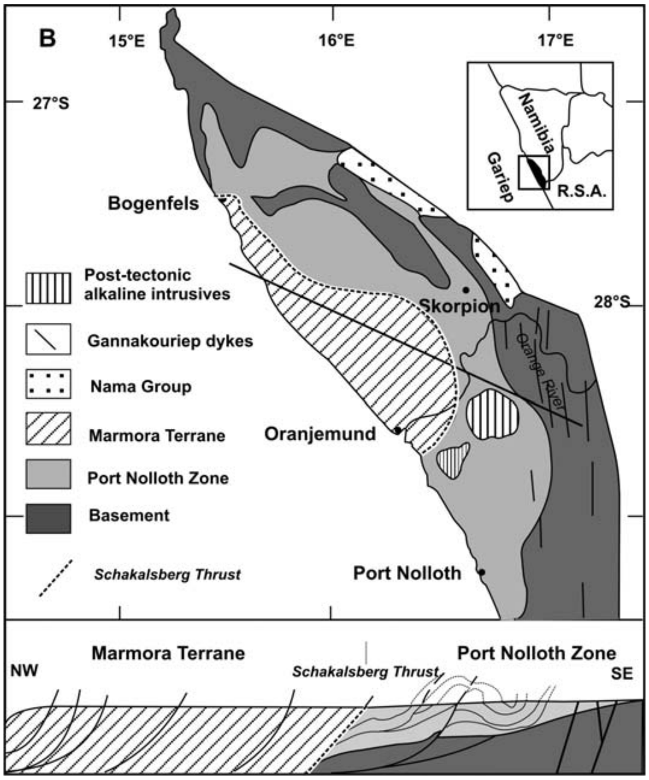

Fig. 7: Distribution of tectono-stratigraphic units of the Gariep Belt with a NW-SE cross-section. After Frimmel & Frank (1998).

The extensional phase ended with the deposition of the glaciogenic Numees Formation between 590 and 564 Ma, which correlates with the Varangerian glacial episode (Jasper et al. 2000).

After the deposition of the Gariep group ceased, the closure of the Adamastor Ocean resulted in continental collision, and thus in deformation and metamorphism. The compressional history ceased at about 500 Ma, followed by the deposition of predominantly siliciclastic Nama Group sediments into peripheral foreland basins (Stanisstreet et al. 1991, Germs & Gresse 1991, Jasper et al. 2000).

A number of pre-, syn-, and post-tectonic intrusions occur throughout the Gariep Belt. The 100 km wide N- to NE-trending, mafic /ultramafic Gannakouriep dyke and sill swarm intruded both Palaeo-Proterozoic basement rocks, which are part of the 1.0 Ga old Namaqua-Natal Metamorphic Belt, and late Proterozoic Gariep cover rocks. The dyke swarm has been dated at 717 ± 11 Ma (Reid et al. 1991), marking the final stretching of the continental crust prior to the opening of the Adamastor Ocean in the West (Jasper et al. 2000).

Several post-orogenic intrusive bodies of alkali granite and syenite cross-cut the main tectonic fabrics and lithological contacts in the central part of the Gariep Belt. They occur along the so-called Kuboos-Bremen line, which strikes SW-NE, south of the Orange River. The Kuboos pluton is clearly younger (507 ± 6 Ma) than the main phase of deformation in the Gariep Belt (545 ± 2 Ma). While the earlier 545 Ma tectonothermal event is explained by the closure of the Adamastor Ocean between the Kalahari and the Rio de la Plata cratons, the younger 500 Ma tectonic pulse is linked to subduction beneath the western margin of Gondwana (Frimmel 2000b).

2.2.2 Tectonic Evolution

The tectonic evolution of the Gariep Belt is subdivided into an earlier sinistral transpressive phase (Davies & Coward 1982, Gresse 1994) with south-southeastwards directed thrusting and a later easterly to northeasterly verging deformation that affected the western part of the Gariep Belt (Gresse 1994). The earlier Gariepian deformation phase (G_1) is dated at between 542±4 Ma and 546±10 Ma by Onstott et al. (1986) and Reid et al. (1991); using a metamorphic overprint age on earlier Gannakouriep dykes that predates the deposition of the Gariep Sequence. The later (G_2) event is correlated to the late Pan African/Brasiliano event (at approx. 500 Ma) that affected the Nama foreland deposits up to 50 km farther towards the east at the Neint Nababeep Plateau (Gresse 1994).

The structural style of the G_1 event is dominated by thrust structures that strike north-northeast to south-southwest in the central and northern part of the belt and northeast-southwest in the southern part of the belt. The most prominent structure is the Schakalsberg thrust that represents a major terrane boundary (Fig. 7), juxtaposing the oceanic allochthonous Marmora Terrane (towards the west) with the continental para-autochthonous Port Nolloth Zone towards the east (Davies & Coward 1982, Hartnady & von Veh 1990, Frimmel 2000b, Jasper et al. 2000).

The initial stages of the southeastwards directed thrusting imparted shallow north-north-westerly dipping planar fabrics in high-strain zones. Upper greenschist facies metamorphic assemblages define the fabrics. The transport direction of the thrusting is defined by the axial orientation of sheath folds and the strained long axes of pre-tectonic shape fabrics. The early formed thrust structures and associated planar fabrics became subsequently deformed during the later stages of the same deformation event. This event produced fold structures that are defined by the foliation/bedding and thrusts that affected both the Gariep Sequence and underlying basement rocks. The folds verge towards the east and southeast, and in the Rosh Pinah area, towards the west. The variable vergence of the folds is explained in terms of a progressive shear model where the initial folds form with axial orientations at large angles to the transport direction and that are then rotated by subsequent shear in either a clockwise or anti-clockwise sense, thus producing folds with an easterly vergence as well as folds with a westerly vergence. Thus, it is important to note that apparent westerly- or easterly-directed thrusting in west-east cross sections commonly has a major south-southeasterly directed thrust component attached to it.

Effects of the later G_2 event are restricted to the presence of slickensides from an area south of Port Nolloth (Gresse 1994), the presence of northeast- verging folds from the northern part of the belt (Davies & Coward 1982), the folding of the Nama sediments farther towards the west and perhaps a general tightening of earlier folds. No large-scale thrusting was observed.

2.2.3 Stratigraphy and Lithology

Regional stratigraphic correlations of the Late Proterozoic rock sequence within the Gariep Belt have been subject of considerable debate and several stratigraphic schemes have been proposed by different authors, e.g. SACS (1980), von Veh (1993), Frimmel (2000), and Alchin et al. (2005) (Fig. 8). This unsatisfactory situation is a result of poor, isolated outcrops, the complexly deformed rock sequence, rapid lateral facies changes of both metasedimentary and metavolcanic rock types.

Due to the on-going controversial discussion about the stratigraphical positions of the different lithotypes, this metallogenic study uses lithological terms rather than stratigraphical terms. The stratigraphical terms are also avoided, since this study is based mainly on drill core data that did not allow a reliable stratigraphic classification. However, the different stratigraphic units are summarised below in order to give a general overview of the occurrence of the different Late Proterozoic lithotypes and their possible stratigraphic positions.

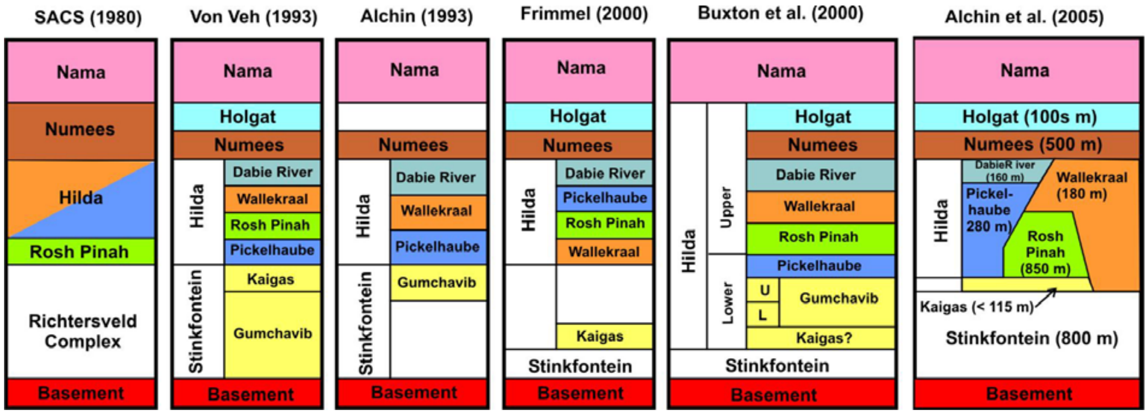

Fig. 8: Schematic stratigraphic schemes for the Late Proterozoic Gariep sequence.

The basement, on which the rocks of the Port Nolloth Zone rest, is part of the some 1.0 Ga old Namaqua-Natal Metamorphic Belt and includes 1730 - 1900 Ma Vioolsdrift Suite granites and 2000 Ma volcanics of the Haib Subgroup (von Veh 1993, Alchin 1993).
The Late Proterozoic Gariep cover rocks, which lie in most places tectonically on the Palaeoproterozoic basement rocks, approximately correspond to sequences assigned regionally to Stinkfontein and Hilda Subgroups, including Gumchavib, Pickelhaube, and Rosh Pinah Formations (Fig. 8). However, no reliable correlation is presently possible because these formations are characterised by rapid lateral facies changes and multiple deformation. Genetically, the Gariep sequence is related to progressive opening of a failed intracratonic rift graben in the east (Rosh Pinah/Skorpion Graben) that was separated by a basement horst from a half graben to the west (Fig. 9), which developed into the Adamastor ocean (Alchin et al. 2005).

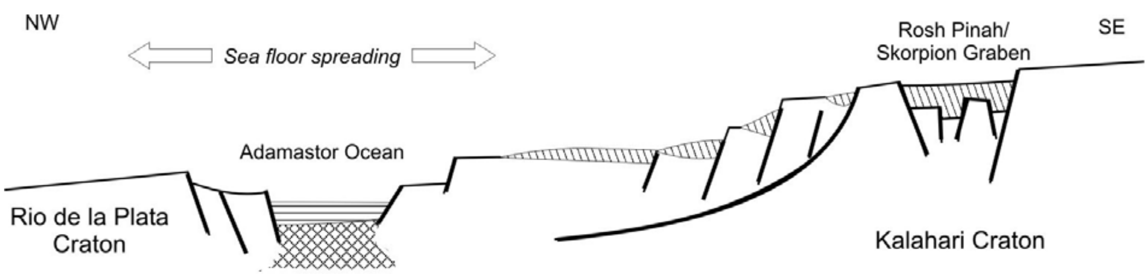

Fig. 9: Position of the Rosh Pinah Graben in thinning and breaking Mesoproterozoic continental crust around 700 Ma, after Alchin et al. (2005).

The Stinkfontein Subgroup is exposed mainly in South Africa along the southern and south-eastern front of the Gariep Belt (Frimmel 2000b). Northward thinning of the subgroup is believed to be due to stratigraphic onlap. The Stinkfontein Subgroup contains siliciclastic sediments in its lower part, containing mainly quartzarenites, feldspatic arenites and conglomerates of the so-called Lekkersing Formation. The Lekkersing Formation is conformable overlain by the Vredefontein Formation, which includes feldspatic arenites and minor metamorphosed felsic volcanic rocks (Frimmel 2000b, Alchin et al. 2005).
The Hilda Subgroup, containing a mixed sequence of shallow marine siliciclastic and carbonate sedimentary rocks with intercalated volcanic rocks, rests either unconformable on

Palaeo-Proterozoic basement rocks or para-conformable, on the Stinkfontein Subgroup or Kaigas Formation, though recently published stratigraphic schemes by Alchin et al. (2005) consider at least the Kaigas Formation to be the lowermost part of the Hilda Subgroup.

The Lower Hilda Subgroup consists mainly of fine- to medium-grained meta-arkoses, -subarkoses, - and sandstones (von Veh 1993, Alchin 1993) of the so-called Gumchavib Formation (Fig. 10). Some distal fine-grained deposits of this formation are apparently exposed some 15 km SSE of the Skorpion deposit (Buxton et al. 2000). Debris flow deposits of the Kaigas Formation might represent a basin marginal deposit, and thus a proximal Gumchavib facies (Buxton et al. 2000). However, the existence of the Gumchavib Formation is still controversially discussed as chemostratigraphic data (Fölling et al. 1998) indicate that the Gumchavib Formation might be a facies equivalent of the Pickelhaube Formation in the Hilda Subgroup (Frimmel 2000b).

Extensive carbonate rocks appeared the first time at the base of the Upper Hilda Subgroup (von Veh 1993, Alchin 1993, and Buxton et al. 2000), though according to Frimmel (2000b), these carbonates form the uppermost part of the Upper Hilda Subgroup. The stratigraphic scheme published by Frimmel (2000b) is based on an age dating of marbles from the lower Pickelhaube Formation (Fig. 10), which yielded $^{207}Pb/^{206}Pb$ isochron ages of 728 ± 32 and 545 ± 13 Ma for the carbonate and residue fractions, respectively. The former is interpreted as dating early diagenesis, whereas the latter refers to the metamorphic event. Thus, ages obtained from the marble by Frimmel (2000b) indicate that it is younger than the felsic volcanic metavolcanic rocks from the Rosh Pinah Formation (see below). However, the absolute isochron ages of the marbles from the Pickelhaube Formation have been re-interpreted recently by Frimmel & Lane (2005) as representing a more distal facies that was deposited contemporaneously with the Rosh Pinah Formation. In fact, the bulk of the Rosh Pinah Formation is correlated with the lower Pickelhaube Formation in those areas that were not affected by rift volcanism (Alchin et al. 2005).

Regionally, the carbonates of the Pickelhaube Formation consist of extensive calcitic marble, which rest either directly upon basement or gradationally upon the siliciclastic rocks of the Gumchavib Formation (Alchin 1993, Buxton et al. 2000). It is agreed that the marble of the Pickelhaube Formation represents an extensive shallow water carbonate sequence probably developed during shallow water sub-tropical conditions. The marble that forms an antiform on the eastern side of the Skorpion deposit, most likely belongs to the Pickelhaube Formation.

Bi-modal metavolcanic rocks and siliciclastic metasediments of the Rosh Pinah Formation (Fig. 10) rest on top of the carbonates of the Pickelhaube Formation according to von Veh (1993) and Buxton et al. (2000). Conversely, the comparison of isotopic ages of metarhyolite of the Rosh Pinah Formation (Borg & Armstrong 2002) and marble of the Pickelhaube Formation (Frimmel 2000) indicates that the Rosh Pinah Formation is older and thus might underlie the Pickelhaube Formation. However, the latest stratigraphic research shows that the Rosh Pinah Formation has been laid down around the same time as the Pickelhaube Formation (Alchin et al. 2005), which also finds support in a Pb-Pb carbonate age of 728 +-32 Ma obtained on the latter (Frimmel & Lane (2005).

The lower part of Rosh Pinah Formation consists mainly of metarhyolite, rhyolitic agglomerates and ignimbrites. The upper part of the Rosh Pinah Formation consists predominantly of meta-tuffites, impure, marly metacalcarenite and largely dolomitic marble beds. Additionally,

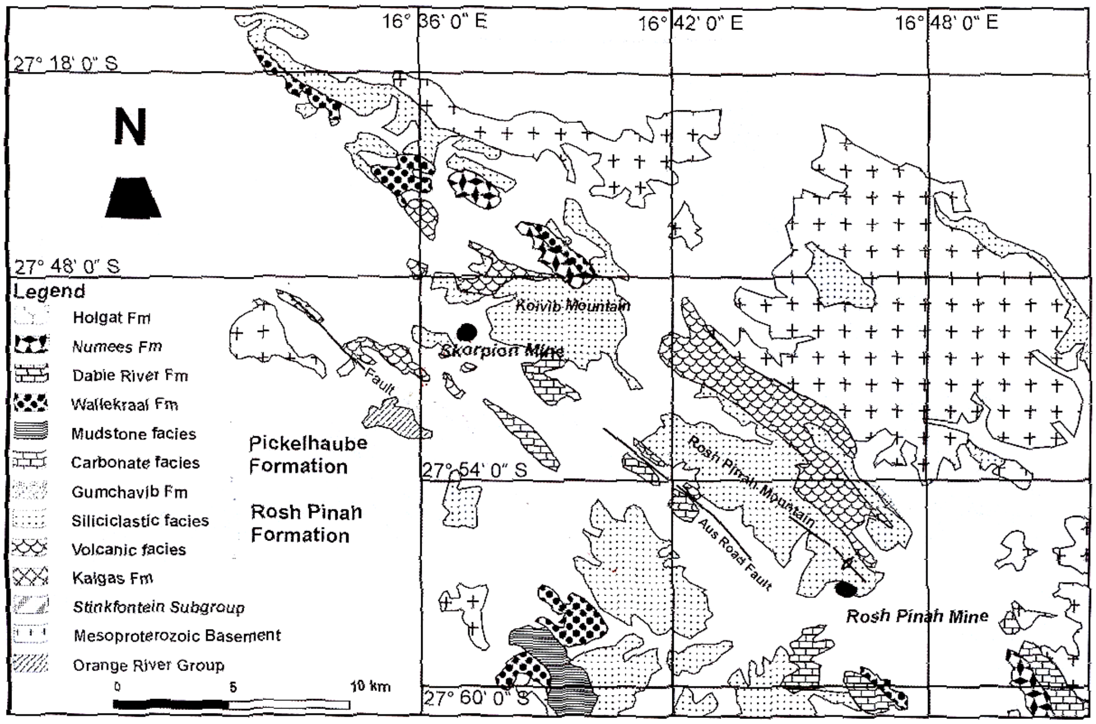

Fig. 10: Simplified geological map indicating the stratigraphy of the Rosh Pinah/Skorpion area. From Alchin & Moore (2005).

two black shale horizons occur in the middle and upperparts of the sequence, respectively, with the lower one being rich in sulphides. In its lower part, the Rosh Pinah Formation hosts the stratiform Pb-Zn-Cu sulphide bodies of the Rosh Pinah mine (Alchin et al. 2005). The sulphide precursor of the non-sulphide Skorpion deposit and remnants thereof are most likely also hosted by rocks of the Rosh Pinah Formation as they consist mainly of felsic metavolcanic rocks, e.g. metarhyolites and meta-tuffites, which make up a significant part of the Rosh Pinah Formation.

The Rosh Pinah Formation is overlain by the so-called Wallekraal Formation (von Veh 1993, Buxton et al. 2000), though it is also regarded as a lateral and time equivalent of the Rosh Pinah Formation (Alchin 1993, Alchin et al. 2005). The siliciclastic-dominated Wallekraal Formation consists generally of well sorted quartz-pebble conglomerate and feldspathic meta-arenite grading into metapelites in upward-fining sequences and is exposed some 25 km SSE of the Skorpion deposit (Buxton et al. 2000). The siliciclastic metasediments, which host a significant portion of the Skorpion non-sulphide ore body, might be equivalents of either Wallekraal Formation or the metasedimentary part of the Rosh Pinah Formation described above.

There is a general agreement that massive dolomitic limestone and/or dolomite, assigned to the Dabie River Formation, forms the uppermost part of the Hilda Subgroup. The rocks of the Dabie River Formation are only exposed close to the Orange River and are distinguished from the other carbonates in the subgroup by the presence of stromatolites, pisolites and oolites (Frimmel 2000b, Alchin et al. 2005).

The Hilda Subgroup is overlain by a diamictite with banded iron formation. The so-called Numees Formation is largely exposed about 30 km SE and SSE of the Skorpion deposit (Buxton et al. 2000). Metacalcturbites and metasiliciclastic rocks of the Holgat Formation, which are exposed about 15 km SW of the Skorpion deposit (Buxton et al. 2000), rest on top of the Numees Formation.