

**THE METALLOGENY OF THE UPINGTON AND KENHARDT
AREA, NORTHERN CAPE**

BY

ROBERT BOELEMA

**THESIS SUBMITTED IN PARTIAL FULFILMENT
OF THE REQUIREMENTS FOR THE DEGREE OF
MASTER OF SCIENCE (ECONOMIC GEOLOGY)
OF RHODES UNIVERSITY,
GRAHAMSTOWN**

1994

Extract from "The Hitch Hiker's Guide to the Galaxy"

"Space," it says, "is big. Really big. You just won't believe how vastly hugely mindbogglingly big it is. I mean you may think it's a long way down the road to the chemist, but that's peanuts to space."

From Douglas Adams, The Hitch Hiker's Guide to the Galaxy.

ABSTRACT

In the Upington region, there are three major tectonic crustal provinces; namely the Kaapvaal Craton, Kheis and Namaqua tectonic provinces. The Eburnian-aged (early Proterozoic) Kheis Province developed along the western flank of the Archaean Kaapvaal Craton while the Kibaran-aged (middle Proterozoic) Namaqua Metamorphic Province, superimposed on the Eburnian-aged basement, developed to the east of the Kheis Province. The Namaqua Metamorphic Province is divided into the Gordonia and Bushmanland Subprovinces, the former being further subdivided into various tectonostratigraphic terranes. These are termed, from west to east, the Kakamas, Areachap, and Upington Terranes. The Upington Terrane includes fault bounded grabens with accompanied bimodal volcanism and sedimentation of the Wilgenhoutsdrif and Koras Groups. The Areachap Terrane consists predominantly of amphibolites generated in an island arc environment while the Kakamas Terrane is characterised by volcano-sedimentary sequences which have been extensively intruded by syn to late-tectonic predominantly I-type Keimoes Suite granitoids.

The main styles of mineralisation correlate well with the various tectonostratigraphic terranes. Sedimentary exhalative massive sulphide deposits are characteristic of the Bushmanland Subprovince and are thought to be associated with the deposits at Aggeneys and Putsberg to the west of the area under investigation. These deposits are considered to have been deposited in an east-west-elongated intracontinental basin. The Kakamas Terrane is typified by granite-related mineralisation. In the eastern portion of the Kakamas Terrane, Sn-W and base metal-bearing veins occur while pegmatites are developed in the western portion. These two styles of granite-related mineralisation is considered to reflect differing depths of formation due mainly to varying degrees of thrusting. The Areachap Terrane consists of volcanogenic massive sulphide deposits of the Besshi-type and is considered to have formed in a back-arc environment. In the Upington Terrane, the Wilgenhoutsdrif and Koras Groups consists essentially of minor Cu occurrences mainly disseminated within basalts and in structural trap sites. The possibility for sediment-hosted Cu deposits is not ruled out. More recent surface processes have led to uranium and gypsum deposits in pans, river beds and calcretes.

Eburnian aged tectonic setting remains enigmatic. Kibaran-aged tectonics which best fits the metallogeny of the area under investigation is considered to be of a subduction zone from west to east formed by the collision of the Bushmanland "microcontinent" against the Kaapvaal Craton. Subduction formed an island arc setting in which the massive sulphide deposits were formed in the Areachap Terrane while the Wilgenhoutsdrif Groups developed in a marginal basin. Further convergence led to collision of the two continents and underthrusting of the Bushmanland "microcontinent" which generated predominantly I-type granitoids represented by the Keimoes Suite. The level of emplacement of these granitoids is a reflection of the degree of foreland thrusting and produced shallower level Sn-W and base metal vein-type mineralisation closer to the suture zone and deeper level pegmatites further from the suture zone to the west. The final period of deformation is represented by northward lateral movement which created "pull apart" fault-bounded basins into which the Koras Group was deposited.

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Chapter One

1. INTRODUCTION.

1.1. PHYSIOGRAPHY.

The area under investigation comprises the two 1:250 000 map sheets, 2820 (Upington) and 2920 (Kenhardt). The north-south and east-west boundaries are defined by 28°00' S, 30°00' S and 20°00' E, 22°00' E respectively.

The topography of the study area generally becomes less pronounced from north to south. In the north-east, the area is characterised by flat topped, deeply incised plains of the Nama Group. The central, and largest portion of the study area is dominated by rounded inselbergs, and, further south, by more undulating topography of the Namaqualand Metamorphic Complex (NMC). In the south of the study area the Karoo Sequence predominates, forming a peneplain with numerous buttes and mesas largely ascribed to Jurassic-aged Karoo dolerite. The late Tertiary to Quaternary continental uplift and progressive aridification is dramatically represented by the Orange River gorge and the Kalahari sands which overlie the area in the northeast and form permanent dunes in the area, as well as by the widespread development of calcrete (Partridge & Maud, 1987).

The drainage of the area is dominated by the Orange River and its tributaries namely the Molopo River from the north, and the Hartbees and Sout Rivers from the south, as well as numerous periodical streams which define a well developed dendritic drainage pattern. Towards the south, in the Karoo rocks, the drainage is poor and is characterised by shallow flat-lying pans.

The average annual precipitation in the region is in the vicinity of 127 mm which leads to a semi-arid vegetation comprising various types of *euphorbia aloë*, most notably the "kokerboom" (*Aloe dichotoma*), grasses, especially on the sandy flats, and acacia trees along the banks of the river beds.

1.2. PREVIOUS WORKERS.

Although there is an abundance of literature present describing the geology in the area under investigation, only that relevant to the metallogeny will be discussed. The first activity, with respect to mining in the area, was at the Strausheim No.1 pegmatite situated on the farm N'Rougas Noord, between Keimoes and Kenhardt where beryl, tin, mica and feldspar were mined in 1900. No formal investigations were made into the economic potential of the pegmatites until the early forties when Von Backström and Visser surveyed the area around Kakamas and the Bokvasmaak Nature Reserve in the northeastern portion of the area. This work culminated in reports of the area around Kakamas by Poldervaart & Von Backström (1949) and in the Bokvasmaak Reserve by Von Backström (1963). To the west of the area, in 1937, Gevers, Partridge and Joubert produced an account of the pegmatites south of the Orange River in Namaqualand. Hugo (1969; 1986) conducted a detailed investigation of the pegmatites in the Northern Cape. The exact origin of the pegmatites is still uncertain.

Since the discovery of the Prieska Cu-Zn ore body in the early seventies, the Areachap Group has enjoyed most of the attention of exploration companies, particularly during the seventies and early eighties. Middleton (1976) first described the Prieska ore body and along with Uiterwyk & Frick (1985) concluded that it was of volcanogenic origin similar to the Kuroko deposits of Japan. Wagener & Van Skalkwyk (1986) subsequently interpreted the Prieska ore body as representing a synsedimentary Kupferschiefer-type deposit. Cornell et al. (1986; 1990) established ages for the Prieska ore body, while Voet & King (1986) interpreted the Areachap deposit to be of exhalative origin, and Humphreys (1986), Geringer et al. (1987), investigated the Bokspuits deposits and showed that these are related to the Besshi-type of volcanic exhalative. Theart (1985) reviewed the Areachap and Copperton Cu-Zn mineralisation and drew comparisons between them. In 1981, Gorton described the Kielder deposits northeast of the Prieska ore body and concluded that they were of volcanogenic exhalative origin.

Von Backström (1950) and Wheatley & De Beer (1986) described the Sn-W occurrences at Van Roois Vlei. Smithies & Pirajno (1989) described the Van Roois Vlei deposits as having been produced by a specialised granite whereas Bowles (1988), in a regional investigation of W mineralisation in the Northern Cape, including those in the Riemvasmaak and Van Roois Vlei regions, concluded that the Van Roois Vlei occurrences originated from the remobilisation of a tourmalinite layer exposed in the vicinity. The Renosterkop Sn-W deposit near Kakamas, just south of the Orange River was investigated by Saad (1987) who concluded that it was formed as an endo-granitic greisen.

Attridge (1986) (Jacomynspan Cu-Ni) and Bicker & Ralston (1986) (Lutzputs Fe-Cu-Ag) described other base metal occurrences which are present in the area.

Surficial uranium precipitates were also the target of exploration in the mid to late seventies and were investigated by Treasure (1977) and Hambleton-Jones (1986). Fockema et al. (1962) and Visser et al. (1963) briefly described various occurrences of gypsum in the area.

The stratigraphic nomenclature used in this text, some of which has not yet been approved by SACS, refers to the 1:250 000 geological map of Upington (2820) produced by the Geological Survey in 1988, as well as the 1:250 000 geological map of Kenhardt (2920), which is still in press. With regards to the legend for all colour maps, the reader is referred to the legends in the appendix E.

1.3. AIM OF THIS STUDY.

The Geological Survey of South Africa is currently engaged in an extensive metallogenic mapping programme covering parts of South Africa. All information on the mineral deposits is captured on the South African Minerals Data Base (SAMINDABA) as well as used to produce metallogenic maps on a 1:250 000 scale to supplement the 1:250 000 geological maps. The above has been accomplished with respect to the 2820 (Upington) and 2920 (Kenhardt) maps, and the purpose of this

investigation is to utilise the information acquired during field work and that within SAMINDABA to interpret the metallogeny of the region.

This has been accomplished with the aid of computerised MIPS (Maps and Image Processing System by Micro Images, Nebraska USA) Geographical Information System (GIS), where all geological information is digitised and integrated. The SAMINDABA database was imported into MIPS. Styles of mineralisation were classified according to metallogenic provinces and epochs and an assessment of the resource potential of the more important deposit styles has been made. A tectonic model of the area, using parameters defined by the styles of mineralisation, is also discussed.

Initially, in chapter two, this thesis provides a geological synopsis of views held by various authors to provide the reader with a geological background to the area under investigation. In chapter three, an attempt is made to familiarise the reader with the known mineral deposits in the study area. Chapter four is a metallogenic analysis of these mineral deposits with respect to metallogenic provinces and their possible genesis. This is followed in chapter five by a possible tectonic model and in chapter six by a general summary.

Chapter Two

2. GEOLOGICAL FRAMEWORK.

2.1. INTRODUCTION.

Three main tectonic events occurred during the Proterozoic and early Palaeozoic in Southern Africa. These are known as the Eburnian, Kibaran and Pan African episodes and are represented in Fig. 2.1. During the Eburnian (early Proterozoic) the Okwa Province, the Kheis Province and Magondi Province developed along the western flanks of the Archaean Kaapvaal and Zimbabwe cratons between 2000 and 1700 Ma. The low-grade, volcano-plutonic Richtersveld Subprovince (2000-1800 Ma) is separated from the Kheis Province by the Middle Proterozoic, high-grade Namaqua Province which consists of the Natal, Gordonia and Bushmanland Subprovinces (1700-900 Ma). These subprovinces consist of supracrustal paragneisses, orthogneisses, charnockites, granitoid plutons and pegmatite belts whose metamorphic-plutonic history reached its peak between 1200-900 Ma. The late Proterozoic to Early Palaeozoic period is represented by the Pan African Orogeny and includes the Damara (Miller, 1983), and Saldanha and Gariiep (Davies & Coward, 1982) Belts.

In the Upington region, there are three major tectonic crustal provinces formed by three different tectonic events. These comprise the Kaapvaal Craton and the Kheis and Namaqua tectonic provinces. (The Kaapvaal Craton, which falls outside the study area to the east and will not be discussed in any detail, is considered to have remained relatively stable since the Archaean). The margin of the Kaapvaal Craton is marked by the north-east-trending Langeberg fold belt (Fig. 2.2) which extends from the Marydale area, northwards into Botswana (Stowe, 1983).

Following the period of intense Kibaran-aged deformation, the stabilisation of the continental crust in this area resulted in the deposition of younger sediments which partially obscure the older, underlying Namaqua rocks. Namibian-aged (± 500 Ma) Nama Group sediments were deposited in a shallow marine and continental fluvial

environment (Geringer, 1973) in the northern and northeastern portion of the area. Also in the north, as well as extensive areas in the south, glacial, shallow marine and/or lacustrine environments dominated during Carboniferous to Permian times (from ± 350 Ma) and deposited parts the Karoo Sequence. Large areas in the northeast are covered by Quaternary aeolian sands of the Kalahari Group.

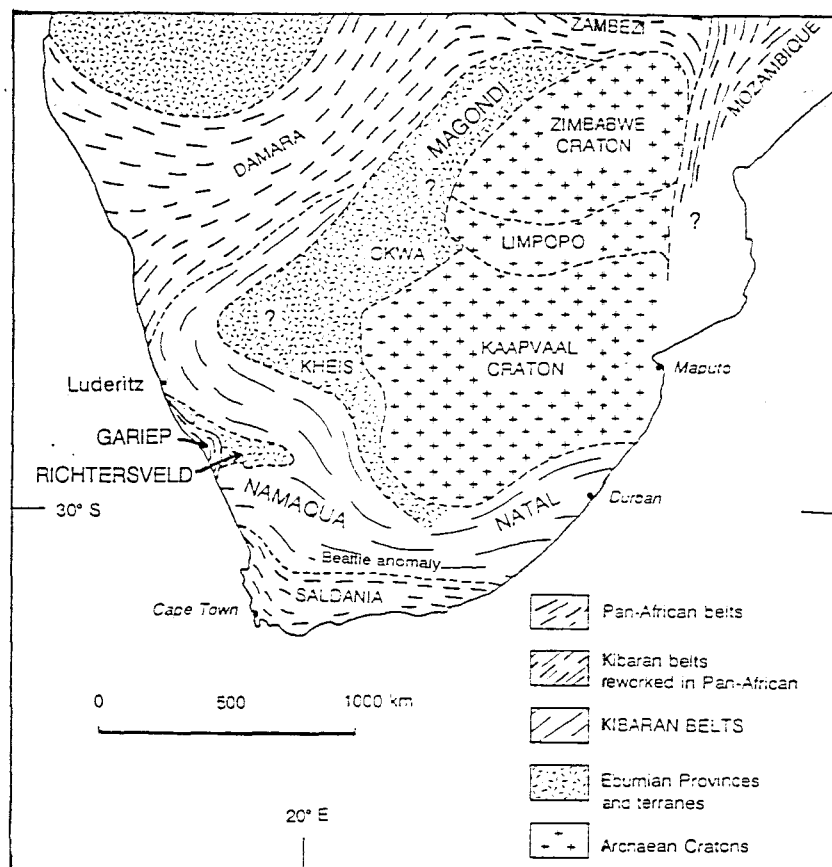


Figure 2.1: The tectonic events in Southern Africa during the Proterozoic and early Palaeozoic (after Thomas et al., 1994). According to Hartnady et al. (1985) the western boundary of the Kheis Province is defined by the geophysical discontinuity situated west of the Okwa Terrane and is further east than is shown in this figure.

2.2 THE KHEIS TECTONIC PROVINCE.

The Kheis Tectonic Province is bounded by the Brakbosch and Blaauwbospan Faults in the south-west and the Kaapvaal craton in the east (Kröner & Blignault, 1976) and

converges southwards towards Marydale where it is wedged in a graben structure between the Brakbosch and Doringberg Faults (Fig. 2.2). The Mid Proterozoic supracrustal sequences of the Kheis Province comprise low- to medium-grade quartzites, phyllites and amphibolites, deposited on the Kaapvaal continental shelf and was subjected to deformation and metamorphism during the Eburnian (1750-2000 Ma) event (Stowe, 1986).

The western boundary of the Kheis Tectonic Province has been defined by Thomas et al. (1994) as being the eastern limit of the Namaqua deformation. Thomas et al. (1994) called this the "Namaqua Front", but did not pinpoint this boundary line as Namaqua ages have been cited within the Kheis Province (Key & Rundle, 1981). The Namaqua-aged deformational overprinting effect decreases rapidly towards the east. Stowe (1986) defined this eastern boundary as the limit of penetrative Namaqua fabric and metamorphic overprint, broadly coinciding with the Brakbos Fault system. Tankard et al. (1982) referred to the Kheis Tectonic Province as the "Eastern Marginal Zone" which is similar to the Kheis domain of Vajner (1974b) and the Kheis and Matsap domains described by Botha & Grobler (1979). The western boundary of the "Eastern Marginal Zone" is defined by the Doringberg lineament (Tankard et al., 1982) which is an assemblage of *en echelon* faults (Pretorius, 1974; Vajner, 1974b; Botha et al., 1977 and Stowe, 1979). Pretorius (1974), Stowe (1979) and Tankard et al. (1982) believe the Dagbreek fault to represent the boundary in the northern part of the "Eastern Marginal Zone". Vajner (1974a,c) and Botha & Grobler (1979) defined the eastern boundary with the Kaapvaal Craton to be along the western margin of the Archaean Kaapvaal granites. Stowe (1986) defined the eastern boundary of the Kheis Province with the Kaapvaal Craton by the lower thrust zone beneath the Matsap-Groblershoop sequence, well east of the Langeberg Range.

The Kheis Tectonic Province is underlain by the Olifantshoek Group which was deformed in the Langeberg fold belt during the lower Proterozoic. The westward dipping Olifantshoek Group displays a broad facies zonation parallel to the craton margin which is suggested to have exerted a controlling influence on sedimentation and structure (Stowe, 1983). The base of the Olifantshoek Group is defined by the

Lucknow Formation consisting of quartzite and shale and the 2070 Ma (SACS, 1980) Hartley andesite lavas. The Matsap Sequence overlies the former and comprises terrigenous sandstones, conglomerates and shales extending along the craton margin for at least 300 km. These units are interpreted by Stowe (1983) to represent a fluvial clastic wedge along the craton margin. This is followed by the quartzite/meta-argillite sequences of the Groblershoop Formation which is in faulted contact with the Dagbreek and Sultanaoord

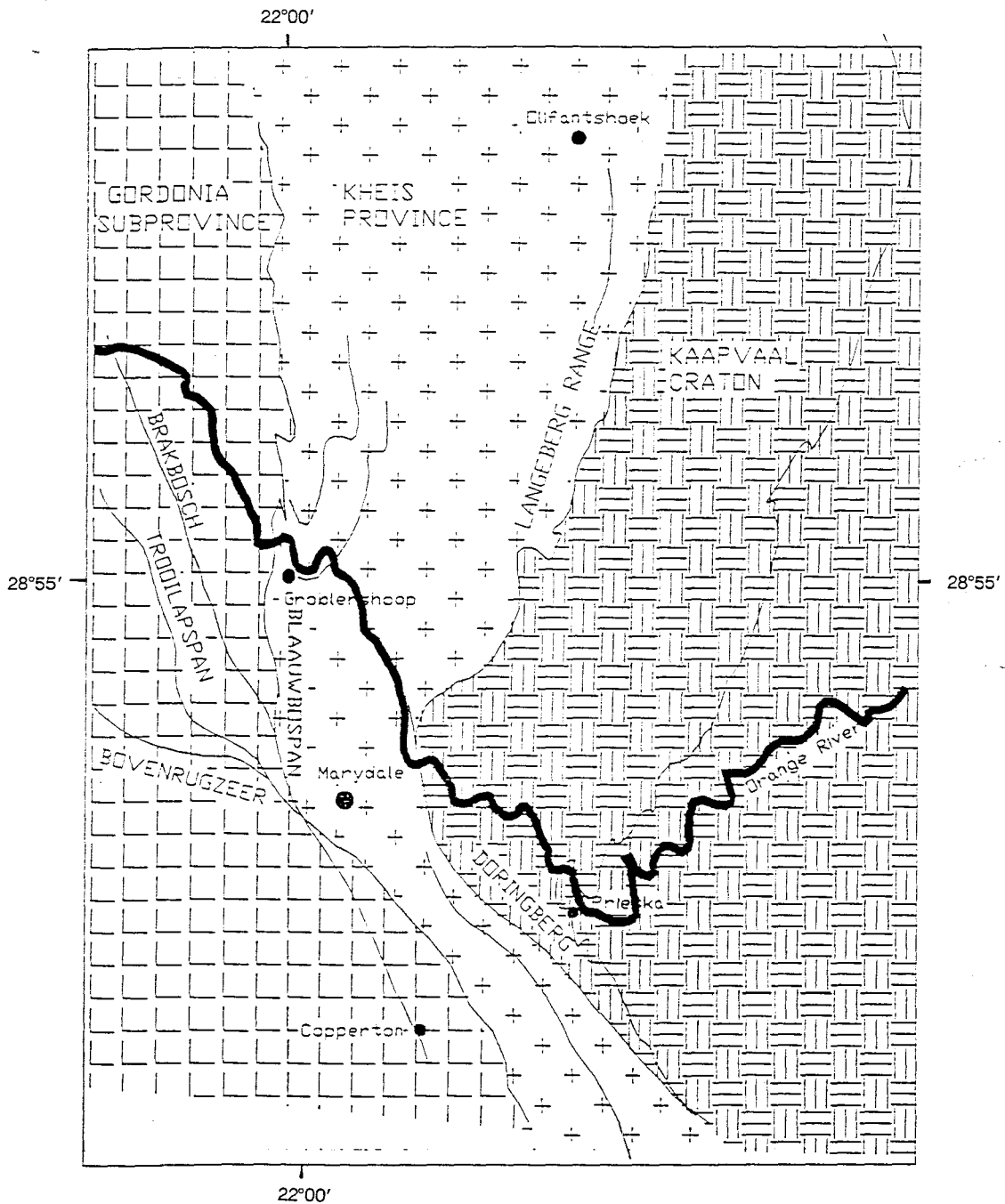


Figure 2.2: The distribution of the various Provinces and their relation to the Kaapvaal Craton.

Formations in the south-west. The Groblershoop Formation is also cut off by the Blaauwbospan Fault. The Dagbreek and Sultanaoord Formations were mapped by Malherbe (1979) and Moen (1980) as underlying the Groblershoop Formation and could represent a distal facies of the Matsap but Vajner (1974c) placed the Sultanaoord Formation above the Groblershoop Formation. Stowe (1983) suggested the Kheis Province represented a "pericratonic shelf on which the Olifantshoek Group was deposited as prograded clastic wedges of progressively increasing maturity".

The Kheis Tectonic Province on the eastern side is underlain by continental basement and Stowe (1983) suggested a stable basement within the province based on stable shelf sedimentation and craton-ward fold vergence and thrust tectonics. He also suggested an increasing basement mobility towards the west. Hartnady et al. (1985) interpreted the Kheis Tectonic Province to be related to thin skinned tectonics where the Kheis rocks were thrust eastwards over the cratonic cover at about 1750 Ma as a series of five nappes. Humphreys et al. (1991), however, reported much higher pressures for the Groblershoop Formation which indicate the Kheis having been to much greater depths within the crust and speculated a larger and deeper Kheis sedimentary basin towards the north, under the Kalahari Group cover.

2.3 THE NAMAQUA TECTONIC PROVINCE.

The Namaqualand Tectonic Province or Namaqua-Natal Province, was recognised by Nicolaysen & Burger (1965) to be part of the Kibaran-aged (1400-950 Ma) orogenic event. The Namaqua-Natal Province covers a vast area in Namibia, where it is well exposed east of Luderitz and is juxtaposed against the Pan African (750-450 Ma) Damara Orogeny. From Namibia, it swings southeast and forms an arcuate belt which wraps around the southern and southwestern margins of the Archaean Kaapvaal Craton from the North Cape through to Natal, in South Africa (Fig. 2.1). De Beer & Meyer (1983) established the southern boundary of the Namaqua Province to be represented by the Beattie static magnetic anomaly. They suggested that this highly conductive, magnetic and dense rock mass at relatively shallow depths could represent partially serpentinitised basic rocks which mark a subducted oceanic crust from the

south.

In the area under investigation, the Namaqua Tectonic Province is subdivided into the Bushmanland Subprovince and Gordonia Subprovince. The Gordonia Subprovince is further subdivided into various tectonostratigraphic terranes, namely, from east to west, the Upington Terrane, Areachap Terrane and Kakamas Terrane.

2.3.1 The Gordonia Subprovince.

According to Hartnady et al. (1985), the Gordonia Subprovince was initially subjected to a period of northward convergent thrusting and folding, followed by an estimated minimum of 140 km of dextral shearing (Stowe, 1984) which caused the prominent northwest trend exposed throughout this subprovince. In the quadrangle under investigation, Hartnady et al. (1985) divided the Gordonia Subprovince into the high-grade Kakamas Terrane and the Upington Terrane which are separated by volcano-plutonic amphibolite complexes of calc-alkaline composition (Geringer, 1979) known as the Jannelsepan Formation (Fig. 2.3). Thomas et al. (1994) recognised similar terranes in this area, but named the Upington Terrane, the Kaaieen Terrane and included the entire volcano-sedimentary sequences of the Areachap Group into the Areachap Terrane, similar to Hartnady et al.'s Jannelsepan Formation. Stowe (1986) proposed the name Upington Terrane for the region of Kheis and Kaapvaal rocks overprinted by Namaqua fabrics and metamorphism that included Vanjar's (1974a) Namaqua Front. According to Slabbert et al. (1989), the salient structural feature of the Gordonia Subprovince is the northwesterly trend of structures and lithologies, which contrast with the northeasterly trend in the Bushmanland Subprovince. The northeasterly trend does, however, become northerly to northwesterly as it approaches the Gordonia Subprovince. For a summary of the stratigraphy of the various tectonostratigraphic terranes within the area under investigation, the reader is referred to the tables in appendix A.

The **Upington Terrane** is primarily composed of 2000 Ma quartzite-pelite equivalents of the Kheis Tectonic Province which have subsequently been overprinted by Kibaran-aged northwest-trending deformation and metamorphism. The thrust tectonics

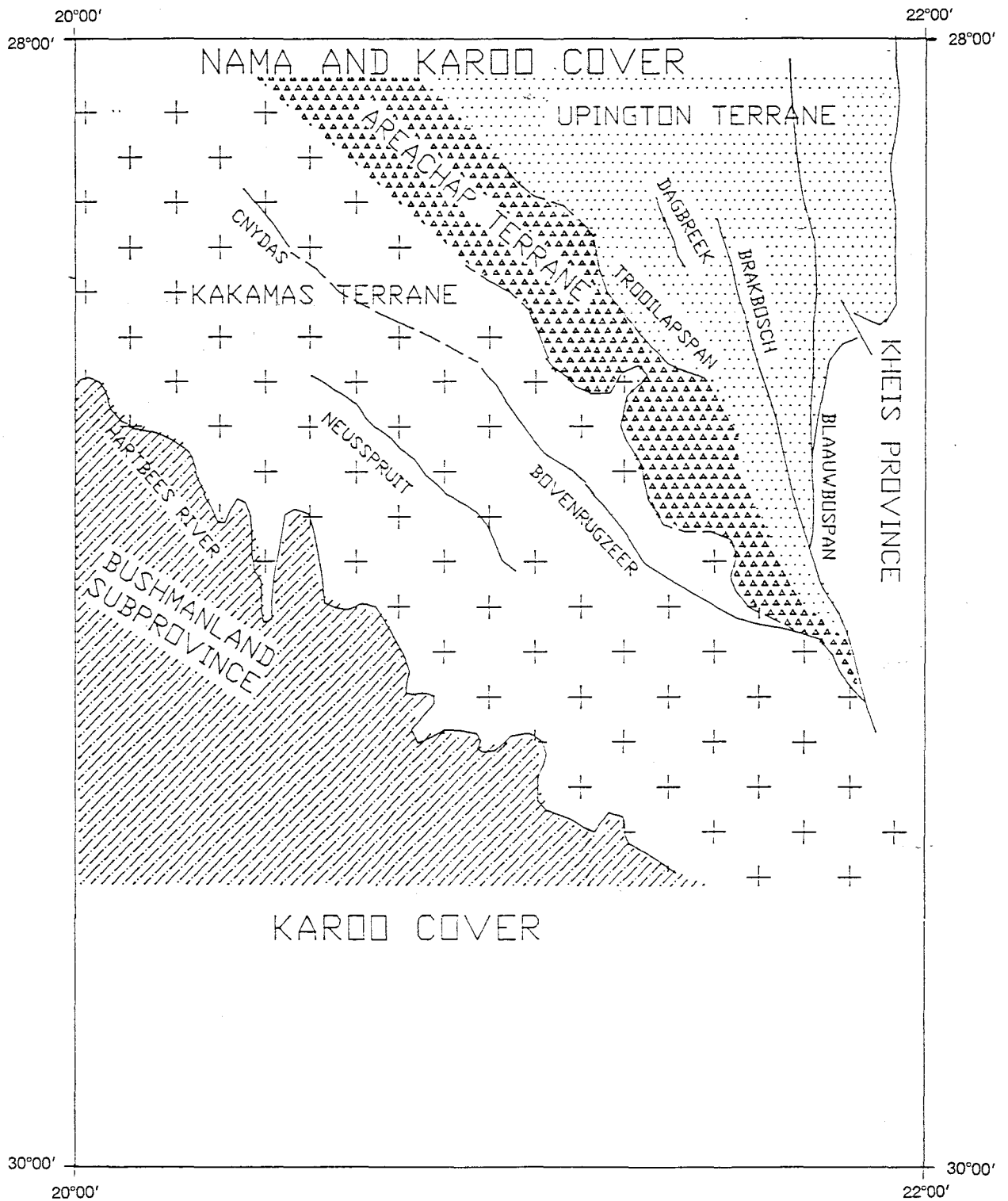


Figure 2.3: The various subprovinces and terranes in the area under investigation, including some of the major faults.

of the Kheis Province extend westwards into the Upington Terrane where, according to Stowe (1983), the rocks were refolded and turned up on edge. Subsequent transcurrent shearing juxtaposed lenticular crustal blocks of different stratigraphic, structural and metamorphic character.

The Upington Terrane is underlain by the Vaalkoppies Group (formally the Kaaien Group) which consists of the Dagbreek Formation of predominantly quartzite with bands of migmatite, amphibolite and serpentinite and the Sultanaoord Formation composed of quartzite and phyllite (Fig. 4.8). The Dagbreek-Sultanaoord successions show progressively increasing sediment maturity while the top of the sequence is marked by fine-grained orthoquartzites with occasional magnetite seams (Stowe, 1983).

The Vaalkoppies Group is overlain by the 1350 Ma rocks of the Namaqua-foliated Wilgenhoutsdrif Group (Moen, 1980; 1987; 1988) which marks the eastern margin of the Upington Terrane (Fig. 4.7). The Wilgenhoutsdrif Group consists of pillowed mafic and felsic lavas, volcanoclastics, banded ferruginous cherts, conglomerates as well as sericitic and chloritic phyllites. The upper contact of the Sultanaoord Formation with the Wilgenhoutsdrif Group is represented by the flaser-structured Kameelpoort quartzite which displays mylonite and serpentinite bands in places. This has been interpreted by Stowe (1980) and Moen (1980) to be an early thrust plane. Moen (1980) described the geochemistry of the mafic lavas as being tholeiitic with "ocean floor" and "within plate" trace element affinities. The high Ti:Zr ratios together with the large amounts of associated rhyolites indicate a continental environment (Moen, 1980).

Overlying these are the relatively undeformed volcano-sedimentary sequences of the Koras Group (Fig. 4.7) which has been dated at 1150-1200 Ma (Botha et al., 1979). The volcano-sedimentary sequences of the Koras Group consist of felsic lavas (calc-alkaline, rhyolitic to dacitic) and basic lavas (basaltic to andesitic; calc-alkaline to tholeiitic, with associated pyroclasts and tuff) as well as conglomerates with angular to well rounded clasts ranging from pebbles to large boulders and medium to coarse-grained sandstones commonly displaying cross-bedding and ripple marks. Moen (1987)

interpreted the Koras Group to have been rapidly deposited in a series of half-grabens in an extensional regime, possibly during the final stages of the Namaqua event. He suggested the Brakbosch and Blaauwbospan fault systems as being possible conduits for the felsic and basic volcanism.

The western boundary of the Upington Terrane is considered by Stowe (1983) to be represented by the ± 1100 Ma Straussberg-Trooilapspan Shear in the north which joins the Brakbosch Shear in the south. The Trooilapspan Shear also marks the western extremity of the Dagbreek Formation and a change to a much higher metamorphic grade. This shear is partially obscured by Quaternary sediments in the south and the intrusion of syn to late-tectonic granites in the north and is characterised by migmatites in the Dagbreek Formation to the east of Upington.

The Upington Terrane could therefore represent an eastern marginal thrust belt developed during the Namaqua orogen as a result of possible arc-continental collision just prior to northward movement of the Namaqua plate and the accompanying Koras volcanism (Hartnady et al., 1985; Stowe, 1986 and Thomas et al., 1994).

The **Areachap Terrane** comprises the Areachap Group which is a narrow belt of amphibolitic grade rocks incorporating the following: the basal metapsammites of the Sprigg Formation, the metavolcanics of the Jannelsepan Formation and the upper metapelitic rocks of the Bethesda Formation (Fig. 3.1; 3.2). The Areachap Terrane has also been extensively intruded by the Keimoes Suite granitoids. The eastern and western boundaries of this terrane have been defined in the previous sections. It extends from north of the Areachap mine, northwest of Upington, where it is covered by Nama and Karoo rocks, southwards towards Copperton, where it disappears under the Karoo cover. The Areachap Group has been cut off by the Bovenrugzeer Shear near Putsonderwater and the correlation of the group with the Copperton Formation is purely on the grounds of styles of mineralisation and their amphibolite contents.

Geringer & Ludick (1990) found two distinctly different amphibolite units within the Jannelsepan Formation. A lower massive, hornblende-biotite amphibolite with no

pyroxene and an upper pyroxene amphibolite with the diopsidic clinopyroxene as the major mafic constituent. Chemically, they found that the former resembles a low-K tholeiite and calc-alkaline basalt whereas the latter shows strong shoshonitic affinities. This, in addition to trace element ratios and REE distribution curves, points towards an arc-related association for the Jannelsepan assemblage (Henderson, 1984; Morrison, 1980; Thompson & Fowler, 1986). Geringer & Ludick (1990) proposed a two stage model for the Areachap Terrane: initially low-K tholeiite and calc-alkaline volcanism were produced during subduction of the oceanic crust. This was followed by shoshonite dominated volcanism during the more mature stage of the arc. The second stage involved the large-scale development of calc-alkaline granite of the Keimoes Suite subsequent to closure of the ocean and continental collision or underplating.

The northwest-trending **Kakamas Terrane** (Fig. 2.3) features higher metamorphic grades as well as a host of early, syn and late-tectonic plutons, predominantly of granitic composition, that include 1200-1100 Ma biotite-hornblende gneissic granodiorites, granitoids and charnockites. Hartnady et al. (1985) suggested on geochemical and mineralogical grounds, that the quartz porphyry of the Koras Group represented an extrusive phase of the late-tectonic granites. These granitic plutons, which will be discussed in more detail in Section 2.3.3., intruded the highly deformed predominantly amphibolite-grade metasedimentary sequences. Apart from these granitoids, which are grouped together as the Keimoes Suite, the Kakamas Terrane consist mainly of rocks belonging to the Hartbees River Complex and the Korannaland Sequence. Metapelitic, felsic and basic gneisses as well as calc-silicate rocks form the bulk of the Hartbees River Complex.

The western boundary of the Kakamas Terrane is considered by Thomas et al. (1994) to be along the Hartbees River Thrust where the Bushmanland Subprovince is juxtaposed against the Kakamas Terrane, (Fig. 2.3). The Hartbees River Thrust consists of a zone of thrusts rather than a single thrust fault (Harris, 1985). The western margin of this terrane has, however, often been correlated with the Neusspruit shear zone (Fig. 2.3) (Van Bever Donker, 1983). The eastern boundary of the

Kakamas Terrane is marked by the contact between the plutonic rocks of the Keimoes Suite with distorted sequences of migmatitic, biotite and aluminous gneisses of the Bethesda Formation and the kinzigite of the Rateldraai Formation, which forms the base, as well as the western extremity of the Areachap Group. Calc-silicates, metapelites and semipelitic to psammitic metasediments form the bulk of the Korannaland Sequence and are separated from the migmatised pelitic gneisses and volcanogenic amphibolites of the 1600-1300 Ma (Barton & Burger, 1983; Cornell et al., 1990) Areachap Group by granitoids of the Keimoes Suite. In the southern portion, this divide is formed by the Bovenrugzeer Shear.

The extensive, well mineralised pegmatite belt (Hugo, 1969), which stretches from the Vioolsdrif area in the west, into the Upington area where it bends southwards to follow the northwest trend of the Gordonia Subprovince, and disappears beneath Quaternary cover southeast of Kenhardt, is present almost exclusively within the Kakamas Terrane. The pegmatites will be discussed in more detail in Section 3.5.4.

2.3.2 The Bushmanland Subprovince.

The Bushmanland Subprovince is bound in the east by the Hartbees River Thrust (Fig. 2.3) and consists of intensely deformed, high-grade gneisses, granulites and granitoids. Controversy over the differentiation between the pre-Namaqua (Eburnian-aged) basement and highly deformed supracrustal Namaqua rocks as well as over the genesis of certain gneisses has not been resolved. The Bushmanland Subprovince may comprise a number of thrust-bounded tectonostratigraphic domains (Colliston et al., 1989; 1991) although the precise nature and geometry of these is still a matter of debate. A summary of the stratigraphy of the Bushmanland within the area under investigation is given in Table A-1, in the appendix A.

The most distinctive and economically important lithological assemblage is the quartzite-aluminous schist which is widespread throughout the Bushmanland. At the Aggeneys and Gamsberg massive sulphide deposits, Moore (1980) interpreted the precursors to these paragneisses to have been deposited in a continental/shallow

marine to deep marine environment. Hartnady et al (1985) concluded from Sm-Nd isochron age dating (quoted from P.J. Betton) that the Bushmanland metasedimentary gneisses accumulated between 1700 and 1600 Ma ago. The dominant tectonic fabric in the Bushmanland is east-west, and has been related to regional metamorphism at 1200 Ma (Clifford et al., 1981). The \pm 1200 Ma and younger Namaqua-age megacrystic and augen gneisses and charnockite of batholithic proportions such as the Gladkop, Klein Namaqua and Spektakel Suites intruded the paragneissic rocks in a sheet-like fashion. Reid & Barton (1983) interpreted the origin of these granitoids as having been derived from anatexis of a 2000-1900 Ma crustal protolith. In the area under investigation, the regional tectonic trend swings north north-west, having been influenced by the adjoining Gordonia Subprovince. The late-tectonic granitic plutons are also subject to the Gordonia trend (Hartnady et al., 1985). Subordinate, mineralised mafic-anorthositic suites, such as the Koperberg Suite, occur throughout the Bushmanland Subprovince and are mainly confined to belts of paragneisses (Joubert, 1974) as well as along the Pofadder Lineament, where they have been altered to amphibolites (Joubert, 1986). Albat (1984) pointed out the proportional increase, towards the west and south, of granitic rocks, as well as the increase in metamorphic grade from amphibolite grade in most of the Bushmanland Subprovince to granulite facies in the Garies-Kliprand area.

2.3.3 Intrusive Rocks.

Numerous intrusive rocks appear in the Gordonia Subprovince between the Hartbees River Thrust in the west and the Brakbosch Fault in the east (Fig. 2.4; 2.5 & 2.6). On the basis of their structural relationships, Stowe (1983) grouped the granites in the area as pre-tectonic, syn-tectonic and post-tectonic with respect to the main Namaqua event and described them as follows:

The **pre-tectonic** gneisses (Fig. 2.4) are characterised by intense foliation and lineation as well as the common occurrence of deformed mafic xenoliths isoclinally folded with axial-planar foliation which bears testimony to their intrusive nature. The pre-tectonic granites are best developed in the Kakamas Terrane and display irregular contacts

with the country rocks. The Riemvasmaak Gneiss, for example, is intrusive into the metasediments of the Korannaland Sequence and the Koelmanskop Metamorphic Suite and is, in turn, intruded by the syn to late-tectonic granites of the Keimoes Suite. Slabbert et al. (1989) consider the "Namaqua" foliation to be related to the Kibaran event and the tectonic fabric of the Riemvasmaak Gneiss to be a result of the older Eburnian event. The pre and syn-tectonic granites cannot always be distinguished from one another. These granites resemble the I-type granites of White & Chappell (1977) (Geringer et al., 1988).

Small rounded xenoliths, weaker foliation and discordant, sharper contacts distinguish the **syn-tectonic** granites (Fig. 2.5) from the older pre-tectonic granites. They range from tonalitic to alkaline in composition. Two populations based on modal quartz and feldspar composition are recognised by Van Zyl (1981) and termed the Straussberg and Keboes suites (Stowe, 1983).

Intrusive bodies of the Straussberg suite, east of Upington, are irregularly oval shaped with sharp contacts and contain numerous ellipsoidal-shaped xenoliths. Van Zyl (1981) indicated a decreasing strain inwards from the margins of the plutons and that the enclosing gneisses show a high strain envelope. This was interpreted as indicating syn-tectonic intrusion during "which the foliated and strained margin was ductile, but the core remained fluid".

The Keboes suite occurs between Louisvale and Keimoes, and is differentiated from the Straussberg suite by its "alkali granite composition, irregular form with foliated concordant contacts and the occurrence of numerous more mafic zones and inclusions" (Van Zyl, 1981).

The syn-tectonic granites seem to form a granite series (Pitcher, 1979) from the katazonal granites in the west of the Kakamas Terrane, which intruded high-grade metamorphites to mesozonal intrusions in the central parts and epizonal granites further east along the Upington Terrane (Geringer et al., 1988). This, according to Stowe (1983), is a reflection on the variable amounts of uplift and ductility which

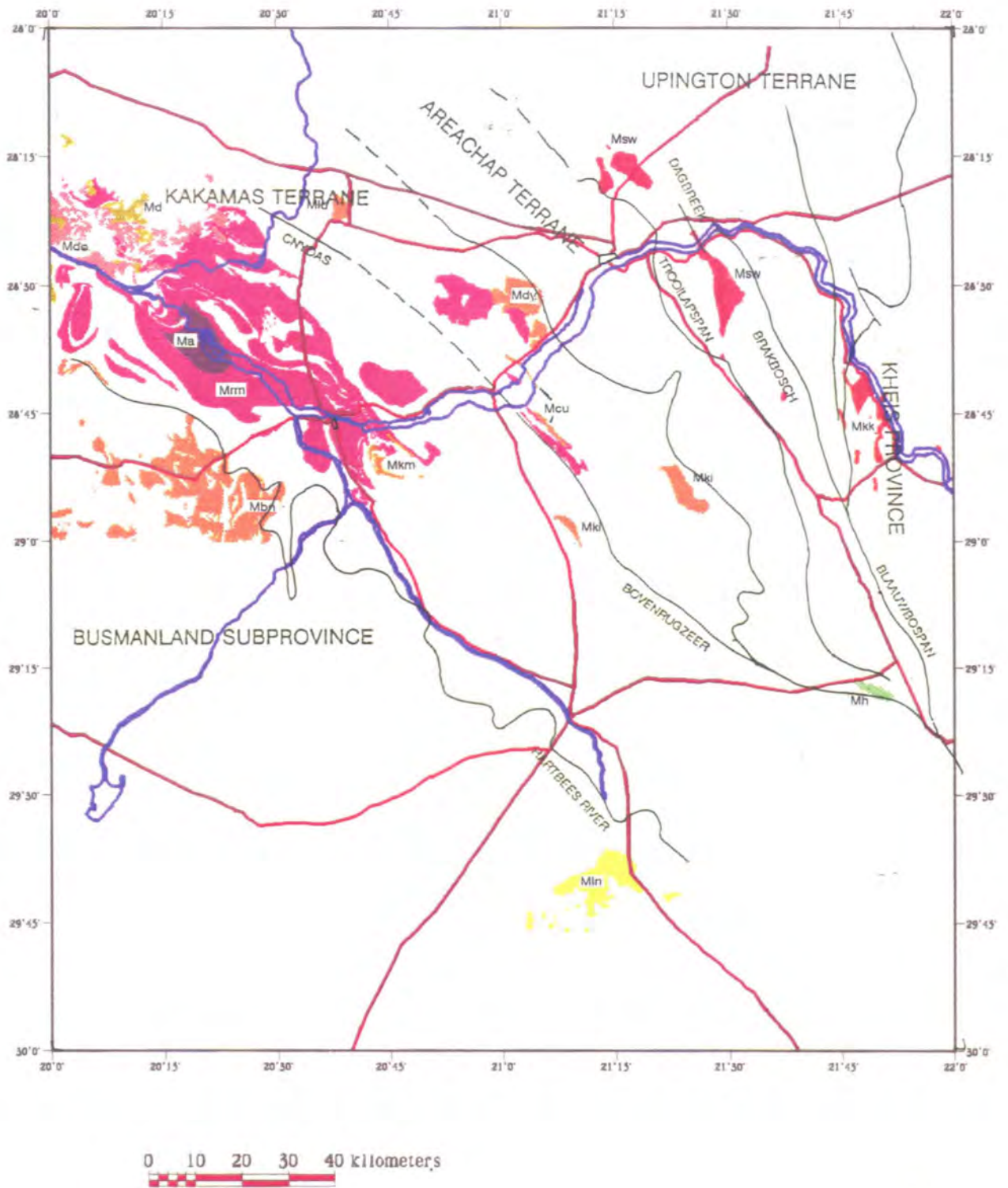


Figure 2.4: The distribution of pre- to syn-tectonic intrusives in the area under investigation.

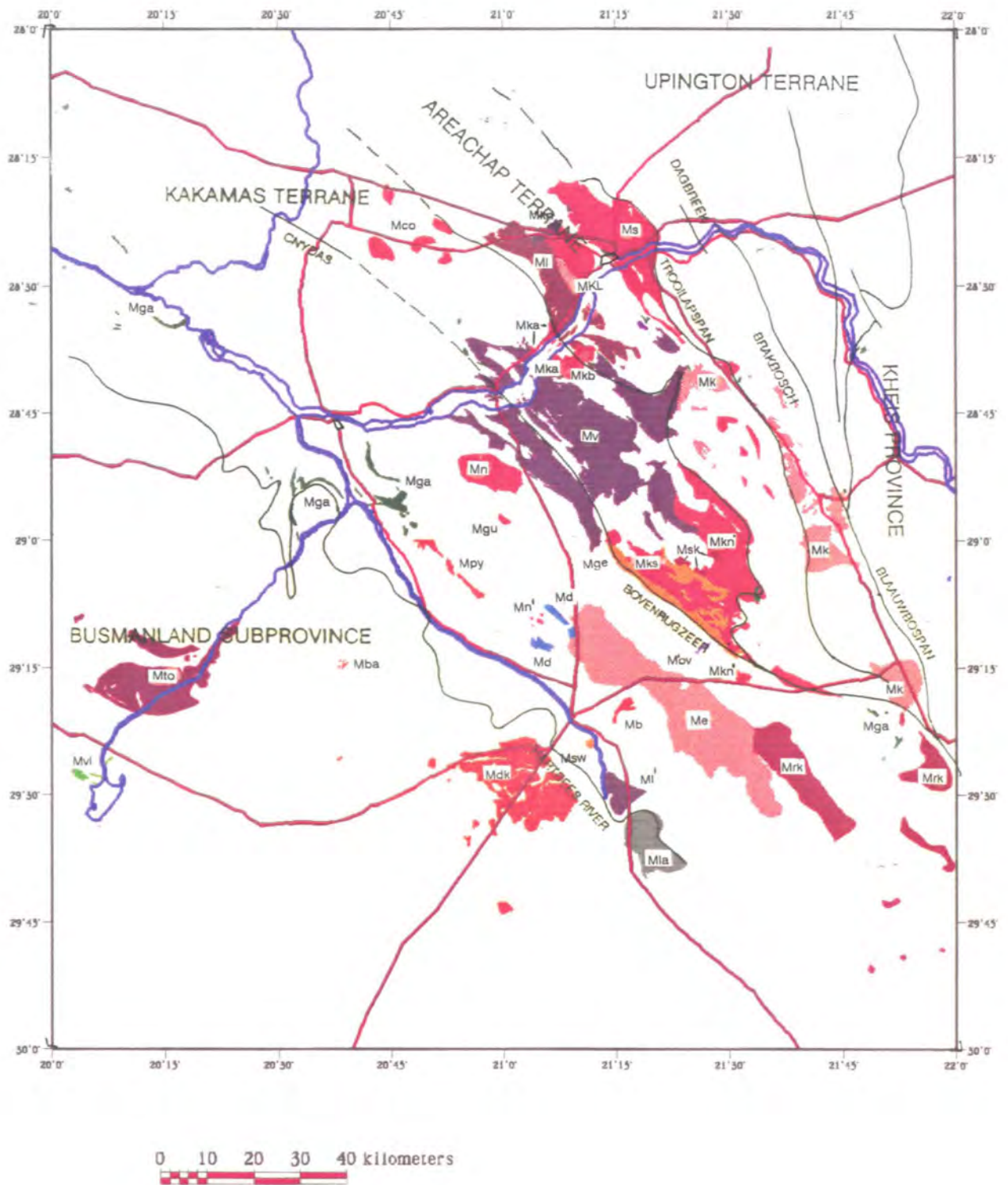


Figure 2.5: The distribution of syn to late-tectonic intrusives in the area under investigation.

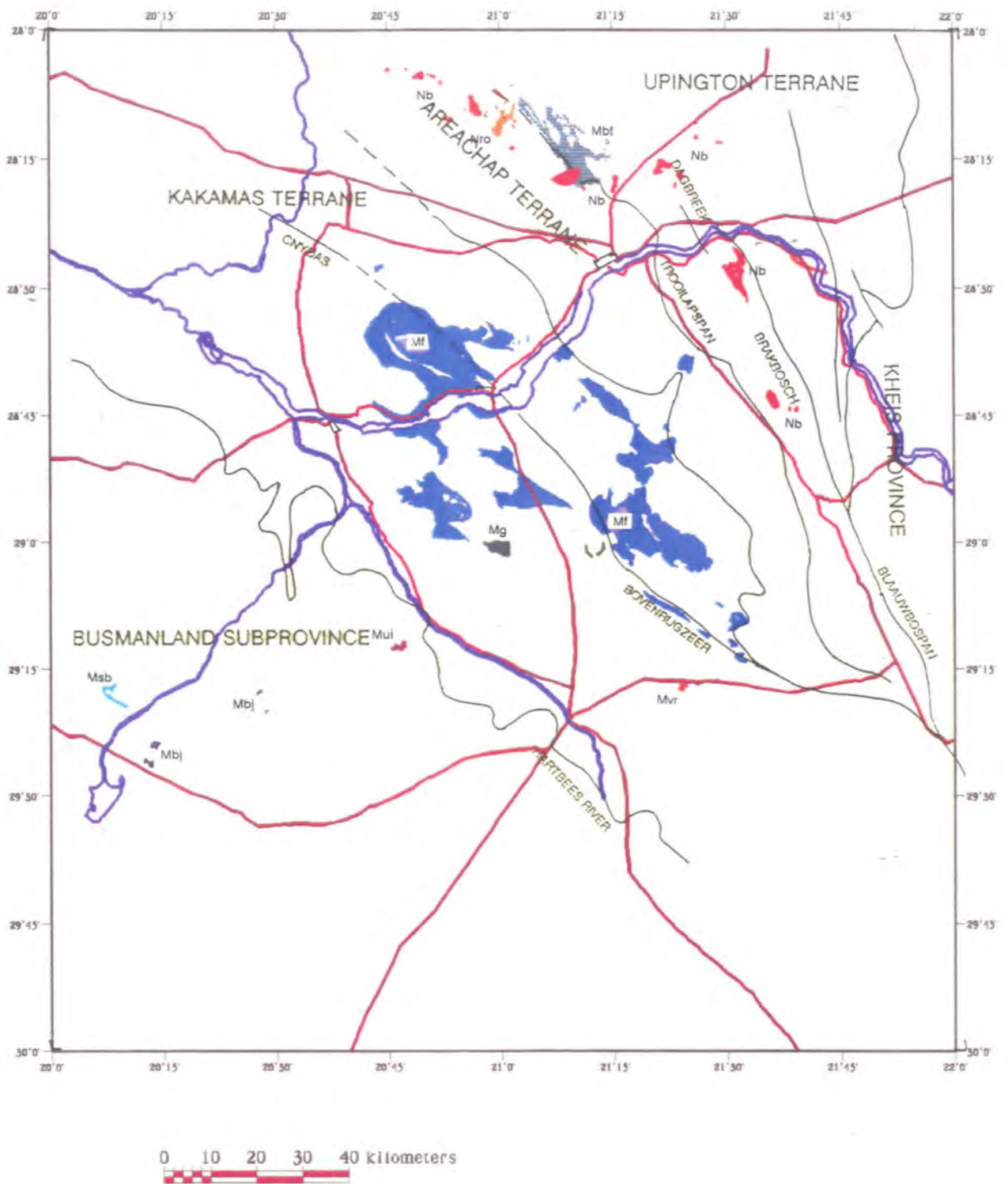


Figure 2.6: The distribution of post-tectonic intrusives in the area under investigation.

occurred in the area exposing deeper crustal levels in the west and higher-level intrusions in the east.

Post-tectonic (Fig. 2.6) granites show marked contact metamorphic aureoles and contain no foliation. The charnockitic adamellite (Poldervaart, 1966; Von Backström, 1964; Van Bever Donker, 1980) is the least fractionated of the Keimoes Suite granitoids. Von Backström (1964) suggested that it was derived from "a tholeiitic magma situated in depth and partially contaminated by xenoliths of country rock", whereas Shultz (1978) indicated that geochemically, these rocks were compatible with an argillaceous rather than an igneous source (S-type). Geringer et al. (1988) suggested that the late to post-tectonic calc-alkaline granites of the Keimoes Suite, the majority of which are of the I-type, form a composite batholith along the eastern margin of the Namaqua Tectonic Province. They are related to the Namaqua tectogenesis and intruded during the closing stages of the folding event between 1250 and 1100 Ma, before the development of the major transcurrent shears.

Some characteristics of the Keimoes Suite are given in Table A-6, in the appendix. Most granitoids of the Keimoes Suite are enriched in TiO_2 , Fe_2O_3 , FeO , MgO and P_2O_5 , SiO_2 , Al_2O_3 and CaO correspond with standard values, while Na_2O and K_2O are slightly depleted when compared to standard quartz monzonite (Geringer et al., 1988). Geringer et al. (1988) noted that the Keimoes Suite granites are highly enriched in Rb and Th with slightly lower Ba values and a significant drop in Nb values. Ce and Sm are enriched relative to their adjacent elements. They pointed out that these normalised curves correspond well with volcanic-arc granites from Chile and concluded that the intrusion of the Keimoes Suite occurred as a result of compression formed by the underriding of the Kaapvaal Craton by the Namaqua Plate.

2.4 DISCUSSION

Numerous geodynamic models have been proposed for the area by various authors, and are generally consistent with either thermal plumes (Kröner, 1979; Botha & Grobler, 1979; Stowe, 1980) or plate tectonics (Botha & Grobler, 1979; Malherbe,

1979; Van Zyl, 1979,1981; Moen, 1980, Stowe, 1980, 1983; Van Zyl, 1981). Some of these ideas are incorporated in the following summary.

The precise history of the pre-Eburnian orogeny is unclear. What is understood, however, is that crustal shortening during the Eburnian followed sedimentation on a broad shelf along the Kaapvaal margin culminating in the deposition of the Olifantshoek sequence. This crustal shortening caused thin-skinned thrusting, documented in the Kheis Tectonic province by Hartnady et al. (1985). The eastern Namaqualand Province comprising the Kaapvaal Craton, Kheis Province and Upington Terrane form a complex eastern continental foreland to the Namaqua Orogen. This assemblage of crustal domains was possibly accreted during the Eburnian orogenic event at about 1750 Ma prior to the deposition of the early Kibaran supracrustal sequences.

The onset of the Kibaran history in the eastern Namaqualand probably commenced with extension of the relatively thin Eburnian crust (Thomas et al., 1994) resulting in rifting and the formation of an ocean basin, the dimensions of which are uncertain. This ocean domain is represented by the Areachap Terrane. The circumstances surrounding the formation and timing of the Wilgenhoutsdrif Group and its relationship with the Areachap Group are somewhat enigmatic. Contact relations discussed by Moen (1988) clearly indicate that the Wilgenhoutsdrif Group postdates the Sultanaoord Formation, and a likely age at around 1400 Ma is estimated (Barton, 1983). Moen (1988) showed that, chemically, the tectonic setting of the Wilgenhoutsdrif lavas is transitional between oceanic (spreading centre or ocean island) and continental, and not orogenic. A continental influence is also supported by the abundance of felsic lavas. Geringer (1979) pointed out that the geochemistry of portions of the Areachap Group indicates an orogenic origin for the amphibolites. Both the Areachap Group (Geringer, 1979; Geringer & Ludick, 1990) and the Wilgenhoutsdrif Group (Moen, 1988) have been related to subduction due to closure of an intracontinental ocean during the Kibaran Orogeny, but are separated by the Brakbosch Fault and Vaalkoppies Group in the north.

Convergence of the Bushmanland "microcontinent" against the Kheis-Kaapvaal Craton led to the main Kibaran orogenic event at about 1100 Ma and caused the intense deformation and medium- to- high-grade metamorphism observed in the region. This collision event also caused widespread melting inducing the generation and emplacement of the granitoids between 1200 and 1050 Ma. Further northward-directed movement resulted in the formation and reactivation of numerous steep transcurrent dextral shears along the eastern margin of the Namaqua Tectonic Province which initiated the formation of a series of strike-slip rift basins into which the Koras group of volcano-sedimentary sequences was deposited.

Chapter Three

3. METALLOGENIC FRAMEWORK.

3.1. INTRODUCTION.

The aim of this chapter is to provide an introduction to the various mineral deposits present in the area under investigation and to familiarise the reader with some of the opinions held by various authors on the genesis and models of the respective deposits.

There are four main phases of mineralisation present in the Northern Cape, namely:

- a) Pre-deformational base metal deposits related to the Areachap Terrane and Bushmanland Subprovince.
- b) Syn to post-deformational deposits related to acidic and ultramafic intrusives.
- c) Deposits related to metamorphism.
- d) Deposits related to surficial processes.

With the exception of the base metal deposits, these phases of mineralisation are generally, but not exclusively, bound by the different tectonostratigraphic terranes (as discussed in chapter two). All the deposits with the exclusion of surficial deposits and hydrothermal vein deposits related to intrusion of the Karoo dolerites were formed during the Kibaran-aged, and possibly the Eburnian-aged events (the Proterozoic).

The Proterozoic.

3.2 THE KHEIS PROVINCE.

The Kheis Tectonic Province is a poorly mineralised unit with no known mineral deposits of any economic significance. Near Groblershoop, to the east of the study area, some minor Cu occurrences are present in greenstone successions within the Groblershoop Formation of the Brulpan Group, but no literature is available on these

deposits.

3.3 THE UPINGTON TERRANE.

The Upington Terrane is characterised by numerous poorly mineralised base metal occurrences which are predominantly associated with faults and shear zones in the Koras and Wilgenhoutsdrif Groups (Fig. 4.7). No significant examination has been made on the metallogeny of mineral deposits associated with the Koras or Wilgenhoutsdrif Groups. Copper-nickel sulphide deposits, occasionally developed with talc, occur in amphibolitic and chloritic metavolcanics of basic affinity (greenstones) of the Wilgenhoutsdrif Group. Copper sulphides are also known along breccia zones and shear zones within rocks of the Koras Group and were probably introduced by metal-carrying brines along faults associated with the development of rifting during Koras times.

Lenses of serpentinite with asbestos veins sporadically intrude the quartzites of the Dagbreek Formation (Vaalkoppies Group) and are situated within close proximity to the Brakbosch Fault (Fig. 4.8).

3.4 THE AREACHAP TERRANE.

3.4.1 Stratiform deposits.

3.4.1.1 The Areachap Cu-Zn deposit.

The Areachap Cu-Zn stratiform massive-to-partially disseminated sulphide deposit is situated on the farm Areachap 462, some 20 km northwest of Upington (Fig. 3.1). The deposit has been described by Voet & King (1986) and is contained within the steeply dipping Jannelsepan Formation of the Areachap Group. The rocks that host the sulphides constitute mafic and felsic units comprising amphibolite, amphibole gneiss, biotite gneiss, pelitic gneiss and lenses of calc-silicates which have been subjected to amphibolite-grade metamorphism. The largest concentration of sulphides is contained

within a 40-80 m thick banded, foliated, garnet-biotite schist and grey quartzite thought by Voet & King (1986) to originally represent shales, impure quartzites and arkoses. The hangingwall is a gneissic amphibolite while an amphibole gneiss and massive amphibolite form the footwall. The contacts between the ore body and the host rocks are always sharp and the footwall contact appears to be faulted or sheared. There appears to be no clear wall-rock alteration, although chlorite and anthophyllite have been recorded near the ore body.

The ore-body outcrops at surface where it has suffered supergene enrichment. Gossan is developed to an approximate depth of 70 m. The ore minerals consist mainly of pyrite and pyrrhotite with chalcopyrite, sphalerite and iron oxides being important secondary constituents. Accessory minerals include galena while minor amounts of tellurium and bismuth occur in the galena. The abundance of Zn increases towards the outer periphery of the ore body. The ore-body is estimated to contain 8,9 Mt of disseminated sulphide grading 0,47% Cu and 2,24% Zn or 6,7 Mt of massive sulphide grading 0,59% Cu and 2,88% Zn. The average grade of Ag is 4,6 g/t and of Au is 0,07 g/t.

The presence of metamorphosed mafic and felsic lavas interbanded with metapelites (with which the ore body is associated) as well as the outward zonation of Zn and iron oxides are consistent with a syngenetic volcanic exhalative origin for the Areachap deposit.

3.4.1.2 The Bokspuits Fe-Cu deposit.

In the Bokspuits area, northwest of Putsonderwater (Fig. 3.1), there are numerous sulphide-bearing bodies, mainly associated with the Jannelsepan Formation of the Areachap Group, that have been subjected to amphibolite grade regional metamorphism (Humphreys, 1986).

The **Bokspuits** sulphide-bearing deposits (Geringer et al., 1987) are located on the farm Bokspuits 118. The host rocks are defined mainly by amphibolites of the

Jannelsepan Formation, but the ore bodies are also found within banded iron formation, carbonated basic tuffs and ferruginous chert. The mineralisation is therefore not associated with any particular lithology, but is confined to stratigraphic horizons, confirming its stratabound nature. Geringer et al. (1987) indicated that the geochemistry of most of the amphibolites pointed to an igneous origin and interpreted some of the amphibolites intercalated with the calc-silicate rocks as possible reworked volcanic material or marl associated with carbonaceous sediments.

The mineralisation at Bokspuits occurs along the limbs of the north-south-trending Kraalkop antiform and in the Kraalkop synform. The mineralisation along the eastern limb of the antiform is hosted by an amphibolite schist, whereas on the western limb, the mineralisation is associated with a highly ferruginous chert and muscovite schist near the top of a massive amphibolite unit. The ore body outcrops as gossan which displays boxwork textures. The mineralisation in the Kraalkop synform is not exposed on the surface and occurs as intercalated lenses within a sequence of amphibolite, biotite-quartz-feldspar gneiss and mica schist. The highly chloritised zone of mineralisation is also marked by intensive calcite and quartz veining. A high degree of deformation has thus affected the ore body and a metamorphic overprint on the ore as well as the host rocks is also present (Humphreys, 1986).

The sulphides are represented by pyrite, chalcopyrite, sphalerite and minor pyrrhotite and the oxides include magnetite, hematite and ilmenite. There is a notable absence of galena within the ore. There appears to be a mineral zonation outwards from Cu-rich centres (Theron, 1979) and is displayed as follows: chalcopyrite-magnetite-pyrite ± sphalerite; to pyrite-magnetite; to pyrite; to hematite. Hematite mineralisation is associated with a chert-rich horizon and marks the outer margin of the ore body. Theron (1979) estimated some 1,7 Mt of disseminated sulphide ore grading at 1,5% Cu in four separate ore bodies.

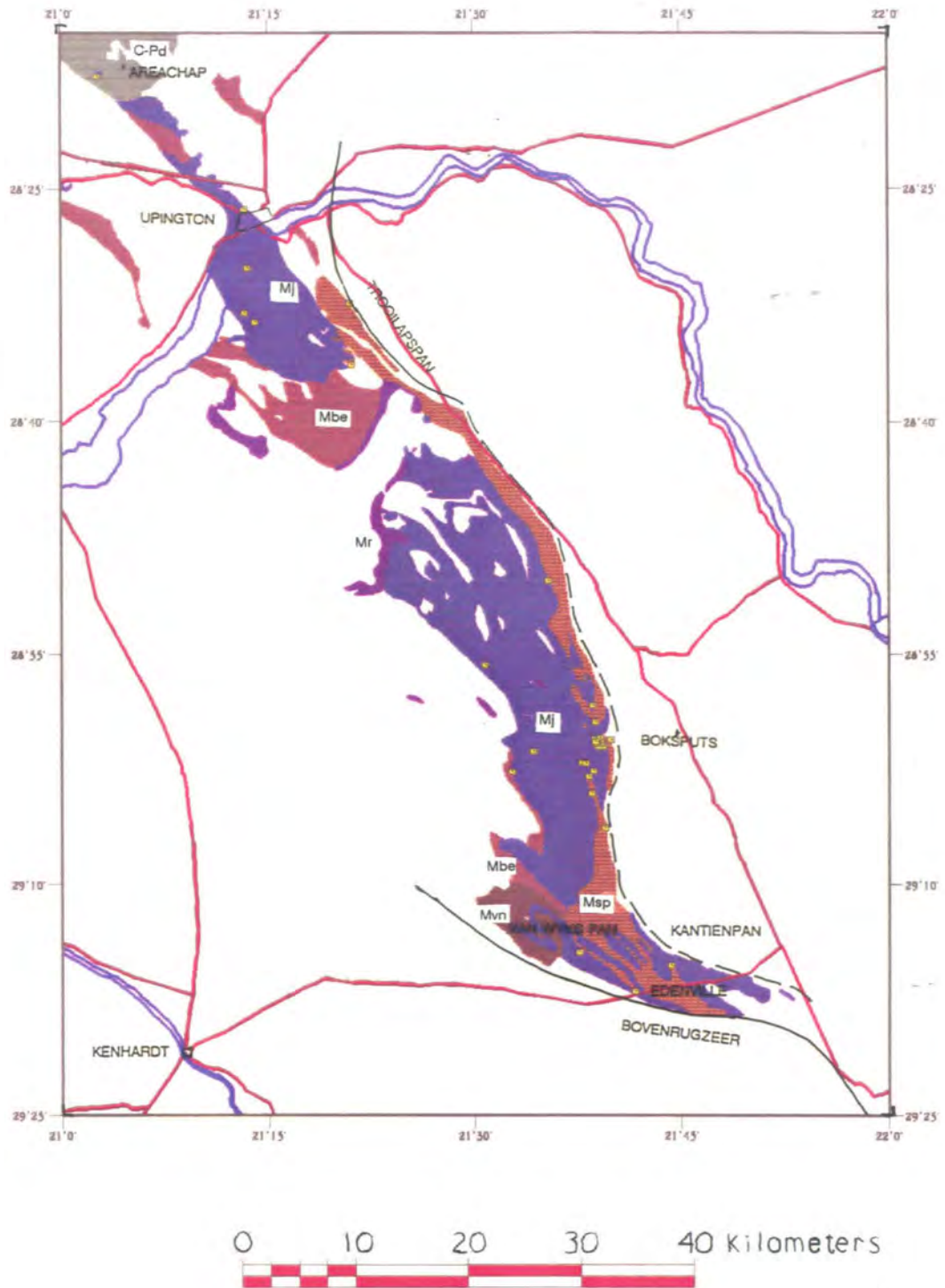


Figure 3.1: The distribution of stratiform massive and partly disseminated sulphide deposits of the Areachap Terrane in the area under investigation including major shear zones.

This deposit has been interpreted by Theron (1979) and Geringer et al. (1987) as a syngenetic volcanic-exhalative massive sulphide deposit. Geringer et al. (1987) went further to suggest, on geochemical grounds, that it is a possible Besshi-type volcanic exhalative deposit.

Other deposits within the Bokspits area include the Van Wyks Pan, Kantienpan and Edenville deposits (Fig. 3.1), which have not been described in the literature. There are three small Cu-Zn-sulphide deposits along the north-eastern boundary of the farm Van Wyks Pan 107. Two of these are situated west of an oblique shear, the one hosted by a quartz-chlorite-amphibole-garnet rock and the other by a serpentine-chlorite rock, both belonging to the Van Wyks Pan Formation. A third ore body, east of the oblique shear, is contained within a quartz-mica schist and quartz-mica-garnet schist which is possibly related to the Bethesda Formation.

3.4.1.3 The Prieska Zn-Cu deposit.

The Prieska stratiform massive sulphide deposit falls outside the quadrangle under investigation, but is discussed here for the sake of completeness, as it is considered to be part of the Areachap Terrane. It is situated at Copperton, some 60 km southwest of Prieska within the volcano-sedimentary sequences of the Copperton Formation (Fig. 3.2) which Cornell et al. (1990) dated at 1205 Ma. This deposit has been described by Middleton (1976), Theart (1985) and Wagener & Van Schalkwyk (1986).

The ore zone lies on the north-eastern limb of a north-south-trending synform and is exposed at the surface in the form of gossan. It is overlain by a quartz-feldspar-hornblende-biotite gneiss in the hangingwall and a streaky quartz-plagioclase-hornblende-biotite gneiss defines the footwall. The host rocks to the sulphides consist of pyritic quartzite, banded gneiss with tourmaline and manganese-rich magnetite zones. Unlike the dominantly chloritic hydrothermal alteration zones observed in the other deposits, the Prieska body displays both siliceous as well as chloritic alteration (Schade & Cornell, 1989).

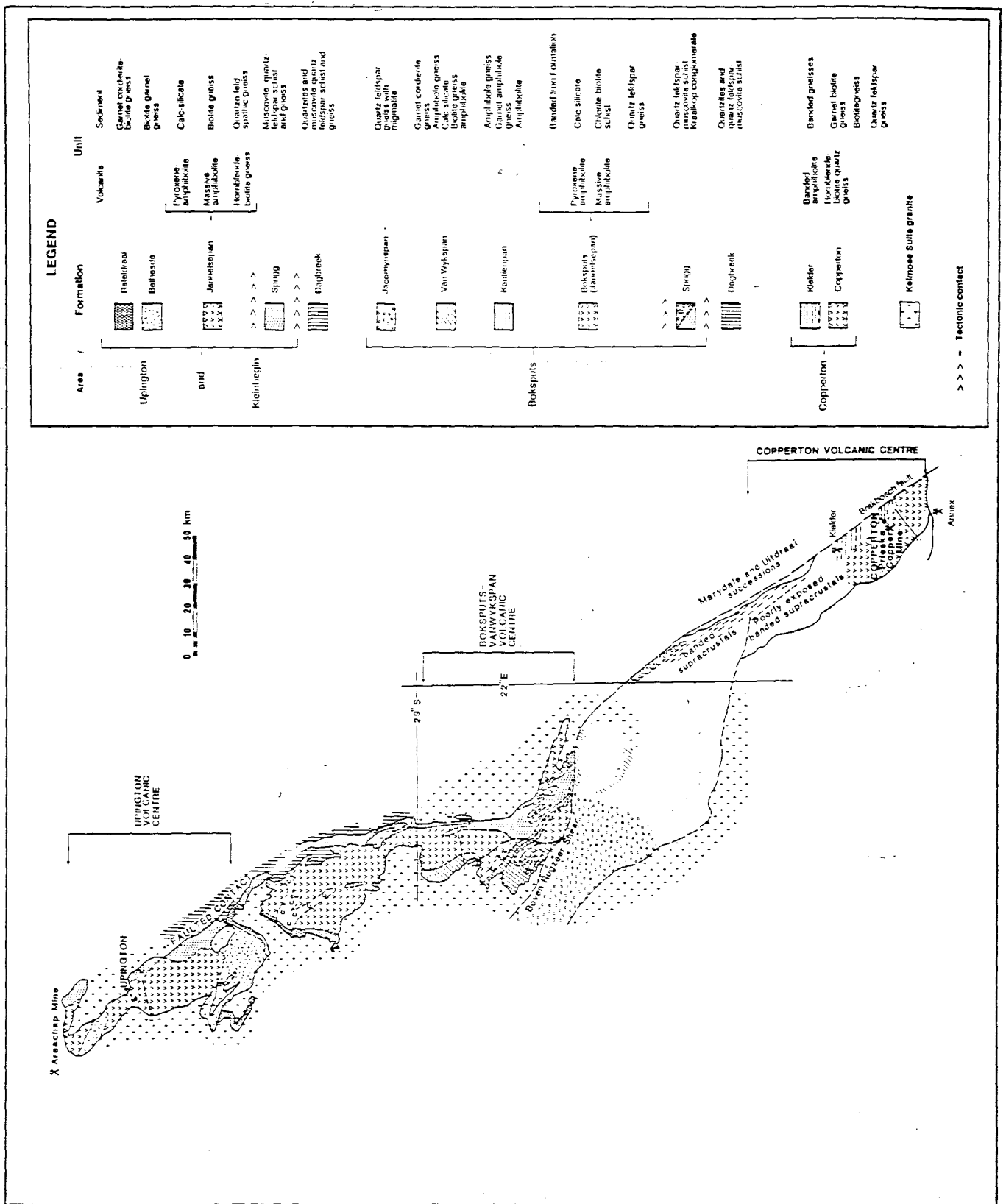


Figure 3.2: The extent of the Areachap Terrane and the areas of mineralisation (after Geringer et al., 1994).

The sulphide mineralisation consists of pyrite with interstitial carbonate and tourmaline, sphalerite with interstitial carbonates and barite, disseminated chalcopyrite and pyrrhotite as well as minor amounts of galena and molybdenite. 47 Mt of ore with a grade of 3.8% Zn and 1.7% Cu was estimated. From 1972 to 1991, 428 600 tons of Cu and 1 009 000 tons of Zn was produced. The Prieska deposit has been interpreted by Middleton (1976) and Uiterwyk & Frick (1985) as having been of an exhalative origin. Isotope data investigated by Theart et al. (1989) strongly supports a volcanogenic origin. Wagener & Van Schalkwyk (1986), however, prefer a sedimentary (Kupferschiefer-type) origin. Some 5 km from the Prieska deposit, the Annex ore body was discovered by airborne magnetics under a 30m thick tillite horizon, but has not been exploited.

3.4.1.4 The Kielder Zn-Cu deposit.

Gorton (1981) described three massive sulphide bodies (Zn > Cu >> Pb) at Kielder, some 12 km northwest of Copperton (Fig. 3.2) and named them K1, K3 and K6. The general characteristics of these ore bodies are summarised in Table 3.1.

Table 3.1: Characteristics of the K3, K1 and K6 ore bodies at Kielder.

K3 ORE BODY	K1 ORE BODY	K6 ORE BODY
Proximal	Proximal-intermediate	Distal
Chloritisation/silicification at base	Chloritisation/silicification at base	No alteration zone
	Symmetrical Mn-rich halo around ore body	Higher Mn-content in wall rocks
		Associated with BIF horizon in hanging wall
Host rocks lack significant amphibolite units	150m thick amphibolite unit in footwall	80m thick amphibolite unit in hanging wall
Significant amount of pyrrhotite, barite	Minor pyrrhotite	Minor, sporadic pyrrhotite
Strong vertical zonation (pyrrhotite-sphalerite)	Vague vertical zonation (Zn increases outwards)	Little vertical zonation
No lateral zonation	Py/Po increases and Py/Magnetite decreases laterally outwards from centre	
		Felsic siliceous chalcopyrite and galena stringer zone below massive sulphides
	Minor galena	Ubiquitous galena

Summarised after Gorton (1981).

These deposits also fall outside the study area, but within the Areachap Terrane, and are included in the discussion for the sake of continuity. These bodies occur as stratabound massive sulphide lenses within granulite-grade quartz-feldspar gneiss, basic granulite and amphibolite of the Copperton Formation. Gorton (1981) interpreted the Copperton Formation in this area to represent interlayered sediments and lavas and provided an age of ± 1300 Ma. The mineralogy consists of pyrite, pyrrhotite, sphalerite, chalcopyrite and galena with gangue minerals of barite, chlorite, phlogopite, apatite, tourmaline and quartz. He also recognised proximal and distal deposits and concluded that these deposits were derived from a volcanic exhalative origin.

3.4.2 Vein Deposits.

3.4.2.1 **Amethyst-bearing vein deposits.**

There are a few fault related amethyst-bearing veins in the area (Fig. 4.14). The most notable deposit is situated on the farm Vaalkoppies 4,8 km east of Upington, where an east-striking vein is present within a fault zone in the Straussberg Granite of the Keimoes Suite and is associated with some Cu mineralisation.

3.5 THE KAKAMAS TERRANE.

3.5.1 Stratabound Deposits.

3.5.1.1 **The Renosterkop Sn-W-Zn deposit.**

The Renosterkop Sn-W-Zn deposit (Saad, 1987; Bowles, 1988) is situated just east of the village of Augrabies, 20 km northeast of Kakamas along the Orange River (Fig. 3.3). Tungsten mineralisation occurs within a zone of quartz-biotite-topaz rock in the form of localised patches of coarse-grained wolframite and is structurally underlain by a quartzo-feldspathic gneiss, known as the pre-tectonic intrusive Riemvasmaak Gneiss. The quartz-biotite-topaz rocks are also host to Sn mineralisation which occurs as low-

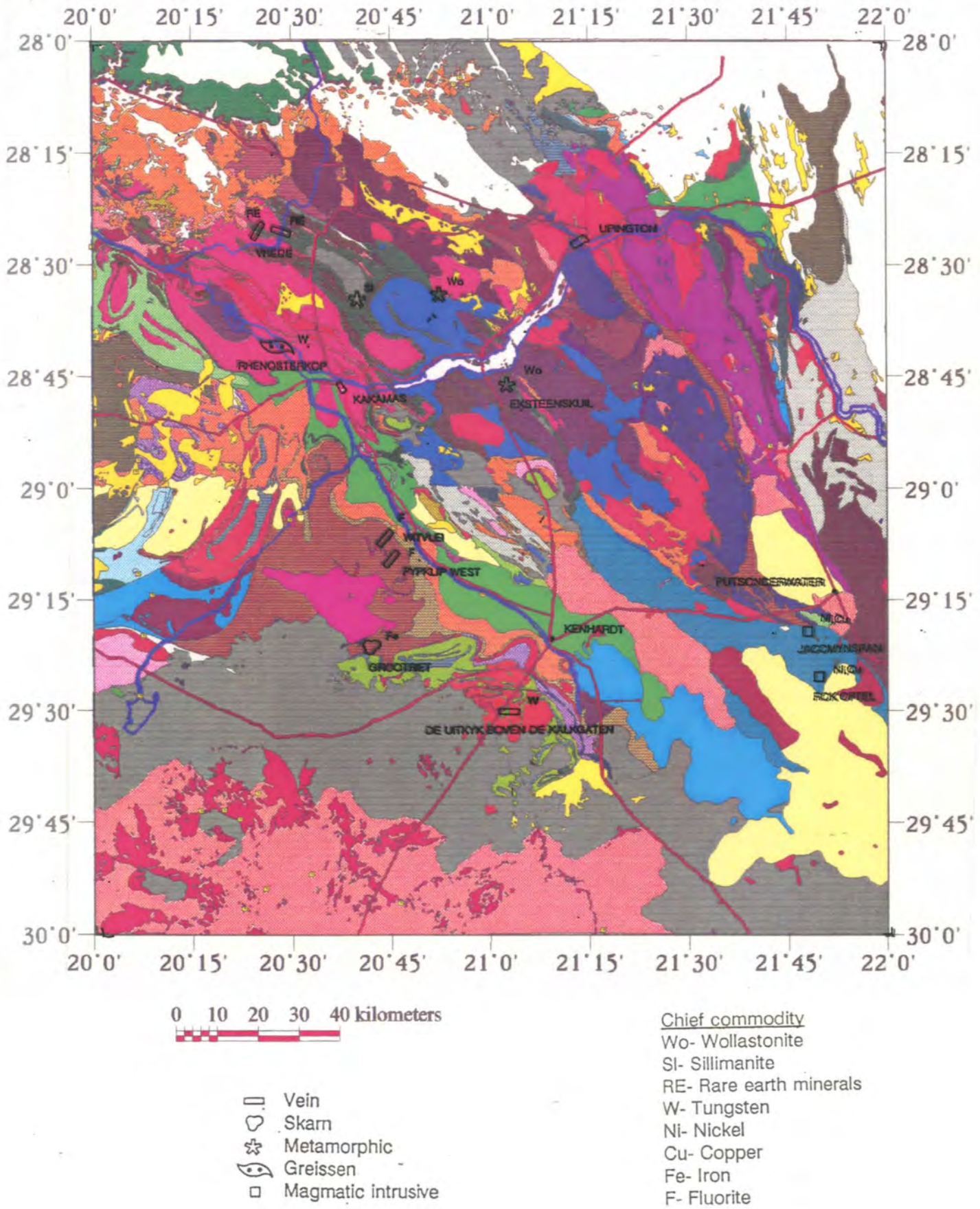


Figure 3.3: Distribution of deposits unrelated to specific metallogenic provinces and mentioned in the text.

grade heterogeneously disseminated cassiterite. Other ore minerals disseminated within the host rock are pyrite, sphalerite, chalcopyrite and arsenopyrite. The quartz-biotite-topaz host rock is surrounded by a chlorite-bearing alteration zone (chlorite replacing biotite) followed by a 50m thick biotite-rich alteration zone (biotite replacing hornblende of the Riemvasmaak Gneiss). The mineralising fluids probably originated from a highly fractionated granitic source at depth.

3.5.1.2 The Rozynebosch Pb-Ag (Zn-Cu) deposit.

The Rozynebosch deposit lies on the Farm Rozyne Bosch 104, just east of the Hartbees River Thrust on the western boundary of the Kakamas Terrane (Fig 4.1). The mineralisation is present within the Rozynebosch Formation (Vyfbeker Metamorphic Suite) consisting of a feldspathic unit (quartz-feldspar \pm biotite gneiss) and a calc-silicate unit (amphibolite, dolomite, marble and calc-silicate rocks with minor granitic gneiss and biotite gneiss). The lenticular-shaped stratabound ore body of disseminated sulphides is hosted mainly by a garnetiferous leucogneiss. The mineralogy of the sulphides includes galena, sphalerite, chalcopyrite and pyrite. Grades in the disseminated ore are up to 2% Pb and 18 g/t Ag. Cu and Zn values are negligible.

Little research has been done on this deposit, and the relationship between the Rozynebosch deposit and other stratiform base metal deposits such as Gamsberg and Broken Hill (SEDEX) further west in the Bushmanland Subprovince and the deposits related to the Areachap Terrane (VOLEX) in the east is enigmatic. The model Pb ages for the Rozynebosch deposit of between 1200-1150 Ma (Köppel, 1980) are younger than the model Pb ages of between 1350-1300 Ma (Köppel, 1980) for the Gamsberg and Broken Hill deposits and the 1600-1300 Ma (Cornell et al., 1990) ages for the Areachap Group. Thomas et al. (1994) suggested a replacement origin for the Rozynebosch deposit.

3.5.2 Vein Deposits.

3.5.2.1 Sn-W-bearing vein deposits.

3.5.2.1.1 The Van Roois Vlei Sn-W Deposit.

The Van Roois Vlei deposit (Wheatley & De Beer, 1986; Bowles, 1988; Smithies & Pirajno, 1989) is situated about 45 km west of Upington on the farm Van Roois Vlei 443 and lies along the northern periphery of the Melkbosrand dome (Fig. 4.9). The Melkbosrand dome comprises the pre-tectonic Riemvasmaak Gneiss intrusive and is surrounded by pelitic quartz-biotite gneiss with minor layers of tourmaline-bearing quartzite of the Toeslaan Formation (Biesje Poort Group, Korannaland Sequence) which has undergone amphibolite-grade regional metamorphism.

Mineralisation occurs in several east-west parallel and subvertical quartz-tourmaline veins (Van Backström, 1950) which were traced down to a "basal" grey granitic gneiss (De Beer et al., 1984). The ore mineralisation includes wolframite, scheelite, cassiterite (often with magnetite inclusions) and lesser molybdenite, chalcopyrite, hematite and ilmenite which are erratically distributed throughout the veins. Quartz, tourmaline and fluorite are the main gangue minerals. Mineralisation appears to be zoned, with cassiterite and pervasive tourmalinisation of the host rock predominating in the east, while chloritisation, albitisation, K-feldspathitisation and sericitisation have also been recognised and are believed to have developed after the tourmalinisation. Exploration by Shell South Africa (Pty) Ltd revealed some 2,5 Mt grading at 0,38% WO_3 and 0,4% Sn representing a reserve of 15 000 tonnes of metal in the main vein zone.

Smithies & Pirajno (1989) visualised a highly fractionated, specialised granite at depth for the source of the mineralising fluids due to the abundance of granophile elements within the ore and gangue and suggested deposition of the ore in a boron-dominated greisen. Bowles (1988) speculated on the possibility of the ore having been remobilised from a tourmaline-rich horizon found in the area, which is thought to represent boron-

rich metasediments derived from a metal-rich volcanic exhalative.

3.5.2.1.2 The Bokspuits W-Sn deposit.

The Bokspuits deposit, situated 2 km west of the Van Roois Vlei deposit on the farm Bokspuits 462 (Fig. 4.9), appears to be closely related to the Van Roois Vlei deposit in style and composition. Mineralised veins, emplaced parallel to the foliation of the host rock (Toeslaan Formation) lie axial planar to tight, steeply northwest plunging F2 folds. Here the biotite-rich gneiss of the Toeslaan Formation is structurally wedged within the Riemvasmaak gneiss. Scheelite, with lesser wolframite and cassiterite, are the ore minerals while pyrite and pyrrhotite are the dominant sulphides. Fluorite seems to be more abundant here than at Van Roois Vlei.

3.5.2.1.3 The Dyasons Klip W deposit.

This deposit occurs on the farm Dyasons Klip 454, 20 km west-southwest of Upington (Fig. 4.9), in the form of NNW-SSE striking veins which commonly pinch and swell. The mineralising veins of quartz and pegmatitic quartz-muscovite contain both wolframite and scheelite as well as tourmaline, arsenopyrite and trace amounts of molybdenum (Bowles, 1988). A porphyroblastic-to-megacrystic pre-tectonic intrusive granite gneiss of the Dyasons Klip Gneiss hosts the mineralised veins.

3.5.2.1.4 The McTaggarts Camp W deposit.

Just east of Dyasons Klip lies the farm McTaggarts Camp 453 (Fig. 4.9), where five mineralised, parallel quartz and pegmatitic quartz veins, hosted by the Dyasons Klip Gneiss, occur (Bowles, 1988). Apart from scheelite and wolframite, the veins also contain arsenopyrite and, in one vein, secondary copper minerals, as well as feldspar, tourmaline and muscovite. These veins have a similar strike to the Dyasons Klip deposits although their dip is much steeper towards the west, and are probably related to the same mineralising system. Some of these veins appear to have intruded older shear planes.

Table 3.2: Characteristics of the tungsten-bearing vein deposits.

VAN ROOIS VLEI PROVINCE			
Van Roois Vlei	Boksputs	Dyasons Klip/McTaggart's Camp	Kalkpunt
Hosted by Toeslaan Formation	Hosted by Toeslaan Formation	Hosted by Dyasons Klip Gneiss	Hosted by Dyasons Klip Gneiss
Quartz-tourmaline veins	Quartz-tourmaline veins	Quartz and pegmatitic quartz-muscovite veins.	Quartz veins
Ore minerals include wolframite, scheelite, cassiterite, molybdenite, chalcopyrite, hematite, ilmenite.	Ore minerals include scheelite, wolframite, cassiterite, pyrite, pyrrhotite.	Ore minerals include wolframite, scheelite, arsenopyrite, molybdenite.	Ore minerals include wolframite, scheelite
Gangue minerals include quartz, tourmaline, fluorite.	Gangue minerals include quartz, tourmaline, fluorite.	Gangue minerals include quartz, tourmaline, feldspar, muscovite.	Gangue minerals include muscovite, tourmaline, feldspar.
Tourmalinisation, chloritisation, albitisation, K-feldspathitisation and sericitisation of wall rocks.			Ferruginisation along vein boundaries.
2.8 Mt grading at 0.38% WO ₃ and 0.4% Sn			
RIEMVASMAAK PROVINCE			
Collinskop	Bok-se-puts	Molapo West	Ekouputs
Hosted by Collinskop and Bok-se-puts Formation. Associated with kinzigite.	Hosted by Collinskop and Bok-se-puts Formations. Associated with kinzigite.	Hosted by Bok-se-puts Formation. Associated with quartz-tourmaline rock.	Hosted by Twakputs Formation and Witwater Gneiss. Associated with tourmalinite units.
Quartz veins.	Quartz veins.		Feldspathic quartz veins
Ore minerals include scheelite, wolframite, pyrite, chalcopyrite, arsenopyrite.	Ore minerals include wolframite, scheelite, chalcopyrite, pyrite, loellingite, bornite, azurite.		
Gangue minerals include quartz, tourmaline	Gangue minerals include quartz, feldspar, garnet, epidote, sericite, biotite, tourmaline.		

Summarised from Bowles (1988).

3.5.2.1.5 The Kalkpunt W deposit.

Two wolframite and scheelite-bearing quartz veins occur on the farm Kalkpunt 452 (Fig. 4.9) (Bowles, 1988) just east of McTaggart's Camp and also display a similar strike to the Dyasons Klip deposits. Their similarity to the Dyasons Klip and McTaggart's Camp deposits extends further in that they are hosted by the porphyroblastic Dyasons Klip Gneiss and the quartz veins contain muscovite, tourmaline and feldspar. All the veins indicate some ferruginisation along the vein

margins.

3.5.2.1.6 The Collinskop W deposit.

This deposit lies 4 km north of the Orange River in the Riemvasmaak Reserve 498 (Fig. 4.12) and has been described by Bowles (1988). A number of quartz-wolframite veins appear over a strike length of about 1 km within a biotite-garnet-sillimanite rock (kinzigite) and a migmatized garnetiferous biotite granulite of the Koelmanskop Metamorphic Suite. These mineralised veins lie subparallel to the contact between the host rock and the porphyroblastic Donkieboud Granite Gneiss of the Eendoorn Suite. Scheelite is a minor constituent, relative to wolframite, and pyrite and chalcopyrite occur in all the veins while arsenopyrite appears only locally. The veins appear to have been emplaced along a complex system of fractures.

The kinzigite, which appears to have a close spatial relation to the mineralised veins, may represent a restite, formed during ultrametamorphism or partial melting of the surrounding garnetiferous biotite granulite (Moen, quoted by Bowles, 1988). This led Bowles (1988a) to suggest a possible link between the tungsten mineralisation and the formation of the kinzigites.

3.5.2.1.7 The Bok-se-puts W deposit.

Several parallel wolframite-scheelite-bearing quartz veins (Bowles, 1988) are situated in the Bok-se-puts area of the Riemvasmaak Reserve 498 (Fig. 4.12). The northwest-striking veins are hosted by biotite schist, which forms lenses within a garnetiferous granulite. The mineralised veins are conformable to the tectonic fabric of the biotite schist. Additional minerals in the ore are chalcopyrite, pyrite, loellingite, bornite and azurite, while the gangue minerals include quartz, feldspar, garnet, epidote, sericite, biotite and tourmaline.

3.5.2.1.8 The Molopo W deposit.

Some small tungsten deposits are found in the southeastern portion of the Riemvasmaak reserve, just west of the Molopo River (Fig. 4.12). Little is known about these deposits, but Bowles (1988) pointed out the association of the mineralisation with thin bands of quartz-tourmaline rock which is interbedded with a garnetiferous biotite granulite. Bowles (1988) suggested a similar genesis of these deposits to those in the Broken Hill Block, Australia, where the origin of the tungsten is related to remobilisation from metal-rich quartz-tourmaline metasediments (Barnes, 1983).

3.5.2.1.9 The Blouputs W deposits.

A few shallow tungsten prospects are found some 2,5 km south of the Orange River on the farm Blouputs 10 (Fig. 4.12) in the form of veins, closely associated with narrow stratiform tourmaline beds which are interbedded with pelitic gneiss (Bowles, 1988). Further south, several feldspathic quartz veins strike north-northwest, conformable to the foliation of the host garnetiferous biotite gneiss of the Twakputs Gneiss (Koelmanskop Metamorphic Suite).

3.5.2.2 Base metal-bearing vein deposits.

2.5.2.2.1 The Lutzputs Fe-Cu-Ag deposits.

North-northeast to northeast trending mineralised quartz veins are located mainly on the farm Cnydas East 439, some 60 km west of Upington (Fig. 4.13). The deposits, described by Bicker & Ralston (1986), occupy shear zones within metasedimentary pelitic gneisses, kinzigite and biotite gneiss of the Toeslaan Formation, which forms part of the Biesje Poort Group (Korannaland Sequence). The host rocks have undergone amphibolite-grade metamorphism as well as intense shearing.

Copper-silver mineralisation is hosted by magnetite-hematite-bearing quartz veins which show considerable variations in grade, width and strike length. The ore bodies commonly display gossan as well as iron staining in the gneiss along the margins of the bodies. Copper staining is present on the iron and quartz veins. Contacts with the

commonly hydrothermally altered host gneiss are usually sharp. The alteration is in the form of propylitisation (chloritisation, kaolinisation and some silicification). The main mineralisation consists of magnetite, hematite, arsenopyrite, chalcopyrite and silica gangue which do not necessarily occur together. Other minerals include pyrite, pyrrhotite, freibergite, sphalerite, bismuthinite and gold as well as secondary minerals such as chalcocite, bornite and digenite. Although grade is highly variable, the best intersection indicated values of 4,23% Cu and 720 g/t Ag.

Bicker & Ralston (1986) suggested that these deposits show some similarities to the Warrego Cu-Au deposit in the Tennant Creek-Davenport basins, Northern territory, Australia (Large, 1975, Goulevitch, 1975). Du Toit (in prep) suggested a possible link between these deposits and the nearby, undeformed granitic rocks of the Cnydas Subsuite which forms part of the Keimoes Suite. The Cnydas Subsuite has been differentiated from a calc-alkali parent magma and is characterised by the presence of Cu and Zn in its differentiated phases (Jankowitz, 1987). Jankowitz (1987) also pointed out the systematic depletion of Cu and Zn from the oldest to the youngest phases of the Cnydas Subsuite. Du Toit (in prep) speculated on the possibility of this Subsuite having only provided the heat source for the deposition of these ore bodies.

3.5.2.3 Fluorite-bearing vein deposits.

A few fluorite-bearing quartz veins occur some 20 km west of Upington in the region of the Dyasons Klip tungsten deposits (Fig. 4.9 & 4.14) within east-to-northeast-striking shear-zone-related veins. These veins are present on the farms Dyasons Klip 454, Geelkop 456, Blaauwskop 36 and Bloemsmond 455 and are hosted by the Dyasons Klip Gneiss, Riemvasmaak Gneiss and Toeslaan Formation. The veins are possibly related to the same granitic source as the tungsten mineralisation in the area.

Other small occurrences also occur near Kakamas and Riemvasmaak. Little attention has been paid to these deposits and their economic potential is considered to be inconsequential.

3.5.2.4 Amethyst-bearing vein deposits.

Only a few of these deposits occur in the Kakamas Terrane, the most notable of these is found on the farm Zoovoorby 458 within a northeast-trending fault-related vein, hosted by the Vaalputs Granite.

3.5.3 Deposits related to intrusives.

3.5.3.1 The Jacomynspan Cu-Ni deposit.

The Jacomynspan deposit is an approximately eastward-striking mafic-to-ultramafic dyke-like intrusive situated on the farms Jacomynspan 176 and Hartebeestpan 175, approximately 15 km west of Putsonderwater and has been described by Attridge (1986) (Fig.3.3). The 5 km long body is intrusive into the porphyroblastic quartz-feldspar-biotite-garnet gneiss of the Jacomynspan Group which has undergone amphibolite-grade regional metamorphism. The development of gossan as well as outcrop of the mineralised body is poor although gossanous fragments within pedogenic calcrete occurs, particularly in the vicinity of the mineralised zone (Tordiffe et al., 1989). The upper portion of the ore body comprises a quartz-feldspar-biotite-tremolite schist where the mineralisation consists of between 1% and 3% poorly disseminated chalcopyrite, pyrrhotite and pentlandite. Non-schistose hypersthene containing 10 to 20% disseminated sulphides occurs within the schist. The lower portion of the body consists of a poorly mineralised (< 1% sulphides) quartz-feldspar-biotite-amphibole gneiss which, according to Attridge (1986), represents a hybrid zone between the upper tremolite schist and the footwall Jacomynspan Group gneiss.

Ore reserve estimations on the portion of the ore body on the farm Jacomynspan indicate 114,18 Mt of sulphide and oxide ore grading at 0,17% Cu and 0,25% Ni.

The orientation of the sulphide mineralisation along foliation planes within the schist led Attridge (1986) to deduce that the mineralisation occurred before the final stages of regional metamorphism. Retrograde metamorphism resulted in the transformation

of the basic-to-ultrabasic assemblages to tremolite, actinolite and serpentine. The intrusive body becomes more ultramafic (serpentinitic) towards the east on the farm Hartebeestpan. The body probably represents a differentiated mafic to ultramafic body which intruded into an early fault or shear zone.

To the south of the Jacomynspan deposit, on the farm Rok Optel 261 (Fig. 3.3), several Cu-Ni deposits of similar character are present within the Jacomynspan Group. Grades of 0.3 % Cu and Ni and 1.2 g/t Ag are found in norite and feldspathic amphibolite as well as occasionally in pyroxenite. The intrusive body grades into a harzburgite nearer the base.

3.5.4 Pegmatites.

Pegmatites in the area under investigation form part of an extensive pegmatite belt of some 450 km in length and between 40 and 50 km wide. The pegmatite belt extends from the Vioolsdrift area, south of the Orange River in the west, eastwards towards the Augrabies area, where it swings southeast and disappears under Quaternary cover in the Kenhardt district, between Kenhardt and Prieska. The pegmatites have been described by numerous authors (Gevers et al., 1937; Von Backström, 1961; 1967; Hugo, 1969; 1986; Geringer & Botha, 1975; Stowe 1986). The most comprehensive study is provided by Hugo (1969). This belt broadly follows the regional trend of the host rocks. In the study area, the pegmatites are largely confined to the Kakamas Terrane, between the Hartbees River Thrust and the Bovenrugzeer-Cnydas Shear (Fig. 3.4). Rb-Sr model ages for muscovite and Pb-Pb ages from zircon indicate ages between 1000 and 950 Ma (Hugo, 1986) and indications are that the pegmatites are representative of the final stage of activity in the Kibaran-aged orogeny of the Namaqualand Metamorphic Complex.

The pegmatites vary in size and shape from thin veins of a few centimetres to bodies of 2 km in length, although the average size of the pegmatites in the study area is in

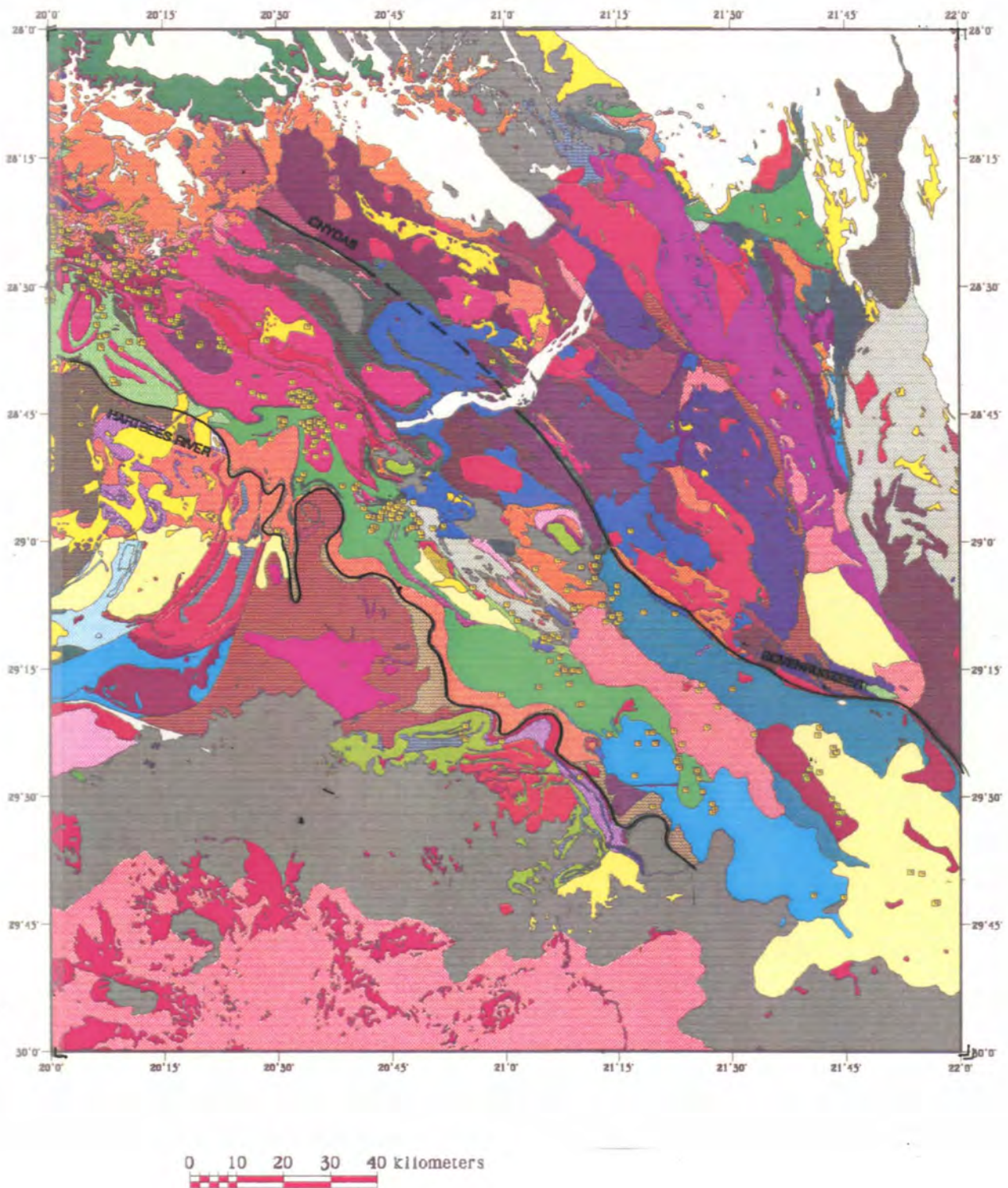


Figure 3.4: The distribution of all pegmatites in the area under investigation and their associated major faults.

the order of 200 m long and 60-100 m wide. Dyke and vein-like bodies are the smallest, while the larger bodies are irregular in form.

Hugo (1969) recognised unzoned and poorly zoned homogenous pegmatites as well as simple and complex inhomogeneous pegmatites. These are briefly summarised below:

Homogenous pegmatites:

Genesis is thought to be of a replacement origin for the following reasons:

- a) Contacts with host rocks are often gradational.
- b) Textures and structures of the host rocks are preserved in the outer parts of the pegmatites.
- c) Xenoliths are rare and where found, their foliation conforms to the foliation of the host rock.
- d) Distortion, bending, dragging and folding of wall rocks is rare.
- e) In the Kakamas area there are an excessive amount of pegmatites. No evidence has been found of such large scale structural readjustments that would be necessary for the host rocks to conform to such a large volume of intruded material.

Unzoned:

Unzoned pegmatites conform to the foliation of the host rocks and usually display a uniform mineralogy consisting of quartz, albite-oligoclase, perthite, biotite, schorl, magnetite, ilmenite and zircon.

Poorly zoned:

These pegmatites are similar in appearance to the unzoned pegmatites with the exception that they occur as isolated pods of coarse-grained quartz and perthite. They are of irregular shape with no obvious wall rock alteration and usually conform to the foliation of the host rocks. Additional minerals include beryl, apatite, schorl and garnet and are usually regarded as uneconomical.

Inhomogeneous pegmatites:

Inhomogeneous pegmatites are characterised by zones within the pegmatite which can be distinguished by mineralogy and texture. The size of the crystals within the pegmatites usually increases towards the centre or core. Inhomogeneous pegmatites are more regular in shape and display alteration of the wall rocks only where the composition of the host rock differs considerably from that of the pegmatite. Their genesis is thought to be as a result of emplacement of pegmatitic fluids into tension cracks and is based on the following evidence;

- a) Sharp, chilled contact and fine-grained border zone.
- b) In schistose host rocks, schists are crumpled, crinkled and distorted to conform to irregularities and space constraints created by the intruding pegmatite.
- c) Shapes of the pegmatites are governed by the shapes of the joints.
- d) Foliations of the xenoliths within the pegmatites do not conform to the host rock foliation.
- e) The compositions of the pegmatites are relatively consistent in different host rocks.

Simple:

These pegmatites show a well developed zoning of border, wall, occasionally intermediate and core zones and are usually found in amphibolites and granulites in the form of veins and lenses. Accessory minerals include biotite, rare earth element (REE) minerals, andalusite, apatite and magnetite. Andalusite, corundum \pm apatite and REE-bearing minerals (gadolinite, allanite, monazite, euxenite, fergusonite and xenotime) are often found in the outer periphery of the intermediate zone. REE-bearing minerals are found almost exclusively in the simple pegmatites and are usually attributed to magmatic differentiation (Jahns, 1955). Heinrich (1953) cautioned that the difference in REE composition in pegmatites could be more mineralogical rather than chemical, where the REE could be dispersed throughout the other minerals.

Complex:

Complex pegmatites are usually more irregular in shape than the simple types and are occasionally concordant with their host rock, although many appear to be emplaced along discordant fractures. The internal structure and composition of these pegmatites

are more complex and are composed chiefly of quartz, perthite, plagioclase (albite-cleavandite and albite-oligoclase) and muscovite. In addition, they have been subdivided into three groups:

- a) Beryl-bearing.
- b) Beryl-columbite-tantalite-bearing (\pm lithia mica and spodumene).
- c) Beryl-cassiterite-spodumene-bearing.

Furthermore, the complex pegmatites contain more tourmaline, apatite, fluorite, triplite, topaz and lithia mica than the other types of pegmatites, which indicate a richer concentration of hyper-fusibles such as B, F and P. These pegmatites are clustered in pegmatite groups (Cerny, 1982a), which could possibly indicate a concentration of hyper-fusibles during the late stages of pegmatite formation or, due to their lower viscosity and subsequent higher mobility, represent crystallisation of pegmatites at greater distances from the source.

3.5.4.1 Examples of specific complex pegmatites.

3.5.4.1.1 Middelpost No. 1 Pegmatite.

This complex pegmatite is situated some 25 km southeast of Kakamas on the farm Middelpost 60 and forms part of a group (Cerny, 1982a) of beryl-bearing pegmatites in the area (Hugo, 1969). The pegmatite is a northwest-striking body, approximately 275m long and 90m wide. Although the mine is at present dormant, over 180 tons of beryl was mined from this pegmatite. The pegmatite contains a thin border zone, wall zone, intermediate zone and core, as well as a replacement body.

3.5.4.1.2 Strausshiem No. 1 Pegmatite.

The Strausshiem No. 1 pegmatite (Hugo, 1969) is situated southeast of the Middelpost No. 1 pegmatite on the farm N'Rougas Noord 108. This complex pegmatite is part of a group of beryl-bearing pegmatites in the vicinity and is thought to be the first pegmatite to have been prospected in the Kenhardt District in 1900. The northwest-striking body is thought to be at least 450m long while the width varies between 15

Table 3.3: Characteristics of three selected complex pegmatites.

PEGMATITE AND (FARM) NAME	BORDER ZONE	WALL ZONE	INTERMEDIATE ZONE	CORE ZONE	REPLACEMENT ZONE
Middelpost No. 1 (Middelpost 60)	Intergrown plagioclase, quartz, perthite and muscovite.		Albite, cleavandite, quartz, muscovite with lesser greenish yellow to blue beryl and black tourmaline (schorl), 1-9m in width.	Coarsely crystalline milky and pink quartz with minor amounts of perthite. Irregular body of 1.5-35m in width and stretches over the entire length of the pegmatite.	Zoned body within the intermediate zone. Zones are from "core" outwards rose quartz - quartz, plagioclase, microcline, perthite - muscovite greisen - quartz, plagioclase - plagioclase, quartz, muscovite, beryl.
Straussheim No. 1 (N'Rougas Noord 108)	Quartz, albite-oligoclase, muscovite and schorl. Thin (up to 8cm) fine grained zone with sharp contact with schistose host rocks.		Muscovite zone: Muscovite, quartz, perthite, plagioclase, beryl, apatite, schorl, garnet. Schorl zone: Fine-to-medium-grained ground-mass of plagioclase, quartz, muscovite, and schorl. Large euhedral crystals of (up to 15cm in diameter) with c-axis perpendicular to zone contacts. Cassiterite-bearing zone: Cassiterite and schorl are disseminated throughout an albite and cleavandite groundmass. Other minor minerals include garnet, apatite and muscovite. Spodumene-bearing zone: Quartz, cleavandite, muscovite with beryl, triplite, apatite, spodumene and schorl. Spodumene is concentrated close to the core.	Milky quartz with euhedral crystal of perthite and large (up to 1.8m in diameter) schorl. Thickness of zone varies from 1-15m.	Muscovite greisen, quartz and cleavandite between the spodumene zone and core zone. Small clusters of blue, green and colourless lithium tourmaline.

Angelierspan No. 1 (Angelierspan 260)		Perthitic graphic granite with minor plagioclase and muscovite. Varies in thickness from 40cm-1m and impregnated by surface limestone.	Coarse-grained perthite, albite-oligoclase, quartz, muscovite and lesser beryl. C-axis is orientated perpendicular to the zone contacts.	Large subhedral white to pink perthite and quartz	
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From Hugo, 1969.

-75m. Beryl, cassiterite and mica was produced at various stages in the mine's history. This pegmatite is one of only a few Sn-bearing pegmatites within the portion of the pegmatite belt under investigation. All the other known Sn-bearing pegmatites are also part of the same group.

3.5.4.1.3 Angelierspan No. 1 Pegmatite.

The Angelierspan No. 1 pegmatite is situated east of Kenhardt on the farm Angelierspan 260 and is largely covered by Quaternary sediments. The pegmatite body extends for approximately 60m in a north-south direction. Approximately 90 tons of beryl was mined. A large mass of beryl, weighing 56 tons was found in the intermediate zone near the core.

Table 3.4 shows some of the production figures of selected pegmatites. These small scale mines were mined on an informal basis with no records being kept. These figures were obtained by Hugo (1969) from local prospectors at the time and should be treated with caution.

3.5.4.2 Vrede allanite.

These deposits have been described by Hugo (1961) and are situated on the farm Cnydas West 438, some 43 km northwest of Kakamas (Fig. 3.3). They occur as epigenetic lodes within the Riemvasmaak Gneiss and are conformable to the foliation of the host rock. The lode morphology is controlled by the foliation of the Riemvasmaak Gneiss where the stringers and veins pinch and swell. A 20 cm wide

Table 3.4. Production figures of some of the more important pegmatites.

PEGMATITE NAME	PRODUCTION YEARS	PEGMATITE GROUP*	COMMODITY
Japie	1956-1958	Riemvasmaak	7 ton REE (gadolinite)
Middel Post No. 1	? -1964	Middel Post	180 ton Beryl
Middel Post No. 2	? -1964	Middel Post	? ton Feldspar, 140 ton Beryl
Straussheim No. 1	1900- ?	N'Rougas	"Several hundred tons" mica, beryl, cassiterite
Straussheim No. 3	1952-1958	N'Rougas	32 ton beryl
Jack No. 2	1964-1965	Rok Optel	> 50 ton beryl
Angelierspan No. 1	?	Rok Optel	90 ton beryl
Angelierspan No. 2	?	Rok Optel	? ton spodumene, Li-mica, Columbite-tantalite, 27 ton beryl.

From Hugo (1969).

zone within the wall-rocks is usually defined by thin veinlets and stringers containing allanite, tourmaline and iron oxides which run parallel to the lodes and have a gradational contact with the host rocks.

The allanite-bearing lodes probably formed from volatile-rich pegmatitic fluids which penetrated and partially replaced the granitic gneiss along its foliation planes. The pegmatitic composition of the lodes and their spatial proximity to other allanite-bearing pegmatites indicate a possible genetic link with the pegmatite-generating process (Hugo, 1961). The mineralising fluids and vapours could either have originated from crystallising pegmatites in the immediate vicinity of these deposits or directly from a blind granitic pluton.

3.5.5 Deposits related to metamorphism.

A few wollastonite and sillimanite deposits occur in the region of Keimoes and Kakamas and are restricted to the Biesje Poort Group of the Korannaland Sequence. The most significant of these deposits is the Eksteenkuil deposit on the farm Eksteenkuil 35,7 km southeast of Keimoes, just south of the Orange River (Fig. 3.3). Wollastonite-rich zones are present within sequences of quartz-muscovite schist, marble, calc-silicate rocks, quartz-amphibole schist and quartz-biotite schist of the

Puntsit Formation, Biesje Poort Group, which represents a well bedded metamorphosed calcareous-psammitic sedimentary sequence. These wollastonite-bearing sequences can be followed for some 14 km along strike.

A few sillimanite deposits are located about 24 km north of Kakamas on the farm Biesje Poort 471 and are hosted by the quartzite, calc-silicate and feldspathic psammites and aluminous gneiss of the Sandputs Formation which belongs to the Biesje Poort Group.

3.6 THE BUSHMANLAND SUBPROVINCE.

3.6.1 Stratiform deposits.

3.6.1.1 The De Tuin Noord Ag-Pb-Cu-Zn deposit.

This low-grade deposit is situated in the De Banken Gneiss Formation of the Brakwater Metamorphic Suite on the farm De Tuin Noord 161 which is about 48 km northwest of Kenhardt. Surface expression is in the form of some minor gossan as well as malachite staining. The mineralisation occurs in the lower medium-grained biotite granite gneiss and calc-silicate of the De Banken Gneiss Formation. Field evidence suggests that this deposit probably extends sporadically westwards along strike. According to Du Toit (in prep.), the mineralisation could be associated with the northward-dipping Kalkputs thrust zone which lies about 4 km to the south of this deposit. He suggested that a mineralised quartzite band lying below the sulphide mineralisation could represent the "sole" of the Kalkputs thrust zone.

3.6.1.2 The Adjoining Geelvloer Pb-Zn-Cu-Au-Ag deposits.

A number of small subeconomic stratabound deposits are found on the farms Hartebeest Vlei 199, Graafwater 198 and Adjoining Geelvloer 197 near the western boundary of the study area. These deposits lie adjacent to the Geelvloer shear zone, which forms part of the Pofadder lineament (Fig. 4.1), and are characterised by their

dominance of Pb-Zn over Cu. The host rocks to the mineralisation are the volcano-sedimentary sequences of the Kraandraai Formation, Grappies Group which have undergone upper amphibolite-grade metamorphism. The Kraandraai Formation consists of marble, calc-silicate rocks, garnet-magnetite quartz rocks (a possible metamorphosed iron formation) and quartzite. The ore bodies consist of lenticular massive and disseminated sulphide horizons which strike sub-parallel to the tectonic fabric created by the easterly trending Geelvloer shear zone. The sulphide mineralogy includes argentiferous galena, sphalerite and chalcopyrite.

Further to the north on the farm Brulkolk 154, two very similar deposits associated with similar host rocks such as marble, iron formation, quartzite and calc-silicate rocks are present. The sulphide mineralogy consists of argentiferous galena, chalcopyrite, pyrite and sphalerite with accessory magnetite. Other very minor sulphide mineralisation in the area occurs in fracture zones within calc-silicate rocks.

3.6.2 Stratiform deposits.

3.6.2.1 **The Grootriet iron deposit.**

The iron deposit situated on the farm Grootriet 162, some 42 km west of Kenhardt (Fig. 3.3), is hosted by the biotite gneiss and calc-silicate rock of the De Banken Gneiss Formation (Brakwater Metamorphic Suite). This deposit represents the largest of numerous small iron deposits within the De Banken Formation. The deposit forms a prominent black koppie of magnetite/hematite ore and is up to 30 m thick. The coarse-grained ore contains interstitial gangue of quartz, epidote, feldspar, amphibole and pyroxene which makes up 30-50% of the ore by volume Pike (1962). Pike (1962) estimated the deposit to contain 92 000 tons of hematite and concluded that the magnetite contains too much titanium for economic exploitation.

Although there are no felsic intrusives exposed in the immediate vicinity, it is possible that the ore body is of a "skarn type" replacement origin within the calc-silicate rock.

3.6.3 Vein deposits.

3.6.3.1 **W-bearing vein deposits.**

3.6.3.1.1 The De Uitkyk Boven De Kalkgaten W deposit.

This wolframite deposit is situated 20 km southwest of Kenhardt on the farm De Uitkyk Boven De Kalkgaten 252 (Fig. 3.3) and has been described by Friggens (1981). Coarse-grained wolframite occurs in 1-40 cm wide quartz and pegmatite veins which parallel the foliation of the quartzo-feldspathic gneiss of the Kokerberg Formation (De Kruis Group) which hosts the veins. The veins are developed in a 50 m wide zone along the contact between the Kokerberg Formation and the intruded biotite gneiss of the syn to late-tectonic De Bakken Granite. In this vein zone, the individual veins are not laterally continuous, but peter out at random and continue at slightly different stratigraphic horizons. Malachite staining as well as molybdenite and scheelite are also present in the mineralised vein. The zone of veins continue for a strike length of about 500 m in a northerly direction. Assay values for WO_3 rarely exceed 400 ppm, but some samples with visible wolframite assayed 6,3% WO_3 and 0,27% Cu.

Friggens (1981) postulated that hydrothermal fluids were emplaced along local bedding plains, situated within folds peripheral to the De Bakken Granite. These structures are therefore potential targets for further exploration.

3.6.3.2 **Fluorite-bearing vein deposits.**

3.6.3.2.1 The Pypklip West F deposit.

Mineralised quartz veins are located along northeast-striking shear zones within biotite gneisses and amphibolites of the De Banken Formation which forms part of the Brakwater Metamorphic Complex (Fig. 3.3). The most important of these veins is situated on the farm Pypklip West 129, some 43 km northwest of Kenhardt where a large vein of 8 km in length is found. The fluorite is associated with calcite, galena and

pyrite. To the north of the Pypklip West deposit a fluorite-bearing vein of 4,8 km is found on the farm Witvlei 103. Numerous other quartz veins, some of which are fluorite-bearing, are present in the vicinity.

Hugo (1962) described the main deposits on Pypklip West and Witvlei, and concluded that the shear zones which host the mineralised veins underwent three tectonically active periods. The first period was characterised by small lateral movements accompanied by epidotisation. Milky quartz was then emplaced as veins and breccia cement in these zones during renewed movement. The third period of shearing resulted in further brecciation and the influx of hydrothermal solutions, which led to the deposition of quartz, feldspar and calcite with minor amounts of galena and pyrite in fissures, cracks and as cement within the quartz breccia. The relationship of the cross-cutting minerals suggests that the quartz and calcite was deposited first, followed by the fluorspar and then the sulphides. Hugo (1962) assumed a 20% fluorspar content and a vertical extent of 30 m and estimated the recoverable fluorite to be in the order of 500 kt.

Von Backström (1964) suggests that these and many other shear zones in the area have a generally uniform dip and strike, irrespective of the nature of the host rock and are much younger than the igneous rocks in the vicinity, probably reflecting a single set of stress conditions. These stress conditions could have resulted in the partial reheating of the rocks at depth, causing the mobile hydrothermal fluids to be transported along the fracture zones.

Post- Proterozoic

3.7 SURFICIAL LEACHEATE DEPOSITS.

3.7.1 The De Bakken Granite uranium.

The De Bakken Granite is a large east-west elongated batholith southwest of

Kenhardt which is covered by Karoo rocks in the south (Fig. 3.5). This coarse grained porphyritic quartz-feldspar biotite granite has a U/Pb age of 900 Ma (Linström, 1977). A few small uranium deposits are found within the outer periphery of the batholith in the form of carnotite-bearing accumulations of non-pedogenic calcretes mainly along the Hartbees and Sak Rivers.

Frick (1986) showed that uranium had been leached from the top 20-27 m of the granite and that the leaching was most intense above the lower limits of the seasonal fluctuations in the water table. The leaching took place irrespective of whether sand, granite debris or pedogenic calcrete formed on the granite surface. The uranium content in the granite below the water table is between 10 and 30 ppm whereas the content above the water table varies between < 1 to 5 ppm. According to Frick (1986), uranium leaching occurred during the Pliocene and was subsequently either deposited in the river calcretes or was lost into the Orange River. The deposition of uranium near the surface, due to vertical leaching, could only have formed at a shallow water table. A likely cause for the required shallow water table could have been glaciation during the Dwyka age, where the glaciers removed the upper weathered granites to expose fresh, unleached granites.

Leaching is still active, but according to Frick (1986), uranium movement appears to be horizontal rather than vertical, and is controlled by the depth of the water table. Uranium then accumulates near the rim of the pans on the De Bakken Granite at sites of ingress of groundwater.

3.7.2 The Geelvloer uranium-gypsum deposit.

The Geelvloer pan, which covers approximately 40% of the farm Geelvloer 196, is situated near the western extremity of the study area (Fig. 3.5). The basement rocks of the area include granitic gneisses, pegmatites and metasediments of the Bossiekom Formation (Grappies Group) as well as the syn to late-tectonic intrusive Vaalhoek Granites. The northern extremity of the pan forms the southern extension of the Sout

Table 3.5: The characteristics of post Proterozoic surficial deposits.

DEPOSIT NAME	COMMODITY	GENESIS	HOST ROCKS	BED ROCKS
De Bakken	Uranium	Leacheate	Syn to late-tectonic De Bakken Granite	Syn to late-tectonic De Bakken Granite
Geelvoer	Uranium, gypsum	Leacheate (U), evaporitic	Alluvial sediments, clay and aeolian sand in Geelvoer Pan	Syn to late-tectonic Vaalhoek Granite (U) and metasediments of the Bossieskom Formation
Aries	Gypsum	Evaporitic, sedimentary	Clay, mud in pan	Quartzite and shale of the Kuibis Formation
Rietput	Gypsum	Sedimentary, evaporitic	Sand, clay in Sout River	N/A
Wit Kalk Kolk	Gypsum	Evaporitic	Ecca shale, Dwyka tillite	Ecca shale and Dwyka tillite of the Karoo Sequence
Steenkampspuits	Salt, uranium	Evaporitic	Dwyka tillite	Dwyka tillite
Renosterkop	Diamonds(?)	Sedimentary	Alluvial gravel	N/A
Brulkolk	Uranium, gypsum	Sedimentary, evaporitic	Fluvial sand and gravel	N/A
McTaggart's Camp-Dyasons Klip	Uranium	Leacheate, sedimentary	Calcrete	Syn to late-tectonic Louisvale Granite
Kleinbegin	Uranium, gypsum	Sedimentary	Calcrete	Quartzites of the Dagbreek Formation
Arribees & Osvlei	Uranium	Sedimentary	Calcrete and sand	Gneisses and sand of the Droëboom Formation

River, but only during periods of heavy rainfall does the pan overflow into the Sout River. The pan floor consists mainly of alluvial sediments, clay and aeolian sand, cemented with gypsum and calcite while the surface silt and sand tend to be highly saline.

Uranium mineralisation, in the form of carnotite, is almost exclusively associated with the contact between the Vaalhoek granite and the overlying Dwyka tillite and is concentrated mainly within porous sand overlying fractures in the granites (Hambleton-Jones et al, 1968). Uranium mineralisation is rarely found in the pan itself. In places the uranium occurs in solution with the very shallow hypersaline groundwater. Concentrations of U_2O_3 reach up to 6 ppm in the north-eastern

extremities of the pan and where it overflows into the Sout River. A possible origin for the uranium could have been Vaalhoek Granite or the De Bakken Granite to the south where Frick (1986) (see Section 3.6.1) has shown leaching of uranium from the portion above the water table.

3.8 EVAPORITIC DEPOSITS.

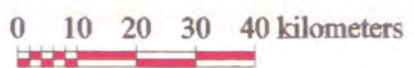
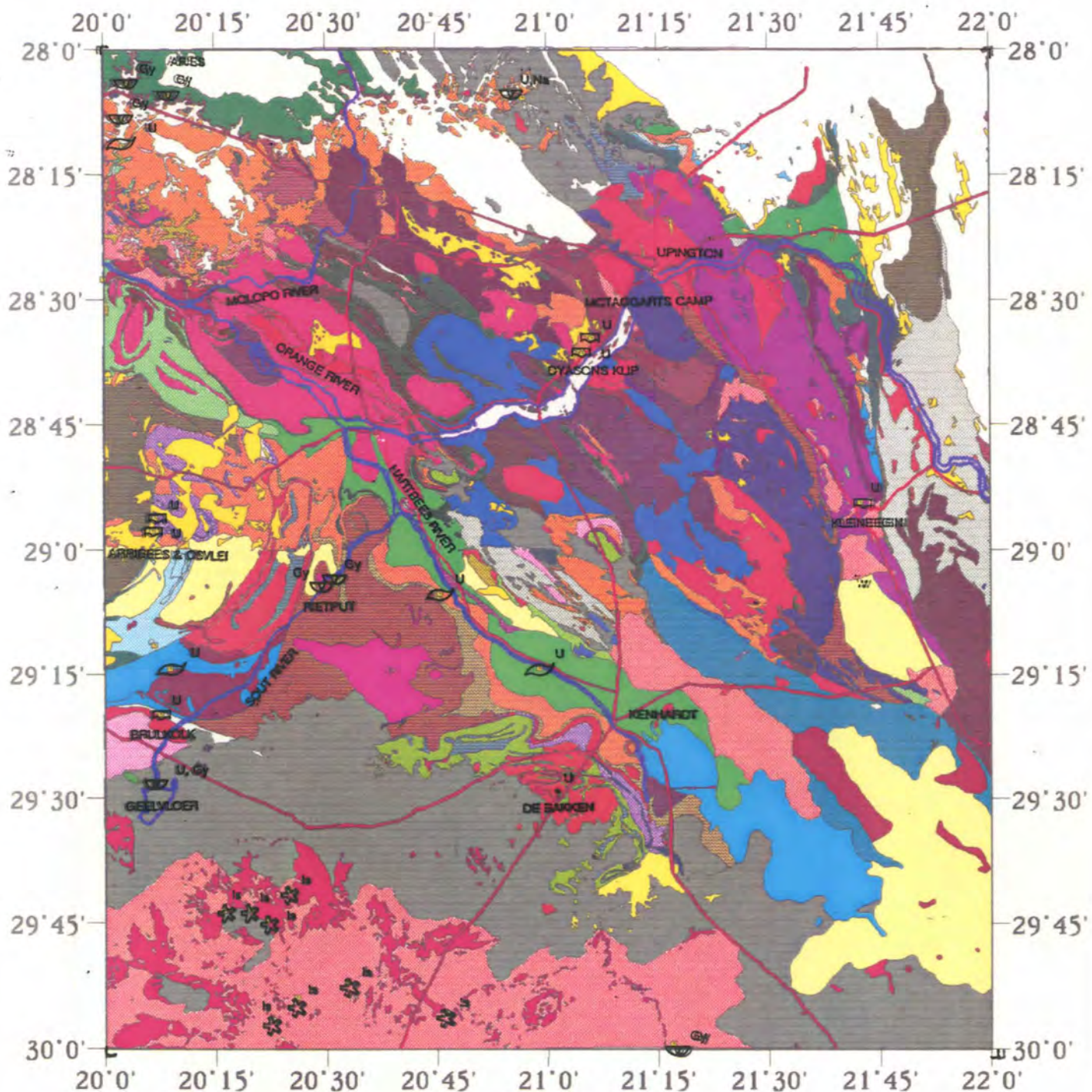
3.8.1 Gypsum.

3.8.1.1 **The Aries gypsum deposits.**

Three gypsum deposits are situated in the northwestern extremities of the area under investigation on the farms Aries 434, Gemsbok Hollow 358 and Zandvlei 433 (Fig. 3.5). The gypsum precipitated in pans which formed as a result of northwesterly trending sand dunes which inhibited the flow of the Bak River. The most important of these deposits is the one on Aries, south of the main road leading to Namibia from Upington. Here Visser et al. (1963) describe a typical profile of the pan consisting of an upper layer of dark brown clay and mud of between 0.45 to 0.6m. This is followed by a 0.6 to 1.5m thick layer of rather pure brownish gypsum which is underlain by a layer of hard, fine grained white gypsum which appears sporadically, but especially in the northeastern portion of the pan. This layer tapers out towards the centre of the pan and is underlain by a brown gypsum layer which is occasionally mixed with clay. The white gypsum contains about 80 to 85 per cent $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ while the brown variety contains 40 to 87 per cent $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ (Visser et al., 1963).

In the northern deposit, north of the main road, gypsum is exposed east of the pan up to 90m above the level of the pan and is partially covered by sand. The gypsum is a hard white type and has an average thickness of 1.2m. In the pan itself, where the ground water level is very shallow, the gypsum is of a brown variety but has limited economical potential.

According to Fockema et al. (1962) the source of the gypsum was probably from the



- Surficial Leacheate
- ∇ Evaporitic
- ∩ Sedimentary
- ⊞ Calcrete associated
- ☆ Vein

- Chief commodity :
- U- Uranium
 - Gy- Gypsum
 - Na- Salt
 - Is- Islandspar

Figure 3.5: The distribution of the more important post Proterozoic deposits in the area under investigation.

Nama shales which lie upstream from the deposits, and contain thin layers of white gypsum within shale horizons. Alternatively, Visser (1963) suggested that the gypsum originated from a more distant source and was precipitated from solution in the river where it was dammed up by the cross-cutting sand dunes. This theory is substantiated by the fact that the Bak River drains a portion of the Dwyka Formation which provides a source for the gypsum at other deposits in the Kalahari.

3.8.1.2 The Rietput gypsum deposits.

On the farm Rietput 98 (Fig. 3.5), gypsum has been deposited on the banks of the Sout River and occurs in the form of selenite crystals as well as gypsum powder. A hard compact variety occurs with sand and clay near the surface. Selenite crystals, as well as some gypsum powder, occurs in a 30-90 cm thick layer of clay and sandy clay. The gypsum content in the clay layer is highly variable and seldom reaches more than 50% (Visser et al., 1963). Gypsum in solution could have been transported from the south along the Sout River where it drains the Geelvloer Pan which is known to contain gypsum (see Section 3.6.2). This would lead to the potential of further gypsum deposits along the Sout River, which have not been found. One would also expect uranium to have been deposited.

3.8.1.3 The Wit Kalk Kolk gypsum deposits.

Numerous large gypsum deposits occur south of the study area in the Brandvlei region. Examples of these are situated near Verneuk Pan and Grootvloer Pan. All these deposits are associated with the Ecca shales and Dwyka tillite of the Karoo Sequence in flat lying, poorly drained areas. The only such occurrence known in the area under investigation is found on the farm Wit Kalk Kolk 318 (Fig. 3.5) in the southern extremity of the study area. It is possible that many of these deposits may be linked to form a large deposit covering a vast area.

Fockema et al. (1962) suggested that the gypsum was formed by the reaction between sulphuric acid and calcium carbonate. The sulphuric acid is thought to have been

derived from the oxidation of pyrite which is present in the Dwyka shale as well as the numerous dolerite sills present in the area. The calcium carbonate is readily available in most of the rocks in the area in varying quantities. Fockema et al. (1962) suggested that these gypsum-bearing fluids then washed into the pans where they soaked into the ground water. Subsequent evaporation caused these solutions to be drawn up by capillary action and to precipitate a layer of gypsum near the surface.

3.8.2 Salt.

Some 50 km northwest of Upington on the farm Steenkampspuits 365 (Fig. 3.5) lies a subeconomic saltpan which drains the surrounding hills although it is predominantly fed by the Doringdam River. The surface of the pan consists of a thin layer of Dwyka tillite which overlies arenites of the Kuibis Formation of the Nama Group. The area is also intruded by numerous Koras-aged Rooiputs granophyres. This deposit also contains uranium. More economical salt deposits occur further towards the north, outside the study area.

3.9 SEDIMENTARY DEPOSITS.

3.9.1 Diamonds.

At Renosterkop, approximately halfway between Kakamas and Augrabies Falls, there are a few old diamond diggings within the Tertiary gravels of the Orange River. These gravels occur just south of the Orange River, but it is unknown how successful these prospects were. Generally, the alluvial diamonds along the Orange River east of Vioolsdrift, are considered uneconomical as the older Tertiary gravel beds that host the diamonds west of Vioolsdrift are not present, or are covered by younger alluvial overburden.

3.9.2 The Brulkolk uranium-gypsum deposit.

This deposit lies in the western portion of the study area within the Brulkolk se Holte

drainage channel of the Sout River system on the farm Brulkolk 154 which is situated about 100 km west of Kenhardt (Fig. 3.5). According to Hambleton-Jones et al. (1986) the uranium was deposited during the mid Tertiary period when the Sout River had a shallow valley and the alluvium rarely attained a depth of more than a few metres. The river was subsequently filled with alluvium and cemented with calcite, gypsum and locally with fluorite. Uranium mineralisation formed by the precipitation of carnotite in red fluvial sand and gravel which is restricted to narrow 0,1-2 m thick patches some 0,75 m below the present surface. The red sand and gravel beds contain U_2O_3 concentrations of >50 ppm. Some of the uraniferous bodies appear to have been eroded away in recent times. The origin of the uranium is a matter of debate, but the De Bakken Granite could have been a possible source as the Sout River drains the north-western portion of the pluton and is known to contain leached uranium (see Section 3.6.1).

3.10 VEIN DEPOSITS.

3.10.1 Iceland Spar.

Numerous calcite deposits, commonly of optical quality iceland spar, occur mainly in the shales of the Prince Albert Formation of the Ecca Group, in the southern portion of the quadrangle under investigation (Fig. 3.5). These deposits are found on the farms Koranna Kolk 284, De Paarden Vleyen 283, Dachab 282, Zoo Afs Kolk 326, Karee Bosch Kolk 329, Gemsbok Rivier 301 and Klein Lemoenkop West 322. The very coarsely crystalline (up to 30 cm) veins of almost pure calcite are hosted by the basal chill zones of intrusive Karoo dolerites and display an irregular pod-like morphology. The source of the $CaCO_3$ is probably from dolomitic limestone present in the Dwyka and Prince Albert Formations, which was remobilised by the intruding dolerite.

3.11 DEPOSITS ASSOCIATED WITH CALCRETE.

3.11.1 The McTaggart's Camp - Dyasons Klip uranium deposits.

The few uranium deposits that occur in this region all seem to be of low economic significance. The most important of these are the occurrences on the farms Dyasons Klip 454 and McTaggart's Camp 453 (Fig. 3.5). The deposit on McTaggart's Camp lies along the Helbrandskloofspruit which flows over the Louisvale Granite. The Louisvale Granite has a gneissic texture and is capped by patches of thin nodular calcrete with rounded pebbles of quartz and older calcrete, within a fine-grained calcareous-cemented, sandy matrix. The uranium anomaly extends for approximately 1 km on either side of the stream. According to Treasure (1977), the mineralisation is present within the calcrete and along its contact with the underlying gneiss. He also noted small patches of yellow secondary mineralisation developed within the calcrete matrix and fractures present in quartz pebbles. Green and yellow secondary mineralisation is furthermore developed along weathered fragments of the Louisvale Granite as well as along its contact with the calcrete. Treasure (1977) interpreted this phenomenon as being related to the groundwater migrating more freely through the upper weathered surface of the Louisvale Granite. In and around a small pegmatite vein at the north-western extremity of the anomalous area, Treasure (1977) found patches of yellow and green secondary uranium minerals. The origin of these uranium minerals could either be synchronous with those in the calcrete, or be a primary constituent of the pegmatite itself (Treasure, 1977).

The second anomaly, as described by Treasure (1977), is situated on the farm Dyasons Klip 454, and extends in a northwest-southeast direction for about 3-4 km. This anomaly, is similar to the McTaggart's Camp occurrence in that it generally conforms to the drainage patterns in the area and the mineralisation is present within calcrete. Patches of calcrete cover overlie Dyasons Klip Gneiss and are in turn partially concealed by surface sediment. The mineralisation is, however, not present in weathered gneissic bedrock and the concentration of uranium minerals seems to be more erratic than the occurrence at McTaggart's Camp.

3.11.2 The Kleinbegin uranium deposit.

On the farm Klein Begin 115 in the eastern portion of the map area (Fig. 3.5), uranium is located within one of the local periodical streams. A flat peneplain characterises the physiography of the area and much of the Dagbreek Formation is covered by Quaternary sand. Uranium mineralisation, in the form of secondary carnotite, is present within gypsum, which is suspected to have acted as a geochemical interface, causing precipitation of the uranium. The primary origin of the uranium is unknown, but it is likely to have been derived from the surrounding Klein Begin Granite.

3.11.3 The Arribees & Osvlei uranium deposit.

Two uranium occurrences of lesser importance are located in the western portion of the study area (Fig. 3.5) on the farms Droëboom 72 and Arribees & Osvlei 95. The region is characterised by a flat alluvial plain which is predominantly covered by red sand and calcrete. Gneisses and quartzites of the Droëboom Group outcrop to the east of the uranium anomalies. According to Treasure (1977), the mineralisation is patchy and is developed both within the calcrete matrix and concentrated around rock fragments (mainly quartz) in the calcrete. The depth to which the calcrete is developed is uncertain, but it appears to be confined to within a metre of the surface sediments Watkins (1976).

Mineralisation occurs in calcrete away from the stream bed and there is no obvious indication of any mineralisation upstream of these anomalies. This led Watkins (1976) to conclude that the mineralisation is a consequence of the stream having eroded into or through a zone of uranium mineralisation. Treasure (1977) speculated that, due to the obvious relationships between the calcrete and the uranium, the deposition of carbonates and uranium occurred concurrently from the circulating groundwaters. The source of the uranium is unknown, but it could have been derived from pegmatites located in the area.

Chapter Four.

4. METALLOGENIC ASSESSMENT.

4.1 INTRODUCTION.

The previous chapter provided an introduction to the different styles of mineralisation within the various tectonostratigraphic units. The purpose of this chapter is to provide an analysis of the metallogeny within the area under investigation. The various styles of mineralisation are generally, but not entirely, restricted to the different tectonostratigraphic units but are more dependant on the age of mineralisation. In this chapter, the metallogenesis is analysed according to styles of mineralisation and are arranged, as far as possible, in chronological order. Many sections start with a summary of the characteristics of the styles of mineralisation relevant to the section. This is done in order to relate and compare these accepted models with the mineralisation in the study area so that a metallogenic analysis of these deposits can be effectively accomplished.

There are three broad ages of economic mineralisation, namely, ages associated with the pre-tectonic events, those related to syn to late-tectonic events and those related to post-tectonic events. Here "tectonic event" refers to the main Kibaran-aged orogeny.

4.2 SEDIMENTARY EXHALATIVE MASSIVE SULPHIDE DEPOSITS.

4.2.1 Introduction and overview.

The general characteristics of sediment-hosted, submarine exhalative Pb-Zn (SEDEX) deposits have been proposed by Large (1981) and are summarised below in Table 4.1.

Table 4.1 Characteristics of typical SEDEX deposits.

They usually comprise one or more lensoidal bodies up to a few tens of metres thick and between a few hundred to a few thousand metres long.
Individual bodies may form within a stratigraphic unit of up to a thousand metres but do not tend to form clusters as is often the case with VMS deposits (Section 4.3.1).
Proximal SEDEX deposits are commonly associated with stockwork or vein type mineralisation adjacent or subjacent to the stratiform mineralisation and this is seen as the feeder pipe form the mineralising fluids.
These deposits are usually characterised by a simple ore mineralogy consisting of pyrite and/or pyrrhotite, sphalerite, galena and minor chalcopyrite while marcasite and arsenopyrite are occasionally present. The Fe-sulphides are commonly the most abundant, while Ag may be present in the form of freibergite or within galena.
Barite is often associated with SEDEX deposits and is usually positioned above the stratiform sulphide mineralisation or occurs as lateral extensions of the stratiform sulphide mineralisation.
Silicification is often present in the footwall sediments of the stratiform sulphides, particularly near the stockwork mineralisation.
These ore bodies commonly display vertical and/or lateral zonation of the sulphides. From the core outwards, the lateral zonation is typically Cu- Pb- Zn- (Ba) while the vertical zonation from the core outwards displays Cu- Zn- Pb- (Ba) zonation. Pyrite and/or pyrrhotite is usually distributed throughout the ore body, sometimes forming a Fe- rich core while hematite may form peripheral to the stratiform sulphide mineralisation.
The stratiform sulphides are usually interbedded with a variety of marine sediments indicative of either a quiet, low energy environment or high energy turbidity currents and slumping.
SEDEX deposits are often associated with thin, fine grained tuffite horizons within the host lithologies, and are usually of a distal, unknown origin.
Many of these deposits are spatially associated with faulting, which was active during the mineralising episode and are generally considered to have been the conduits for the metal-bearing fluids. Subsequent reactivation of these faults may be manifested in major crustal lineaments.
Deposits within a single tectonic province are usually of similar age, but do not necessarily form at exactly the same time.
These deposits are commonly located in miogeosynclinal (passive continental margin or oceanic crustal margin) or basinal (intracontinental) pre-orogenic regimes. Deposits formed in the intracontinental environments are usually not as severely deformed as those formed in the miogeosynclinal settings during post-depositional deformation.
Pb-isotope ratios for galena are homogeneous and non- radiogenic.
S-isotope values from the metal sulphide are usually in isotopic equilibrium, while the S in some of the pyrite and barite becomes isotopically heavier and is considered to be biogenically reduced from sea-water sulphate at the site of mineralisation.

Summarised from Large (1981).

Large (1981) also pointed out that these deposits are often located close to block faults in third-order basins. These fault zones are considered to have been active during the deposition of the host sediments and may become reactivated to form lineaments along which several sediment-hosted Pb-Zn deposits are located.

4.2.2 The Bushmanland Subprovince.

The stratiform Pb-Zn deposits associated with the volcano-sedimentary sequences of the Bushmanland Subprovince within the area under investigation, have not been

studied in much detail and no literature is available on these deposits. The metamorphism and deformation of the ore bodies and their host rocks make it difficult to interpret the genesis of these deposits. The present lack of information makes interpretation of these deposits highly speculative at this stage. Their association with predominantly metasedimentary sequences and their Pb-Zn dominance over Cu leads one to postulate a sedimentary exhalative origin for these deposits. These ore bodies have a relatively simple sulphide mineralogy, consisting of essentially argentiferous galena, sphalerite and chalcopyrite. There are, however, some doubts as to the proto-lithology of some of the amphibolites and leucocratic gneisses in the area and whether they are metasedimentary or metavolcanic. Furthermore, there is an absence of significant Fe and/or Mn oxide formations associated with these deposits as well as a total absence of known barite, stockwork or vein-type mineralisation in the proximity of the ore bodies, or any lateral and/or vertical zonation of the sulphides. This may be explained either by these attributes having been removed through shearing, as all the deposits are located in a highly sheared terrain, associated with the Geelvloer Shear, or that these deposits are not of an exhalative origin.

Agenbacht (pers. comm.), although highly speculative, suggested a possible link between these deposits and the 1200 Ma (Joubert, 1986a) T'Oubep intrusive Suite (Paizes, 1975; Harris, 1985; MacLaren, 1988; Agenbacht, 1992) and that these ore bodies may represent a skarn deposit. The T'Oubep Suite has been interpreted by Agenbacht (1992) as being dominantly calc-alkaline with an I-type affinity.

The location of the Adjoining Geelvloer occurrences in relation to the Putsberg Cu (Pb-Zn) ore bodies, southeast of Pofadder, and the Gamsberg and Aggeneys Zn-Pb (Cu-Ag) deposits is, however, noticeable (Fig. 4.1). There is little doubt about the sedimentary exhalative origin of the Aggeneys-Gamsberg deposits (Ryan et al., 1986). The Putsberg ore body has been interpreted by Viljoen et al. (1986) as being of sedimentary exhalative origin. If an extension of the regional fabric of the Bushmanland Subprovince is made from Aggeneys eastwards, then a connection may be drawn through the Putsberg ore body eastwards to the Adjoining Geelvloer

deposits (Fig. 4.1). According to Large (1981) (see Table 4.1) SEDEX deposits are often associated with third order basins. Moore (1980) suggested that the Namaqualand-Bushmanland paragneiss sequences associated with the Aggeneys deposits represents a single elongate east-west-trending basin. It is therefore possible that these deposits may represent a row of third-order basins within a single large basin extending from Aggeneys through to Adjoining Geelvloer. The host rocks at Aggeneys comprise acid leucogneisses at the base and metasedimentary quartzite-aluminous schist and iron formation in the upper portions. Towards the east, in the Grappies Group, which hosts the Adjoining Geelvloer group of deposits, the rocks consist of more mixed successions of marble, calc-silicates, garnet-magnetite-quartz rocks and quartzite. The host sequences of the Aggeneys and Adjoining Geelvloer areas bear certain similarities (Thomas et al., 1994). Joubert (1986) implied a continuation of the quartzite/aluminous schists following the regional tectonic fabric from the Aggeneys area through the Pofadder region into the area under investigation in the Adjoining Geelvloer area. Tentative stratigraphic correlation along this "basin", therefore, seems to substantiate the hypothesis of these deposits having been connected to a similar genetic origin.

The age of the Adjoining Geelvloer deposits and those of the Putsberg deposits have not yet been determined, but Köppel (1980) and Welke & Smith (1984) gave model Pb ages of between 1350 and 1300 Ma for the Aggeneys deposits. Reid et al. (1987), however, determined a Sm-Nd isochron age of 1650 Ma for the Aggeneys deposits.

The origin of the somewhat younger (1200-1150 Ma; Köppel, 1980) Rozynbosch Pb-Ag (Zn-Cu) deposit is even more enigmatic. This deposit lies within a fault-bounded crustal fragment, just east of the basal thrust within the Hartbees River Thrust Belt within the Gordonia Subprovince. The Hartbees River Thrust is taken at the base of what Harris (1992) termed the Hartbees River Thrust Belt which is a wide zone of thrusting characterised by numerous mylonite zones. The Rozynbosch deposit, therefore, lies within this zone of thrusting. The Rozynbosch deposit lies along the "line" of sedimentary exhalative deposits discussed earlier, but does not lie along the continuation of the regional fabric which curves northwards in the north-

eastern portion of the Bushmanland Subprovince (Fig. 4.1). Whether the Rozynebosch deposit is related to the other occurrences in the Bushmanland Subprovince and has just been repositioned into its present position, or whether the Rozynebosch deposit is an autonomous unit within a specific crustal block is unclear. Harris (1992) suggested that the change in fabric trend from east-west to northerly is related to the Pofadder lineament and that the removal of this late strain distortion would line the Rozynebosch deposit up with the Aggeneys-Geelvloer trend. Köppel (1980) indicated that the Pb includes a significant mantle component and Thomas et al. (1994) speculated on a replacement origin for this deposit within the leucogneissic host.

4.3 VOLCANOGENIC MASSIVE SULPHIDE DEPOSITS.

4.3.1 Introduction and overview.

Volcanogenic massive sulphides (VMS) are formed as a result of a specialised type of hydrothermal system that is developed within a submarine volcanic environment. VMS deposits commonly form clusters and are separated from each other by rocks of similar lithologies. The average area occupied by a typical cluster of VMS deposits is in the region of 850 km², which is equal to a circular area of approximately 32 km in diameter (Sangster, 1980). Sangster (1980) furthermore stated that a cluster often contained an average of 12 deposits (usually between 4-20 deposits) amounting to 92 Mt of ore. The largest deposit in the cluster contains an average of 67% of the total metal, and the second largest in the order of 13%. Within these clusters, the different deposits are often confined to a relatively narrow stratigraphic interval and are seemingly related to structural controls, most noticeably vertical faults. These faults then provide channelways for hydrothermal fluids which result in the formation of stockwork systems at the base of the exhaled fluids where the precipitation of base metal sulphide lenses occur. The upper contact of the sulphide lenses with the wall rocks are usually sharp, while the basal contact is gradational. The essential characteristics of a VMS deposit are shown in Figure 4.2.

4.3.1.1 Mineralogy and zonation.

The common sulphide mineral in massive sulphide lenses is pyrite with subordinate pyrrhotite, chalcopyrite, sphalerite and galena, while non-sulphide minerals include magnetite, hematite, cassiterite as well as quartz, chlorite, barite, gypsum and carbonates. The most characteristic feature of VMS deposits is the chemical, mineralogical and textural zonation of the ores as well as the hydrothermal alteration of the basal feeder pipe (Fig. 4.2). There is usually a decrease in the Cu/Zn ratio away from the feeder pipe reflecting a decrease in chalcopyrite and a simultaneous increase in sphalerite and galena. This outward zonation is as a result of the solubility

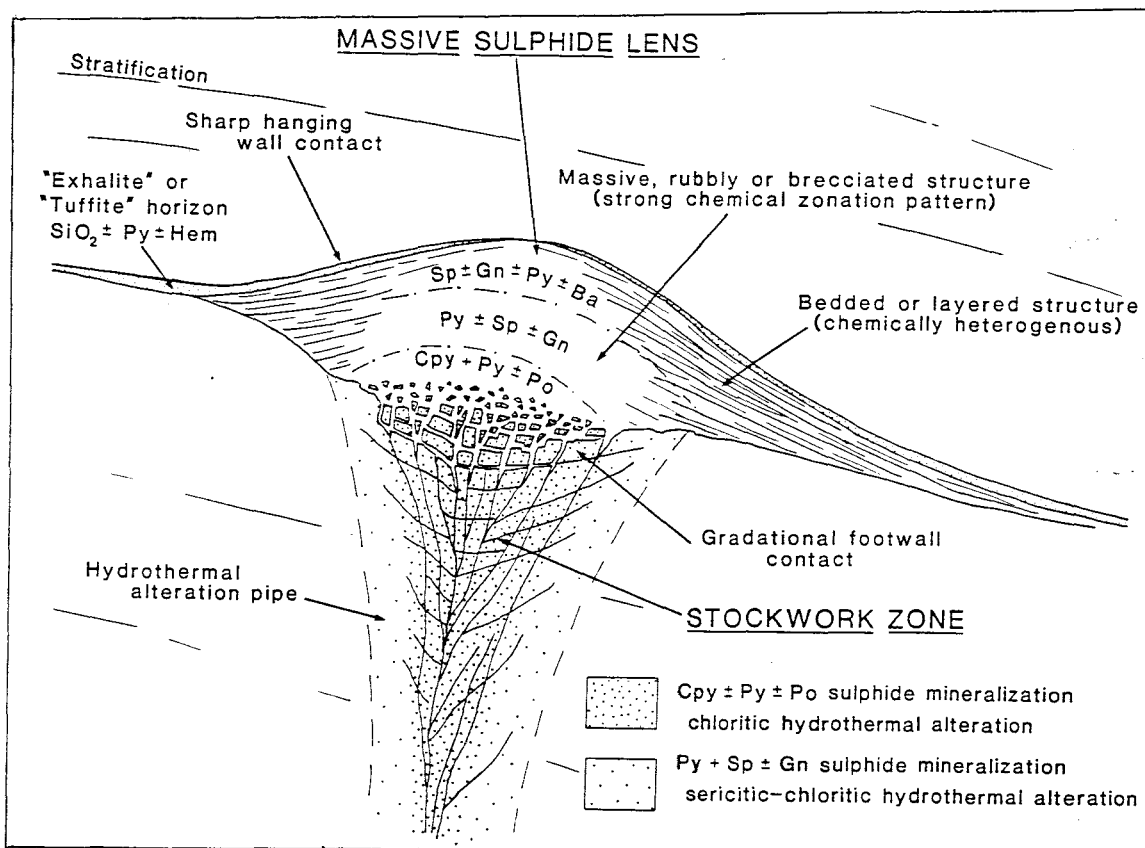


Figure 4.2: Essential characteristics of an idealised VMS deposit (from Lydon, 1988a).

of chalcopyrite in a chloride-bearing aqueous solution being highly temperature dependant (Lydon, 1988b). The solubility of sphalerite (and galena) in chloride-bearing solutions is much less temperature dependant which leads to a significant concentration of Zn and Pb in the lower temperature regimes, away from the exhalative feeder pipe, under comparable physiochemical conditions. Pyrite is generally ubiquitous while pyrrhotite, magnetite and bornite are concentrated in the feeder pipe and central portion of the sulphide lenses. Barite is usually present in the peripheral parts of the massive sulphide lenses where sea-water sulphate is involved in its precipitation (Watanabe & Sakai, 1983; Kowalik et al., 1981). There is often a spatial relationship between the VMS deposits and stratigraphically overlying units of either magnetite-hematite iron formations or manganese-oxide formations. Different hypotheses for this phenomenon have been presented. Barnes (1983) suggested that these sediments are a reflection of marginal or

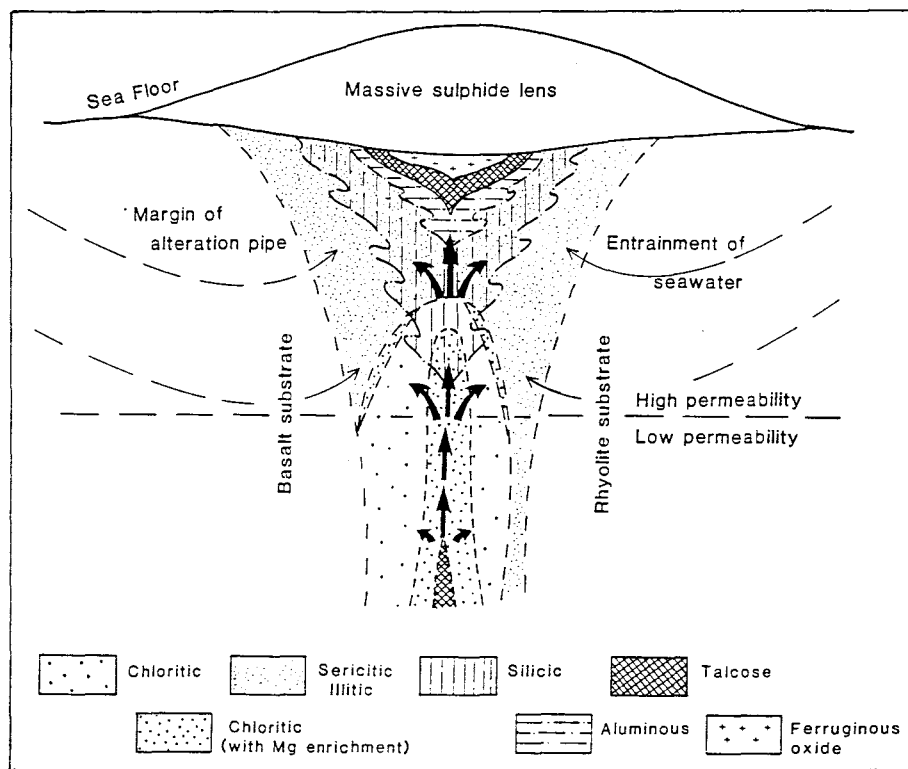


Figure 4.3: Schematic representation of the various alteration assemblages associated with feeder pipes to VMS deposits (from Lydon, 1988a).

terminal oxidation of brine pools, while Large (1977) suggested that they formed as a result of the progressive oxidation of the hydrothermal plumes away from the vent, and Lydon (1988b) proposed that these sediments are "the product of contemporaneous or later low-temperature hydrothermal discharge". The primary zonation could be distorted by mobilisation of the sulphides during deformation and metamorphism and could even result in the development of stockwork-like veins (van Staal & Williams, 1984).

The feeder pipes in the footwall of the VMS deposits (if the deposit is proximal) are also subject to hydrothermal alteration. This is shown schematically in Figure 4.3. Generally, however, the conduits for the hydrothermal ore fluids are characterised by chloritic cores and sericitic margins (Lydon, 1988b). This phenomenon has been interpreted by Riverin & Hodgson (1980) and Lydon & Galley (1986) to represent the decreasing thermal gradient which surrounds the feeder pipe.

Plimer (1978) recognised variations, not only with respect to zonations within a single ore body, but also with reference to the ore bodies within a cluster. He related these variations to the distance of the ore bodies, in time and space, from the volcanic centre and called them proximal and distal. The main characteristics of both proximal and distal deposits are given in Table 4.2. Plimer (1978) cautioned that there are numerous stratiform base metal deposits which display both proximal and distal characteristics. In this model he considered time and space to be the same which in effect also means that the younger deposits would be proximal and the older deposits distal. The typical Fe- and Mn-oxide formations in stratigraphic units above the massive sulphide zones, would, according to Plimer (1978), represent a distal portion of the deposit. It is therefore possible to have a continuum from proximal to distal deposits reflected within a cluster of massive sulphide deposits. If, however, various vents occur within a cluster area, which is highly likely, then the common zoning of metals would not be as obvious.

Table 4.2. Characteristics of proximal and distal stratabound ore deposits.

PROXIMAL DEPOSITS	DISTAL DEPOSITS
Associated with lava, agglomerate, tuffs, minor reworked pyroclastics, chemical sediments.	The more distal, the greater the percentage of clastic and chemical sediments (chert, limestone, dolomite, barite), relative to volcanics (tuffs), make up the host rocks. Some fossiliferous units.
Major facies change or hiatus stratigraphically above the sulphide mineralisation.	Rapid facies changes or hiatus appear to post-date ore deposition.
Spatially associated with alteration pipes, stringer ore zones.	No spatial affiliation with alteration pipes or stringer ore zones. Only slight hydrothermal alteration characterised by silicification, sericitisation and Fe-metasomatism
Common metal zoning from Cu-rich footwall alteration to overlying massive pyrite and chalcopyrite + sphalerite and uppermost pyrite + chalcopyrite + sphalerite ± galena.	Pyrite, pyrrhotite, sphalerite, chalcopyrite and galena constitute the sulphides. Metal zoning from Cu- to Zn- to Pb (± Ag ± Ba ± Mn ± F)-rich ores from the base to the top of the sulphide zone is common.
Sedimentary structures (cross bedding, slump structures) often visible.	

Summarised from Plimer (1978)

4.3.1.2 Classification.

Lydon (1988a) classified VMS deposits according to major ore metal associations. He distinguished two types, namely, the Zn-Pb-Cu-type and the Cu-Zn-type. The Zn-Pb-Cu-type contains a significant amount of barite in the outer periphery of the ore body as opposed to its virtual absence in the Cu-Zn-type. The Au/Ag ratio of the Cu-Zn-type is usually higher than that of the Zn-Pb-Cu-type. The sulphur isotope ratio of the Cu-Zn-type is also significantly lighter than that for the Zn-Pb-Cu-type. The oxide facies Fe- and/or Mn- formations are more commonly associated with the Zn-Pb-Cu-type (although not a ubiquitous feature) but are not necessarily absent in the Cu-Zn-types (Franklin et al., 1981). The presence of felsic and/or sedimentary rocks are more prevalent in association with Zn-Pb-Cu-types whereas mafic volcanic rocks are dominant in Cu-Zn deposits (Franklin et al., 1981). They also noted that subvolcanic intrusions are commonly found associated with the Cu-Zn-type districts, although the Cu-Zn type in the more sedimentary dominated areas such as the Besshi deposits of Japan, do not appear to have prominent subvolcanic intrusives. In the Zn-Pb-Cu districts, however, subvolcanic intrusives are less prominent. There does not seem to be any significant differences in morphology between these two types of deposits. Franklin et al. (1981) noted some differences in the characteristics of the alteration

pipes of the two types of deposits that are summarised in Table 4.3.

Table 4.3. Alteration characteristics of the feeder pipes of the Cu-Zn and Zn-Pb-Cu deposits.

Cu-Zn DEPOSITS	Zn-Pb-Cu DEPOSITS
Well define, often extensive alteration.	Alteration pipes not vertically extensive, but are mineralogically well defined.
Mg-rich chlorite or talc-rich core surrounded by a sericite \pm quartz-rich halo.	Sericite-quartz core surrounded by a halo of Mg-enriched chlorite.
Pervasive Na ₂ O and CaO depletion, some K ₂ O addition. The Mg-rich core is depleted in SiO ₂ and the outer sericitic zone may be enriched in SiO ₂ .	

Summarised from Franklin et al. (1981).

Several authors (for example Hutchinson, 1973; Solomon, 1976; Ohmoto et al., 1983) suggest that the metal ratio Cu:Zn:Pb is partially a reflection of the trace metal composition of the source rocks. The absence of Pb in the Cu-Zn type VMS deposits is ascribed to the source fluids originating from mafic rocks and the abundance of Pb in the Zn-Pb-Cu-type VMS deposit is attributed to a felsic volcanic and/or sedimentary origin.

Hutchinson (1973) and Solomon (1976) also classified VMS deposits according to the major ore element composition but added a third Cu-only type. Hutchinson (1980), in addition to classifying these deposits into major element associations, also attempted to group these classes into tectonic environments in which they were formed. Apart from the Cu-only type, which was related to plate rifting, most other classes are generally related to various stages of plate collision and subduction and the accompanied variation of volcanic composition. From the above it can be seen that the composition of the ore body is at least partly related to the composition of the host rock. It has therefore been suggested by Franklin et al. (1981) that the type of massive sulphide deposit is not directly related to tectonism but rather to the composition of the host rock, which is certainly subject to tectonic control.

Sawkins (1976) identified three types of VMS deposits, namely the Kuroko-type which occurs in felsic, calc-alkaline volcanic sequences of any age in areas of plate convergence, the Cyprus-type which occurs in low-potassium basaltic rocks in the

upper portions of ophiolite complexes at areas of plate divergence, and the Besshi-type which occurs in clastic sediments and mafic volcanics in troughs of no specifically defined plate tectonic setting.

4.3.1.3 Origin of metals.

Ignoring for the moment the tectonic settings in which VMS deposits were formed, Lydon (1988b) summarised three types of hydrothermal systems, proposed by various authors for the generation of these deposits. The most popular contemporary model is that of the convective cell where a magmatic heat source causes subsurface waters, dominantly of seawater origin, to move through the substrata, leaching the ore components from the rocks along their flow paths. Lowell & Rona (1985), among others, doubt the capacity of most intruding magmas to provide enough heat to produce some of the larger massive sulphide deposits. Another possibility, although not widely acclaimed, is one where the ore fluids are derived from volatiles of magmas, the so-called "magmatic hydrothermal model". Sangster (1972) and Solomon (1976) for example, suggested a highly differentiated calc-alkaline magma to be a possible source for some VMS deposits. The model favoured by Lydon (1988b) is what he termed the "stratal aquifer model", where deeply buried connate water under extreme lithostatic pressure could be released along fracture zones, possibly as a result of tectonic activity. He used this theory to substantiate the occurrence of VMS deposits in clusters within a relatively narrow stratigraphic horizon, as the release of fluids from an aquifer under these conditions, would prevail over a relatively short time span.

4.3.2 The Areachap Terrane.

The volcano-sedimentary sequences of the Areachap Group, are host to numerous massive and disseminated stratiform metal sulphide deposits which are predominantly, though not exclusively, associated with the Jannelsepan Formation. These deposits are generally concentrated in four areas namely; at Areachap, northeast of Upington, at Bokspuits, between Upington and Putsonderwater, at Kielder, northwest of Prieska,

and at Copperton, near Prieska. The latter two deposit concentrations fall outside the quadrangle under investigation (Fig. 3.1) but are included in the discussion as this is deemed necessary for the metallogenic analysis of the Areachap Terrane. The Areachap Terrane is largely shear bounded (see Section 2.3.2), but the northern part is also separated from the south by the Bovenrugzeer Shear (Fig. 3.1). The northern Jannelsepan Formation is correlated with the southern Copperton Formation due to the similarities in rock type, age and style of mineralisation hosted by these formations (Theart, 1985).

The poorly constrained age of between 1600-1300 Ma for the Jannelsepan Formation and Copperton Formation (Barton & Burger, 1983) which hosts most of the massive sulphide deposits stands in contrast to the findings of Hutchinson (1973) who demonstrated that, globally, VMS deposits are particularly absent in mid Proterozoic rocks between 1600-700 Ma. He pointed out that during this time span, many of the larger sedimentary exhalative deposits (eg. Sullivan) were formed.

Due to the highly deformed nature of the host rocks, few primary characteristics of the ore bodies are evident, for example with regards to the zonation, feeder pipes, etc. (Section 4.3.2) which makes interpretation of these deposits difficult and often speculative. The Areachap deposit was described by Voet & King (1986) who recognised lapilli and bombs in the amphibolites which they interpreted to represent intermediate lava flows. The protolithologies of quartzo-feldspathic gneisses intercalated with the amphibolites are thought by these authors to represent rhyolitic tuff while garnet-biotite gneiss was either a pelitic sediment or an acid tuff. Generally, however, the sequence (Fig. 4.4) consists predominantly of mafic volcanics with felsic volcanics and metasedimentary units, one of which hosts the ore body.

At Areachap, Voet & King (1986) suggested that the ore was precipitated in a volcano-sedimentary basin. Initial deposition of sediments and basic volcanics was followed by explosive acid volcanics leading to the deposition of rhyolitic tuffs. During periods of inactivity with respect to volcanism, sediments including sandstones, pelites and arkose were deposited. Subsequent subsidence of the basin caused faulting in the

floor which allowed metal-carrying brines to enter the unconsolidated sediments leading to the precipitation of the massive sulphide ore. According to Theart (1985) there is no evidence for a direct relationship between the ore body and an alteration zone for the feeder pipe. He concluded that the Areachap deposit represents a distal deposit within a volcano-sedimentary succession. The amphibolites of the Areachap succession have both tholeiitic and calc-alkaline affinities and have been interpreted by Theart (1985) to represent intrusive or extrusive basalts and marl respectively. Theart (1985), in investigating the REE nature and $\delta^{34}\text{S}$ values of different sulphide minerals from Areachap and Prieska, concluded that the fluids which led to the deposition of the ore bodies were derived from a magmatic source. Theart (1985) suggested that this deposit may be classified as a Besshi-type deposit. The Bokspuits deposits are associated mainly with metasedimentary ferruginous chert and feldspathic amphibole gneisses which are intercalated with metavolcanic amphibolites (Geringer et al., 1987 and Fig. 4.4). Amphibolites intercalated with calc-silicate rocks are interpreted by Geringer et al. (1987) to represent reworked volcanic material (tuffs and pyroclastics) or marl while massive amphibolites associated with the mineralisation at Bokspuits are thought by these authors to represent a low-K tholeiite, suggestive of a fore-arc environment, and the pyroxene amphibolite to be slightly more calc-alkaline in character. Geringer et al. (1986) furthermore suggested that the mineralisation resembles that of the Besshi-type.

The Bokspuits deposits are characterised by a high Cu:Zn ratio, a moderate Fe:Cu ratio and a low abundance of pyrrhotite which point towards a proximal setting for the ore bodies. There is, however, no obvious evidence of alteration and feeder pipes in the host or surrounding rocks. The input of metal-carrying fluids seems to have been weak and the continuous deposition of sediments (quartzite in places associated with ore body) led to the low metal values at Bokspuits. There was, however, a large supply of iron and adequate sulphur resulting in predominantly oxide-sulphide iron mineralisation.

Table 4.4. General characteristics of the main sulphide deposits in the Areachap Terrane.

	AREACHAP	UPINGTON	BOKSPLOTS	KIELDER	PRIESKA
Main metal mineralisation	Zn-Cu	Barren	Cu-Zn	Zn-Cu	Zn-Cu
Additional metals	Accessory Pb, Au, Ag	Barren	Accessory Co, Ni, Bi, Pb absent	Minor Pb, high Ag/Au	Minor Pb, Mo, high Ag/Au
Oxide mineralisation	Magnetite, quartz, phlogopite	Ferruginous chert and BIFs.	BIFs, Fe-oxides and Ferruginous chert.	Barite, chlorite, phlogopite, apatite, tourmaline and quartz	Calcite, phlogopite, anhydrite, barite. Tourmaline horizon, manganese, magnetite or carbonate bearing layers.
Alteration	No direct relation of ore with alteration.	Little or no signs of hydrothermal alteration.	Alteration in the footwall, represented by calcite. Mineralised zone highly chloritised with intensive calcite and quartz vein fillings	Two ore bodies with chloritic/silicification alteration (hydrothermal vent), one ore body with no alteration zone and BIF in the hanging wall.	Ubiquitous phyllic and propylitic alteration. Mg-rich chloritic alteration pipe in the footwall.
Distance from vent	Distal.		Proximal and (distal, speculative).	Proximal and distal.	Proximal (Middleton, 1976; Theart, 1985), distal (Middleton & Van Schalkwyk, 1986).
Grade and tonnage	8.9 Mt at 2,24% Zn and 0.4% Cu.	Barren.	1.7 Mt at 1.5% Cu for only four of the ore bodies.	Grade and tonnage unpublished.	47 Mt at 3.8% Zn and 1.7% Cu.
Host rocks	Garnet-biotite schist and quartzite.	Pyroxene-amphibolite, amphibole-feldspar gneiss with intercalated calc-silicates.	Ferruginous chert, quartzite (western ore bodies), feldspathic amphibolite gneiss and chloritised feldspar gneiss (eastern ore bodies).	Quartz-cordierite and /or sillimanite granulite.	Quartz-feldspar-sillimanite gneiss.

Zonation	Increase in Zn towards outer periphery of ore body.		Outward zonation from Cu-rich centres through magnetite \pm Zn to hematite.	K3 ore body: Strong zoning from Pyrrhotite - barite - Zn, Cu also decreases outwards from the centre. K6 ore body: zoning absent.	Moderate outward zonation from Cu-rich centre to Zn-rich periphery.
Nature of ore	Massive and disseminated.		Mostly intercalated lenses of disseminated sulphides with host rock.	Massive overlain by disseminated lenses.	Both disseminated and discrete bands of massive sulphides.

from Middleton, 1978; Gorton, 1981; Theart, 1985; Voet & King, 1986; Middleton & Van Schalkwyk, 1986; Geringer et al., 1987 and Geringer & Ludick, 1990.

In the Uppington region, Geringer & Ludick (1990) interpreted the lower massive amphibolite (Fig. 4.4) to represent shoshonitic and high-K volcanics which is suggestive of a mature back-arc environment, whereas the pyroxene amphibolite has a protolithology of a low-K tholeiite and calc-alkaline basalt indicating a fore-arc environment. To date, no signs of mineralisation have been found in this area.

The Kielder ore bodies all lie within metasedimentary rocks associated with interbanded volcanic amphibolites. Gorton (1981) interpreted most of the gneisses and granulite protolithologies to be of a sedimentary origin (Fig. 4.4) and the amphibolites to be low-K tholeiites of island-arc affinity. The spectrum of intermediate rock types in the wall-rocks are possibly volcanic sediments and reworked tuffs (Gorton, 1981). The granulite grade of regional metamorphism attained at Kielder makes it difficult to distinguish between metasediments and metalavas of an intermediate composition. At various levels in the stratigraphy, above and below the ore bodies, there are horizons of amphibolite which have been interpreted by Gorton (1981) to represent basalts.

The Kielder deposits have minor barite within the massive sulphides but no massive barite horizons have been found along strike. Gorton (1981) deduced from chemical zonations and alteration of the host rocks that the three ore bodies of the Kielder deposit represent both proximal and distal deposits. He cautioned, however, that this does not necessarily mean that the mineralisation formed from the same exhalative vent as the ore bodies developed on different stratigraphic horizons.

It has been tentatively implied by Gorton (1981) that the Kielder Zn-Cu-type deposit, hosted by metasediments and intimately associated with low-K tholeiite is of a Besshi-type, implying a fore-arc setting, a forerunner to the Kuroko-type.

Controversy exists over the origin of the wall rocks of the Prieska deposit. Middleton (1976) observed volcanic textures such as pyroclastic fragments, amygdales and flow banding while Wagener & Van Schalkwyk (1986) noted sedimentary structures in the host rocks. Theart (1985) failed to recognise any primary igneous or sedimentary

SECTIONS THROUGH THE AREACHAP GROUP AT SELECTED PLACES.

(Not according to scale)

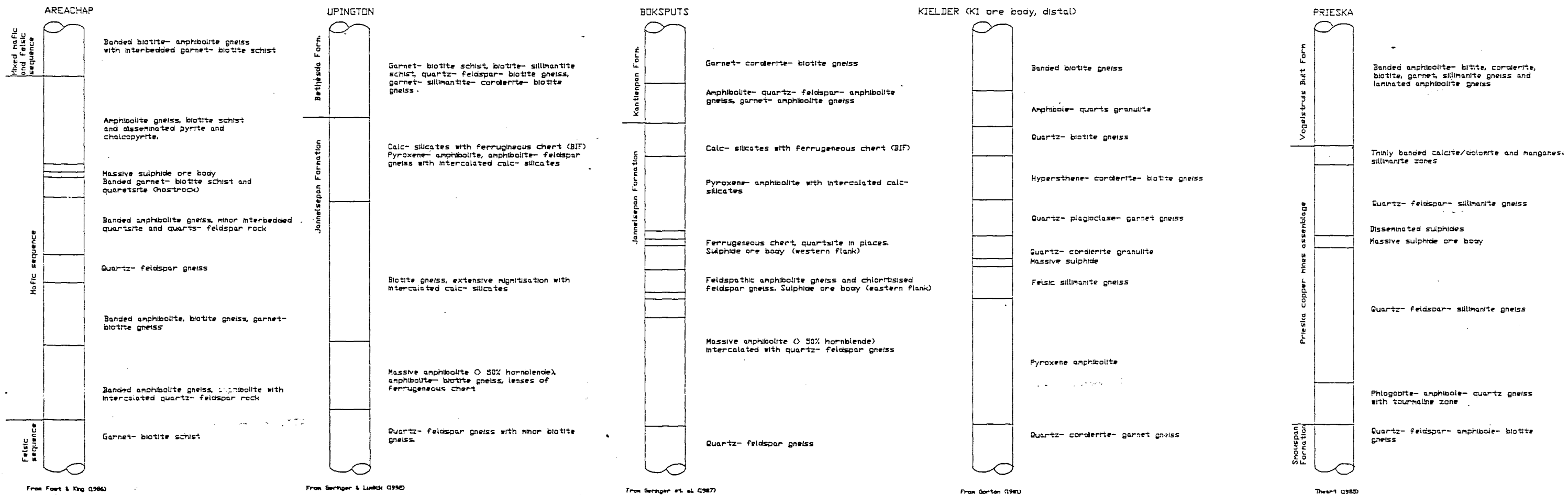


Figure 4.4: Generalised vertical sequences of the principle massive sulphide deposits in the Areachap Terrane.

textures or structures and considered them to have been obliterated during high-grade metamorphism and tectono-metamorphism.

Middleton (1976) proposed an explosive volcanic origin for the Prieska and Annex deposits. The ore formed syngenetically in calc-alkaline lavas and pyroclastics collectively known as the Copperton Volcanic Pile. Wagener & Van Schalkwyk (1986) concluded that the Prieska ore body was formed synchronously with the deposition of sediments in a marine basin where clastic sediments accumulated while the basin was open to the sea, and where chemical sediments were deposited during times when the basin was landlocked. They proposed that the ore body was chemically precipitated at the interface between oxic waters and anoxic brines and that distal island-arc volcanic exhalations may have contributed to the concentration of sulphur, base metals and B in the water.

Theart (1985) used geochemical methods to ascertain the nature of the host rocks for the Prieska deposit and to disprove Wagener & Van Schalkwyk's (1986) syn-sedimentary model. The peraluminous host rocks to the stratiform massive sulphide ore bodies have been interpreted as volcanic rocks that have been reworked during a sedimentary cycle. The basal intermediate to acidic gneisses (Smouspan Gneiss) were interpreted as having been derived from a calc-alkaline dacite, while the amphibolites intercalated with the so-called "Prieska Copper Mines Assemblage" are considered to represent intrusive dykes with a tholeiitic basaltic composition. The laminated amphibolite of the Vogelstruisbult Member may have been an extrusive tuff. The REE character and the low $\delta^{34}\text{S}$ values for the Prieska deposit indicate that the mineralising solutions were derived from a magmatic source.

4.3.3 Discussion.

From the previous section, there is little doubt that all the massive sulphide deposits in the Areachap Terrane, including the Prieska deposit, are of the Volcanic Exhalative type. In the Areachap Terrane, each of the mineralised regions, except for the Areachap deposit, contain proximal deposits (Table 4.4 & Table 4.5). From Figure 4.4,

it is clear that the different sections are difficult to correlate over the entire Areachap Terrane which may be ascribed either to local facies changes or that the sulphides did not form contemporaneously. The Upington area, investigated by Geringer & Ludick (1990), is apparently barren of massive sulphides and lies between the Bokspits and Areachap deposits. It can therefore be stated that each of these mineralised regions represents a centre of exhalation of metal carrying hydrothermal solutions. This is in agreement with the observations made by Sangster (1980), where volcanogenic massive sulphides in a metallogenic province are concentrated in clusters within a 32 km radius. In the Areachap Terrane the clusters are represented by the Areachap, Bokspits and Copperton/Kieller deposits.

In the Copperton/Kieller region there are five known massive sulphide deposits, made up of the Copperton and Annex (Middleton, 1976) deposits and the three deposits of Kieller. The largest of these is the Copperton deposit, containing 47 Mt at 1.7% Cu and 3.8 % Zn, while the grades and tonnages of the smaller deposits have not been published. These deposits are situated within an area of 20 km radius. If Sangster's (1980)

Table 4.5 Factors indicating proximity of Areachap Terrane ore bodies to exhalative vent.

AREACHAP	BOKSPITS	KIELLER	PRIESKA
Minor alteration associated with feeder pipe.	Minor alteration associated with feeder pipe.	K3 ore body: Chlorite and silica alteration. K6 ore body: No alteration.	Chlorite and silica alteration.
Moderate increase in Zn outwards.	Increase in Cu:Zn ratio.	K3 ore body: outward zoning from pyrrhotite - barite - Zn. K6 ore body: zoning absent.	Moderate zoning from Cu - Zn outwards.
No massive barite or BIF horizon.	No massive barite horizon. Associated BIF horizon.	K3 ore body: No massive barite or BIF horizon. K6 ore body: No massive barite. Associated BIF horizon.	Moderate barite.

evaluation of massive sulphide clusters, mentioned earlier, is presumed correct, and the Copperton deposit is assumed to be the largest within the cluster (47 Mt = 67%)

then the remaining tonnages within the Copperton/Kielder cluster region, although highly speculative, could be in the order of 23 Mt and the total tonnages in the cluster would be in the order of 70 Mt. The suggestion by Gorton (1981) that the Kielder deposits did not necessarily originate from a single vent could also indicate that there is a possibility of additional ore bodies in the region.

In the Bokspuits region, the available grade and tonnages of the deposits are not adequate to speculate about the potential of this region as defined by Sangster (1980).

The Areachap deposit is the only known ore body in this region. Much of the area around this deposit is covered by Dwyka Formation tillites towards the north. If Sangster's (1980) model of clusters for volcanogenic massive sulphide deposits is correct, then there is certainly a possibility of additional ore deposits in this region, either laterally or at depth, as the ore body and host rocks are steeply dipping. The possibility also exists that proximal deposits may have been eroded away. According to Sangster (1980), the average cluster contains about 92 Mt. If this value is accepted for a cluster of ore bodies in the Areachap region, then the present deposit only makes up 9.7% of the ore potential of this region. The fact that the Areachap deposit is considered to be distal, increases the likelihood of additional ore bodies closer to the source of the exhaled hydrothermal solutions.

The massive sulphide deposits in the Areachap Terrane are similar in that they all are dominated to varying degrees by Fe with lesser Zn and Cu and little or no Pb. This, and the geochemical nature of the amphibolites, may indicate that the ore constituents were derived from a similar type of mantle-derived source along the eastern margin of the Namaqualand Mobile Belt. Gorton (1981) proposed that all the basic igneous rocks in the Kielder/Copperton region have a common parental magma. The ore bodies are all underlain by a variety of different volcano-sedimentary sequences which militates against the metals having been leached from these rocks in general. It is therefore likely, as suggested by Theart (1985), that the source of these ore fluids is related to a deep-seated magma of mantle origin.

The VMS deposits of the Areachap Terrane are of the Cu-Zn type discussed by Lydon (1988a). Other factors that point to this conclusion include:

- * minor amounts of barite associated with the ore bodies,
- * relatively low sulphur isotope ratios,
- * Mg-rich chlorite alteration associated with the feeder pipes,
- * volcanic host rocks having both tholeiitic and calc-alkaline affinities.

There is general consensus among the various authors who have studied the massive sulphide deposits in the Areachap Terrane that these deposits are of the Besshi-type. This is not stated in absolute confidence, but is done so because the Besshi-type deposit provides the "best fit" for the Areachap massive sulphide deposits when compared to the available "classic models". There are, however, some similarities with Kuroko-type deposits especially with the Copperton/Kielder deposits and to a lesser degree with the Areachap deposit. These are factors such as the more polymineralic nature, which includes minor Pb, barite, anhydrite and the association of the ore bodies with relatively more felsic lavas, although specific felsic domes, usually associated with Kuroko-type deposits have not been recognised in the Areachap Terrane.

Hutchinson (1973) pointed out that Besshi-type deposits are typically associated with continental-margin island-arc environments in a subduction-zone setting. Tholeiitic to calc-alkaline magma produced in this type of tectonic setting could either be derived from the consumption of the descending oceanic plate in the Benioff zone or from partial melting of the mantle between the underthrust plate and the overriding continental plate. Progressive differentiation could produce volcanism spanning the entire compositional range from mafic to felsic rocks. As the maturity of the subduction zone increases, increased felsic activity as a result of anatexis of the subducting crust would be produced. Kuroko-type massive sulphide deposits are often associated with felsic intrusives at this more mature stage of island arc development.

Jakes & White (1972) pointed out that there is a compositional variation of the lavas across the island arc from a low-K tholeiite on the oceanward side of the arc, through

calc-alkaline, to high-K calc-alkaline to shoshonitic at the continental side of the arc. Gill (1970) and Barberi et al. (1974) furthermore suggested that the lavas of shoshonitic affinity are representative of the late phase of volcanic activity in an island-arc setting, indicating a more mature arc environment. Geringer et al. (1986) noted a predominantly low-K tholeiitic character for the amphibolites in the Bokspits area while the amphibolites in the Upington area are predominantly calc-alkaline to shoshonitic in character. They therefore concluded that the Bokspits and Upington regions are indicative of fore-arc and back-arc portions of an island-arc system respectively. Alternatively they proposed that the Upington area represents a more mature arc environment as opposed to a relatively immature arc system at Bokspits (Fig. 4.5 and Fig 4.6).

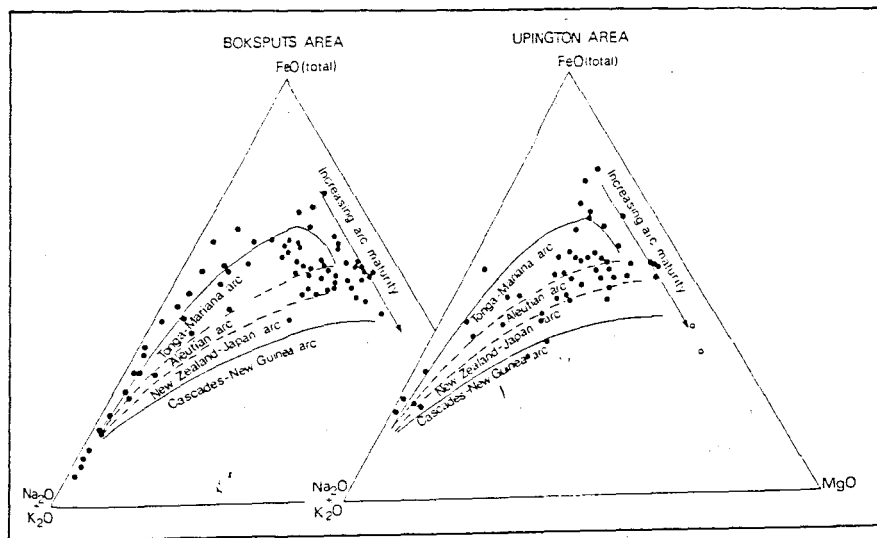


Figure 4.5: AFM-plot (after Brown, 1982) of amphibolites of the Areachap Terrane showing the increased arc maturity of the Upington area as opposed to the Bokspits area (from Geringer, 1986).

Investigations on the amphibolites in the Areachap and Copperton/Kielder areas have not yet constrained the amphibolites to such a degree of arc-related maturity. However, the relatively more polymineralic nature, the relatively higher felsic component of the host rocks and the appearance of barite, indicative of slightly shallower deposition, could all point to a greater arc maturity and a possible propensity towards a Kuroko type VMS deposit. Gorton (1981) indicated a low-K tholeiite as the origin of the amphibolites in the Kielder area which implies a fore-arc/relatively immature setting for the Kielder deposits. The differences between the Bokspuits deposit and the Areachap and Copperton/Kielder deposits could also be a manifestation of the dominance of Cu over Zn in the Bokspuits area and a reversal of this trend in the other areas. This could point to possible Kuroko-type deposits in the Areachap Terrane further north and south under the younger sediments of the Nama and Karoo rocks respectively.

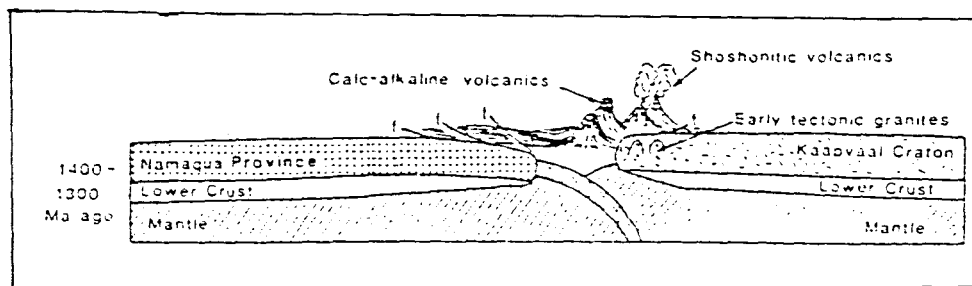


Figure 4.6: The tectonic setting of the Areachap Terrane showing the relationship between the low-K tholeiite, calc-alkaline and shoshonitic volcanism of the Areachap Group (from Geringer & Ludick, 1990).

4.4 THE WILGENHOUTSDRIF GROUP.

The Wilgenhoutsdrif Group (Fig. 4.7) is made up of mainly phyllitic rocks with subordinate quartzite, conglomerate, carbonates and ultramafic bodies, overlain by volcanic and sedimentary rocks. Moen (1988) recognised two cycles of volcanic activity, each composed of acid and basic lava and epiclastic sediments, but did not recognise any intermediate lavas. He classified the igneous rocks as tholeiitic basalt and rhyolite,

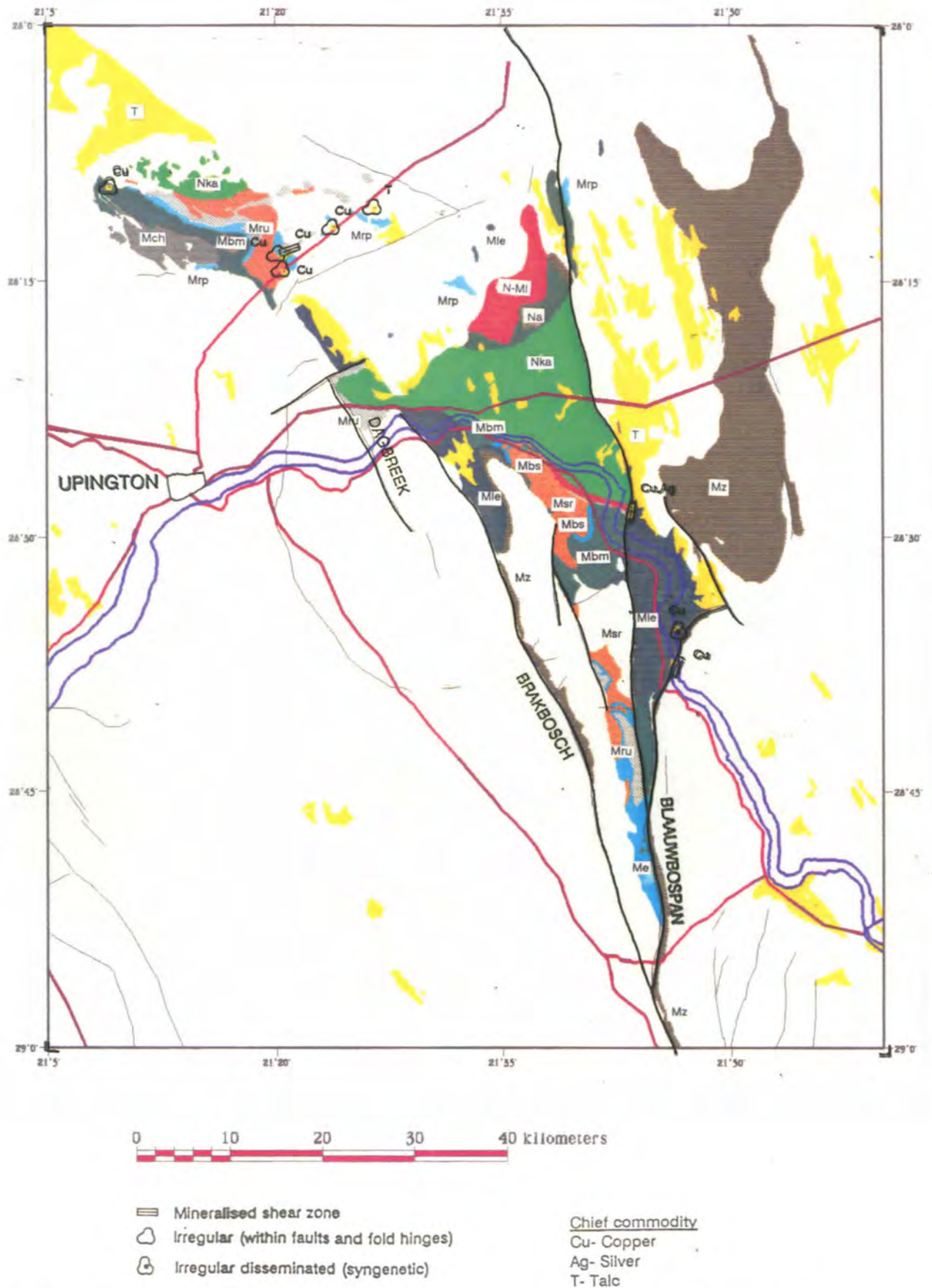


Figure 4.7: Mineral occurrences associated with the Koras and Wilgenhoutsdrif Groups with the more important shear zones.

and showed the former to be geochemically related to both ocean floor and intraplate environments. The acidic rocks were interpreted as having been derived from partial melting of crustal material. Moen (1988) stated that "the basaltic magma must have formed separately in the mantle, and during ascent created a steep geothermal gradient, which could initiate partial melting in the crust".

The Wilgenhoutsdrif Group has been interpreted by Moen (1988) as having been deposited in a tectonically unstable basin with intermittent volcanism controlled by deep faulting. Repeated movement along the faults may be manifested in the cyclic repetition of volcanic activity and sedimentation.

Known mineralisation in the Wilgenhoutsdrif Group is restricted to disseminated Cu mineralisation within the basaltic metalavas which show indications of pillows. Surface features consist of malachite staining and chrysocolla. At one locality, Cu-Ag mineralisation is associated with a hydrothermal breccia along faults containing quartz and calcite. The mineralised faults lie adjacent to a major fault which separates the Wilgenhoutsdrif Group in the east from the Koras Group in the west. Two periods of mineralisation in the Wilgenhoutsdrif Group are thus evident, namely, syn-genetic mineralisation with the influx of basic volcanics into the basin and later, epigenetic invasion of hydrothermal fluids along structural weaknesses, possibly related to Koras-aged tectonism. Much of the Wilgenhoutsdrif Group is covered to the north by Tertiary deposits, under which further mineralisation is likely.

4.5 JACOMYNSPAN Cu-Ni DEPOSIT IN PERSPECTIVE.

The Jacomynspan chalcopyrite-pyrrhotite-pentlandite-bearing ultramafic body is intrusive into porphyroblastic quartz-feldspar-biotite-garnet gneiss of the Jacomynspan Group (Attridge, 1986). The east-west-trend of the intrusive body lies roughly parallel to, and some 3 km south of the Bovenrugzeer Shear where it bisects the Areachap Terrane into the northern Jannelsepan and southern Copperton portions (Section 4.3.2). Attridge (1986) suggested that the ultramafic body intruded syn to late-tectonically. Although few detailed geochemical studies have been performed to

ascertain the origin of the ultramafic rock, its proximity to the Areachap Terrane, as well as its syn-orogenic character must be taken into account. The magma body, possibly of mantle origin, intruded into a pre-existing fault or shear zone (Attridge, 1986) prior to the development of the major period of lateral movement along the eastern margin of the Namaqua front which is manifested in dextral shear zones such as the Brakbosch and Bovenrugzeer shear zones. It seems likely that the ore represents immiscible sulphides present in the magma at the time of emplacement.

Cu-Ni sulphide-bearing ultramafic bodies along major structural discontinuities associated with suture zones related to subduction have been documented by Sawkins (1972), but a genetic model for these intrusives remains enigmatic. Examples of these occur along the sheared boundary between the Superior and Churchill Provinces of the Canadian Shield which have been interpreted as an ancient island arc (Zurbrigg, 1963). A further example is provided by the ultramafic intrusions along the convergence of the Russian and Siberian platforms which formed the Urals during Hercynian times (Hamilton, 1970). The possibility therefore exists, that additional Cu-Ni mineralised mafic to ultramafic intrusives are present in the proximity of shear zones along the Areachap Terrane.

4.6 SERPENTINITES IN PERSPECTIVE.

Lenses of serpentinites, some of which are chrysotile-bearing, occur sporadically within the Dagbreek Formation of the Vaalkoppies Group. Many serpentinites essentially lie in close proximity to the west of the Brakbosch shear zone (Fig. 4.8). Chrysotile mineralisation is contained within fractures which were probably caused by deformation of the ultramafic host rock under tensional stress or dilation associated with the Brakbosch transcurrent shearing event. The chrysotile was probably deposited in the fractures from constituents derived locally by lateral secretion under relatively low-grade greenschist metamorphism (Riordan, 1981).

The origin of the serpentinites is probably related to the intrusion of mantle-derived ultramafic material along the structural weaknesses created by the Brakbosch shear

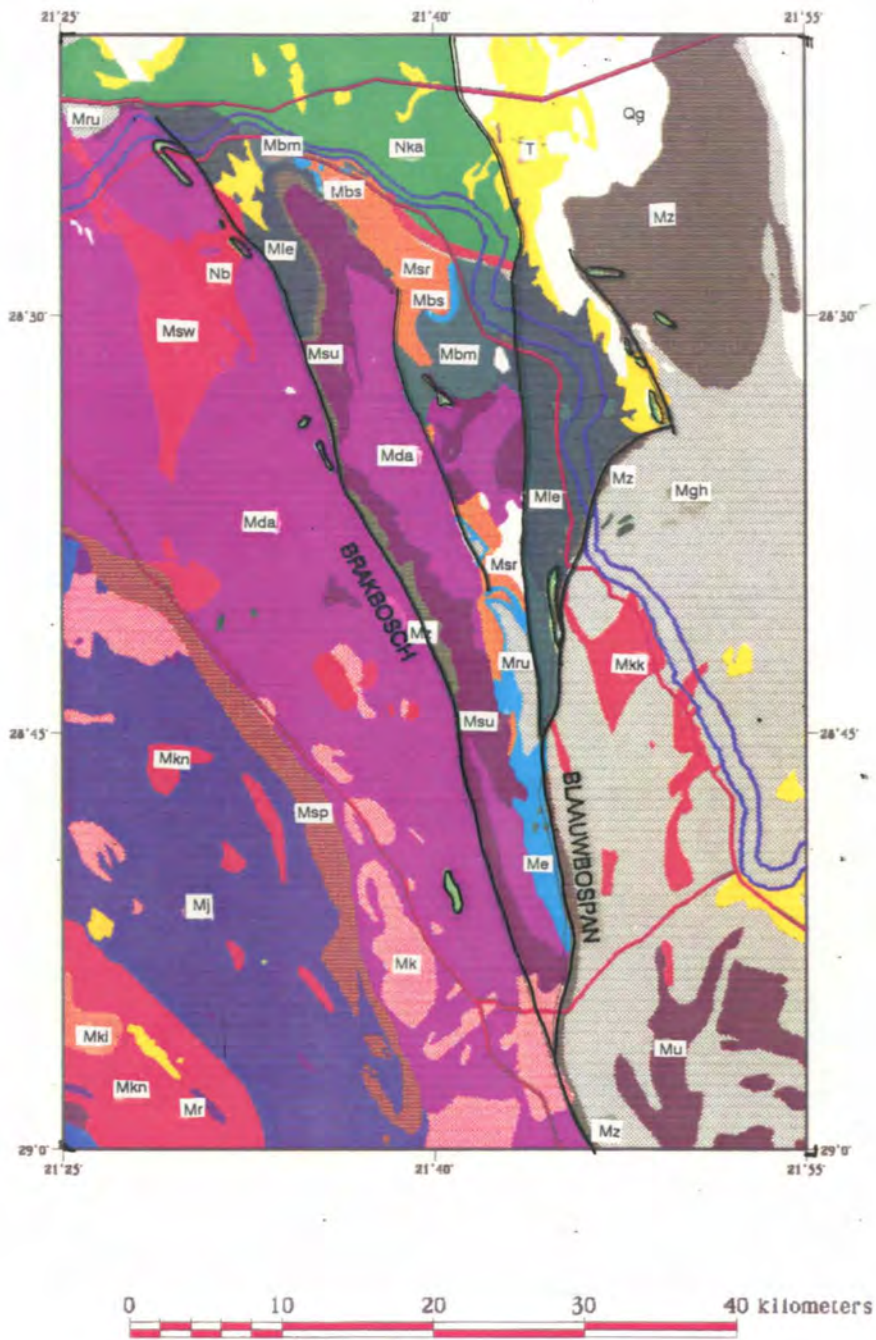


Figure 4.8: The distribution of the serpentinites and their association with the Brakbosch Fault. Serpentinites are lime green.

zone during the collision of the Areachap island-arc. The existence of mobilised serpentinitised mantle material in convergent plate boundaries is not uncommon (Hess, 1955; Moores, 1973). The sporadic nature of the serpentinite bodies could well have been the result of lateral movement by the Brakbosch shear zone and additional serpentinite bodies may exist along the expanse of the shear zone. Other shear zones in the area may also have provided structural weaknesses along which the ultramafic material could have intruded.

4.7 HYDROTHERMAL VEIN DEPOSITS AND THEIR POTENTIAL RELATION TO GRANITOIDS.

4.7.1 Introduction.

The economic potential of an orogenic province is partially dependant on the nature and intrusive level of the associated granitoids. The position of the intruded granitoids is a reflection of deep-seated lineament zones, mainly along the north-west-trending Kakamas Terrane. Granitoid related deposits are largely confined to the Gordonia Subprovince, especially within the Kakamas Terrane. The vein-type structural relationship with the host rocks imply a syn to late-tectonic age of mineralisation. It is therefore probably not coincidental that the majority of vein-type mineralisation and the Keimoes Suite granitoids are both largely confined to the Kakamas Terrane, and that there should be a genetic link.

In the area under investigation, there are two W-Sn metallogenic provinces, both within the Kakamas Terrane, namely, the Van Roois Vlei W-Sn Province, west of Upington (Fig. 4.9) and the Riemvasmaak W-(Sn) Province, near the Namibian border (Fig. 4.12). Fe-Cu-Ag vein deposits are situated northwest of the Van Roois Vlei W-Sn deposits, along the regional tectonic trend, and comprise the Lutzputs Fe-Cu-Ag Metallogenic Province (Fig. 4.13). Peripheral to the Van Roois Vlei Metallogenic Province in the south, there are numerous Cu ± amethyst ± fluorite-veins deposits (Fig. 4.14) which are possibly related to the same granitoid intrusive event which formed the Van Roois Vlei W-Sn deposits.

4.7.2 The origin of ore-forming solutions in granitic systems (an overview).

Stremprok and Skuor (1974) concluded from a review of 260 tin-bearing granitoids from 16 provinces that the differences in major elements between the tin-bearing granites and the barren granites is minor, but noted a general tendency towards excess SiO_2 and K_2O . In this regard they noted the following: "the excess of silica over the granite average may be explained by leucocratic tendency of tin-bearing granites as well as by secondary processes including silicification and greisenisation. The excess of K_2O may in part be explained by potash metasomatism known from these granites". These authors, however, stress that these trends are by no means universal. Analysis of the Keimoes granite Suite by Geringer et al. (1988) indicated "normal" SiO_2 and slightly depleted values for K_2O compared to the standard quartz monzonite.

With respect to trace elements, Taylor (1979) noted that all granites related to tin provinces are geochemically specialised and contain anomalous amounts of one or more of the following elements: Sn, F, Cl, Li, B, Rb. However, Taylor (1979) also pointed out that, for example, in the Cornwall tin province in southwest England, the presence of geochemically specialised granites was rather rare. It is therefore possible that granitic suites that show little or only partial evidence for specialisation could yield tin mineralisation. In general, high concentrations of Sn, F, Cl, Li, B, and Rb could indicate the presence of mineralisation, although their absence does not necessarily mean the lack of any mineralisation. Minerals which act as tin concentrators and may indicate trace element specialisation are biotite, muscovite, hornblende, ilmenite, sphene and to a lesser extent feldspar (Taylor, 1979).

Most, if not all granitic rocks associated with Sn-W deposits are of the S-type, derived from a sedimentary source. The differences between the I-type and S-type granites are highlighted in Table 4.5. Geringer et al. (1988) (see Section 2.3.3) noted, however, that the Keimoes Suite granites are of the I-type affinity. The classification of granites into the magnetite- series and ilmenite- series (Ishihara, 1977) has not yet been undertaken on the granites of the Northern Cape. Although the S-type and I-type granites of Chappell and White (1974) are generally correlated with the ilmenite- and magnetite-

Table 4.6. The characteristics of the I-type and S-type granites.

S-type	I-type
Relatively low sodium. Na ₂ O normally < 3.2% in rocks with approx. 5% K ₂ O, decreasing to < 2.2% in rocks with approx. 2% K ₂ O	Relatively high sodium. Na ₂ O normally > 3.2% in felsic varieties, decreasing to > 2.2% in more mafic types
Mol. Al ₂ O ₃ /(Na ₂ O + K ₂ O + CaO) > 1.1	Mol. Al ₂ O ₃ /(Na ₂ O + K ₂ O + CaO) < 1.1
> 1% C.I.P.W. normative corundum	C.I.P.W. normative diopside or < 1.1% normative corundum
Relatively restricted in composition to high SiO ₂ types	Broad spectrum of compositions from felsic to mafic
Variation diagrams irregular	Regular inter-element variations within plutons; linear and near linear variation diagrams
Muscovite and biotite common	Hornblende common
Sr ⁸⁷ /Sr ⁸⁶ > 0.708	Sr ⁸⁷ /Sr ⁸⁶ 0.704 - 0.706
Metasedimentary xenoliths common	Mafic hornblende-bearing xenoliths of igneous appearance common
Sn association with highly silicic types	Porphyry Cu and Mo (W) deposits associated with felsic to mafic types

From Plimer (1980), modified from Chappell & White (1974).

series granites respectively, there are numerous exceptions and certain ilmenite- series granites have I-type compositions. The main characteristic features that distinguish the magnetite- and ilmenite- series granitoids are displayed in Table 4.7. Ishihara (1981) found a clear relationship between the magnetite- series and ilmenite- series and mineralisation (see Table 4.6). The magnetite- series granitoids contain a higher concentration of sulphur and certain metals which may result in the formation of metal sulphide mineralisation. These granites also have a relatively high chlorine content and Cl/F ratio, compared to the ilmenite- series, which could also account for the development of concentrations of Cu, Pb, Zn, Mn, Au, Ag, and Hg which may be transported as chloride complexes (Barnes, 1979). Sn, W type mineralisation is generally associated with ilmenite- series granites. Ishihara (1981) pointed out that S-type, ilmenite- series granites form at relatively shallower depths and have little chance for magmatic differentiation and concentration of ore components. The composition of these granitoids and their metal ratios depend largely on their original compositions or scavenging from the wall rocks. The ore-forming metals, as well as fluorine or chlorine, then accumulate in the cupolas of the granites.

Biotite, in the solidifying process of a granitic magma, is considered to play an

important role in the behaviour of F, which is mainly present in this mineral (Bailey, 1977). The Fe/(Fe + Mg) ratio in biotite for magnetite- series granites is lower than for the ilmenite-series. The lower Fe content in biotites of magnetite- series granites relative to those of ilmenite- series, is a result of Fe being taken up by the magnetite. Mg-rich biotite can contain more F than Fe-rich biotite (Munoz & Ludington, 1974) and therefore more F may remain in the residual magma of ilmenite- series granites during crystallisation of biotite.

Table 4.7. The main characteristics of the Magnetite- and Ilmenite- series granitoids.

MAGNETITE- SERIES	ILMENITE- SERIES
Abundance of magnetite	Practically free of magnetite
Magnetite often constitutes more than 90% of opaque oxide minerals	Minor opaque oxide minerals, ilmenite usually in ferromagnesian minerals
Sulphides are typically pyrite and minor chalcopyrite	Pyrrhotite common, chalcopyrite less common than in Magnetite- series
Sphene and epidote group minerals	Garnet and monazite
Fe/(Fe + Mg) ratio in biotite and amphiboles relatively low	Fe/(Fe + Mg) in biotite and amphiboles relatively high
Solidify under relatively higher f_{O_2} conditions	Solidify under relatively higher f_{O_2} conditions
Refractive index of biotite between 1.600 and 1.650	Refractive index of biotite higher than 1.650
Lithophile elements (F, Rb, Li, Pb, Sn, Be) show little variation with differentiation index	Lithophile elements (F, Rb, Li, Pb, Sn, Be) increase with differentiation index
Relatively higher chlorine and lower fluorine contents	Relatively higher fluorine and lower chlorine contents
$\delta^{34}S$ values are positive	$\delta^{34}S$ values are dominantly negative
Produces Mo, Zn, Pb, Cu, Ag, Au, Hg, Mn and W (scheelite) deposits	Produces W, Sn, Be, F and pegmatite deposits
Occur in dominantly tensional tectonic setting	Occur in dominantly compressional tectonic setting
Biotite is greenish in hand specimen and greenish-brown z-colour under the microscope	Biotite is greasy black in hand specimen and reddish-brown z-colour under the microscope
Related to I-type	Related to S- and I-type

Summarised from Ishihara (1981).

Tin is usually transported with fluorine complexes, but may also be transported as chlorides. These fluoro- and chloro-complexing mechanisms are essentially the same for base metals (Eadington & Giblin, 1979). Tungsten partitions from the melt into the hydrothermal phase and is transported as hexahalide species (Foster et al., 1975).

It is likely that many other components of ore deposits, especially Li, Be, F, B and highly charged ions such as Sn, W, Mo and Bi behave in a similar fashion (Plimer, 1980). Elements such as Sn, B, F and Cl would readily partition from quartz-tourmaline \pm cassiterite or pyrrhotite \pm cassiterite \pm tourmaline rocks during anatexis of metasediments, whereas W, F, Be, Mo and Bi would readily partition from calc-silicate rocks into the melt (Plimer, 1980). The first components to partition into either the anatectic melt or the final components of the residual fractionates of a granitoid are those which are not easily accommodated in the silicate structures.

The presence of fluorine and boron in rocks substantially lowers the solidus of these rocks and enrichment of F and B by fluid-rock interaction may also allow melting to take place at lower temperatures (Manning & Pichavant, 1983). In high grade-metamorphic rocks, F occurs at either low-level "background" values contained within hydroxysilicates, micas and amphiboles or as higher-level F carried by non-silicates such as fluorite, fluor-apatite and topaz. The presence of an aqueous fluid phase may, however, be necessary before the melting temperature of a F-bearing hydroxysilicate assemblage is reduced (Manning & Pichavant, 1983).

Boron is primarily hosted by tourmaline which has a wide stability range in high-grade rocks. Other borosilicate minerals, though far less abundant, include dumortierite ($\text{Al}_7\text{BSi}_3\text{O}_{18}$), kornerupine ($\text{Mg}_3\text{Al}_6(\text{BAlSi})_5\text{O}_{21}\text{OH}$), grandidierite ($(\text{Mg,Fe})\text{Al}_3\text{BSiO}_9$) and sinhalite (MgAlBO_4) (Manning & Pichavant, 1983). Tourmaline enrichment in metamorphic rocks could be derived from either a B anomaly within sediments or by the introduction of B-rich fluids related to magmatic activity.

There are two basic concepts which may lead to the enrichment of B and/or F. The first is the enriched source concept which includes partial melting of, for example, B-rich metasediments (Bernard et al., 1985) or B-Li-F-enriched metasediments in evaporitic environments (Waters & Moore, 1985) and, secondly, the fractionation concept which involves the concentration of these elements as a result of fractionation during magma evolution and crystallisation (Christiansen et al., 1983). The enrichment of B or F may be manifested in the igneous stage by the crystallisation of topaz,

fluorite, tourmaline or Li-mica, and/or in the late magmatic stage by the development of similar hydrothermal minerals. The main distinction between B- and F-bearing magmas is in the water content (Pollard et al., 1987). B-bearing magmas enhance the solubility of H₂O thus increasing the thermal energy released during crystallisation of residual magmas relative to F-bearing magmas. This phenomena could lead to an outward zonation from F-rich to B-rich precipitation away from the magmatic source, in a system which contains significant amounts of both fluorine and boron. Thus Pollard et al. (1987) concluded that B-rich systems are characterised by tourmaline with tourmaline, silica, muscovite, feldspars and chlorite being the main alteration minerals and breccia pipes, veins and stockwork vein systems being the dominant mineralisation style. F-rich systems are characterised by fluorite and/or topaz and Li-micas with alteration zones of fluorite, feldspar, muscovite and silica in the form of massive greisen or veins and stockwork systems.

Further observations of outward zonation from the granitic source are often noticed in many granite-related ore districts, the most commonly quoted being that of the Cornwall district in southern England (Hosking, 1963). There is usually a typical outward zonation from Sn, W, As and U through U, Ni, Co to Cu followed by Pb, Zn, Ag to Fe and Sb sulphides (Strong, 1990). Taylor (1979), however, cautioned against using the Cornwall district as a classic example as there are many overlaps and spatial variations to such patterns, but conceded that many examples display inner zones rich in Sn and W, fringed by sulphide ores containing Pb, Zn and Cu. Furthermore, Taylor (1979) also warned against the misconception that these zones formed in disc shapes around the source granite or parallel the granitic contact. The shape of the mineralised district, as well as the zones surrounding the granite, is also controlled by a variety of other factors including pressure, temperature, wall rock interaction, structure and f_{O_2} .

4.7.3 The Van Roois Vlei W-Sn Metallogenic Province.

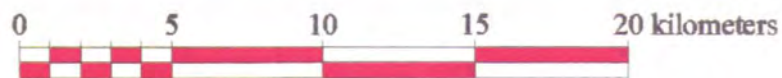
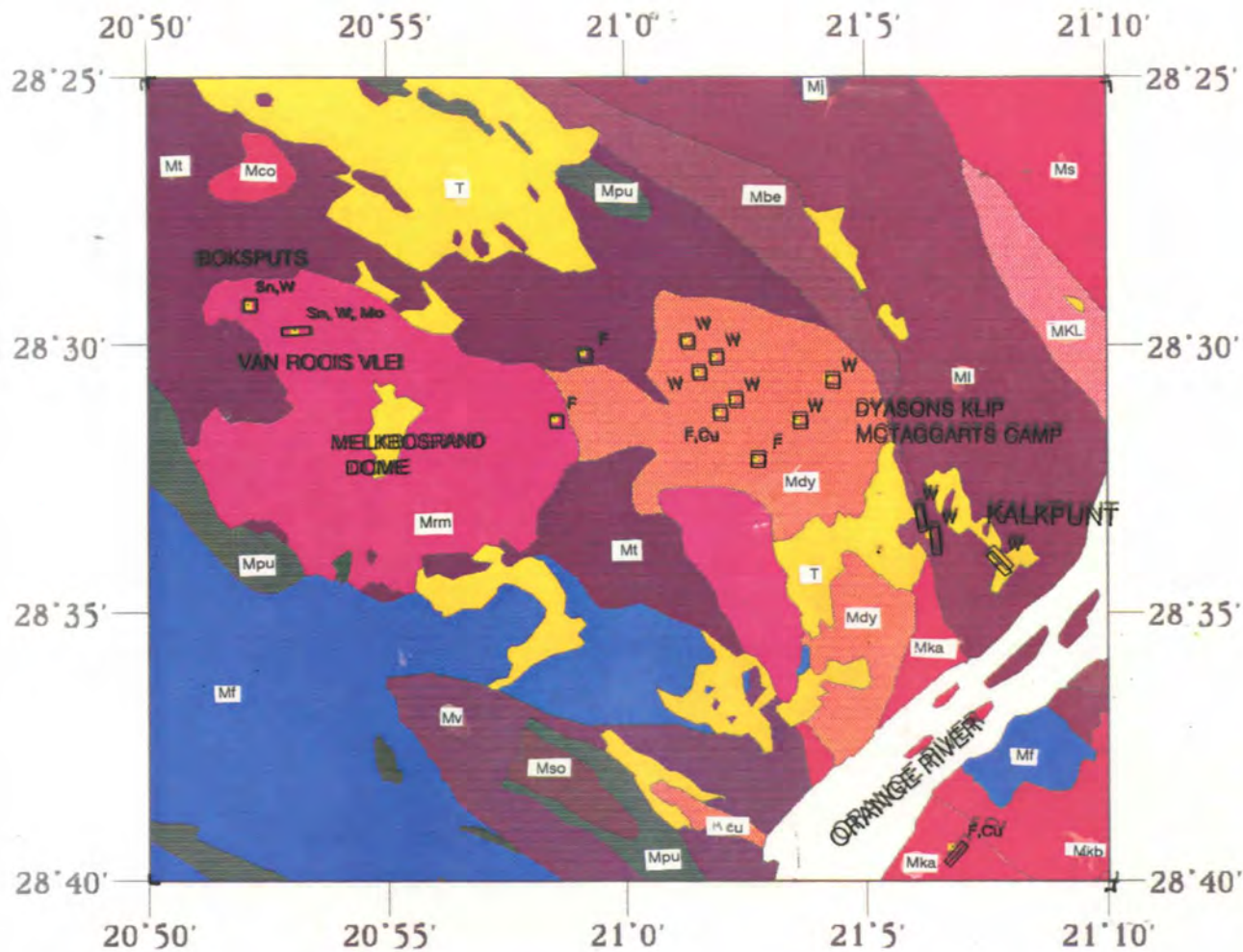
The ore-bearing veins in the Van Roois Vlei Metallogenic Province are hosted by the Dyasons Klip Gneiss in the east and by the Toeslaan Formation metasediments and

Riemvasmaak Gneiss in the west (Fig. 4.9).

In the Dyasons Klip - McTaggart's Camp region in the eastern portion of the Van Roois Vlei Metallogenic Province, the mineralisation is characterised by numerous north-northwest-trending fractures containing scheelite-wolframite \pm cassiterite \pm arsenopyrite \pm chalcopyrite \pm sphalerite \pm molybdenite. The wall-rock mineral association is characterised by quartz-tourmaline-biotite-muscovite-fluorite \pm plagioclase. These veins are hosted by the pre to syn-tectonic Dyasons Klip Gneiss. Large northeast-trending fluorite veins are also present in the area and further east, just north of the Orange River, where north-south-trending wolframite and scheelite-bearing quartz-tourmaline veins occur in the syn to late-tectonic Louisvale granite and in the garnet-biotite schist of the Bethesda Formation.

At Van Roois Vlei the mineralisation is hosted by east-west-trending quartz-tourmaline \pm K-feldspar \pm fluorite veins which are present within the paragneisses of the Toeslaan Formation and the quartzo-feldspathic Riemvasmaak Gneiss. The ore consists of scheelite, wolframite and cassiterite with magnetite, hematite and pyrite being common additional constituents. Molybdenite, sphalerite and chalcopyrite occur in minor amounts (Smithies & Pirajno, 1989). Wall-rock alteration assemblages include potassic metasomatism, tourmalinisation, chloritisation, silicification as well as lesser fluoritisation, sericitisation and ferruginisation (Smithies & Pirajno, 1989).

At the Bokspuits deposit, northwest of Van Roois Vlei, scheelite and lesser wolframite and cassiterite with pyrite and pyrrhotite are present within a series of quartz \pm tourmaline \pm fluorite veins which occur within the metapelitic gneisses of the Toeslaan Formation. Here the Toeslaan Formation has been tectonically isolated within the Riemvasmaak Gneiss (Smithies & Pirajno, 1989). According to Smithies & Pirajno (1989), the wall-rock alteration at Bokspuits indicates a lower



- | | | |
|---|--------------|------------------------|
| □ | F- dominated | <u>Chief commodity</u> |
| ▭ | B- dominated | W- Tungsten |
| ▨ | Peripheral | Sn- Tin |
| | | Mo- Molybdenum |
| | | F- Fluorite |
| | | Cu- Copper |

Figure 4.9: Distribution of Sn-W-F mineralisation in the Van Roois Vlei area.

fluorite:tourmaline ratio when compared to the Van Roois Vlei deposit.

4.7.3.1 Two possible origins for the hydrothermal solutions.

4.7.3.1.1 Late stage specialised granite.

From factors such as the spatial relationship of the hydrothermal veins, similar geological setting, alteration style and type, mineral paragenesis and ore mineralogy, Smithies and Pirajno (1989) deduced that these deposits were formed as a result of the same hydrothermal event. From the relative abundances of fluorite and tourmaline associated with the mineralisation in parts of the Van Roois Vlei Metallogenic Province, these authors proposed that the McTaggart's Camp, Bokspits and Van Roois Vlei deposits are exo-granitic greisen-type deposits whose alteration-mineralisation styles reflect increasing distances from a granitic source. The more proximal McTaggart's Camp deposits locally show a F domination over B, while the more distal Van Roois Vlei deposits are characterised by a total dominance of B over F. At the intermediate Bokspits deposit, fluorite is a common mineral, although still subordinate to tourmaline. However, southeast of the McTaggart's Camp deposits, just north of the Orange River the north-south trending mineralised wolframite and scheelite-bearing veins are dominated again by tourmaline and contain little, if any, fluorite. Wall-rock alteration is largely limited to ferruginisation. This reveals a regional zonation and the possibility that these deposits are again, more distal from the source in the opposite direction to the Van Roois Vlei deposits (Fig. 4.10).

Smithies & Pirajno (1989) superimposed Fe/Mg ratios of tourmaline from the endo-granitic Zaaiplaats and exo-granitic Rooiberg tin deposits in the Transvaal over those of the tourmalines in parts of the Van Roois Vlei Metallogenic Province and confirmed the distal relations of these deposits from a common source (Fig. 4.11). They found that MgO increases away from the source.

Smithies & Pirajno (1989) postulated that cassiterite-bearing leucogranitic dykes within the deposit area provided the source for the ore-forming hydrothermal fluids.

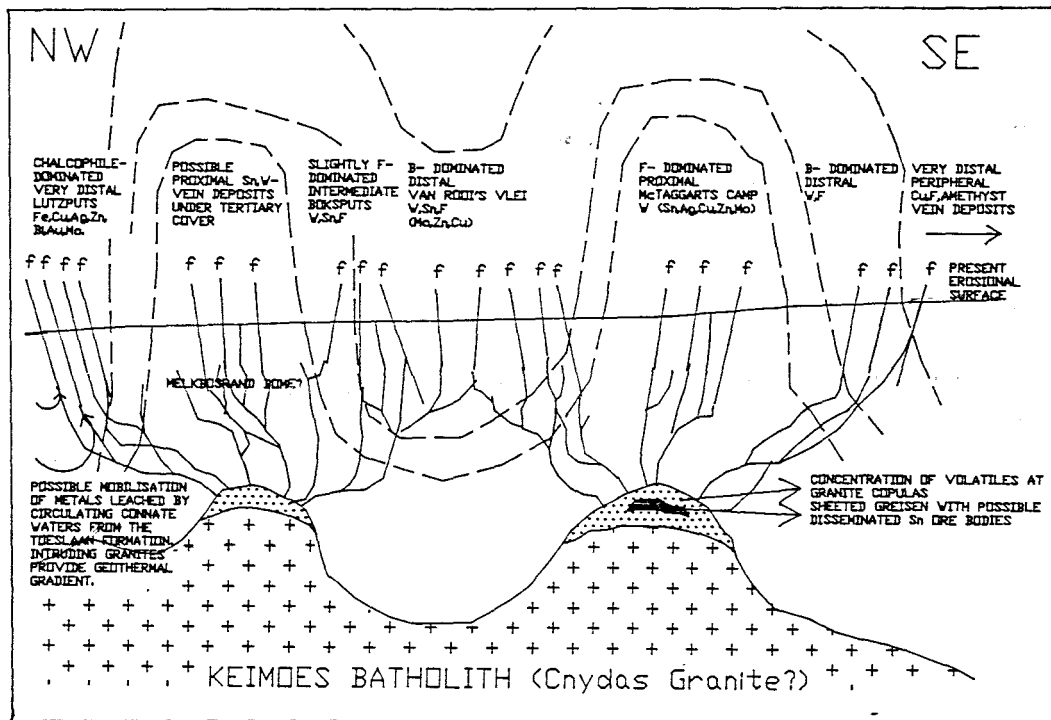


Figure 4.10: Schematic cross section of the exo-granitic vein-type mineralisation of the Van Roois Vlei Sn-W-F Province showing the possible zoning within the province.

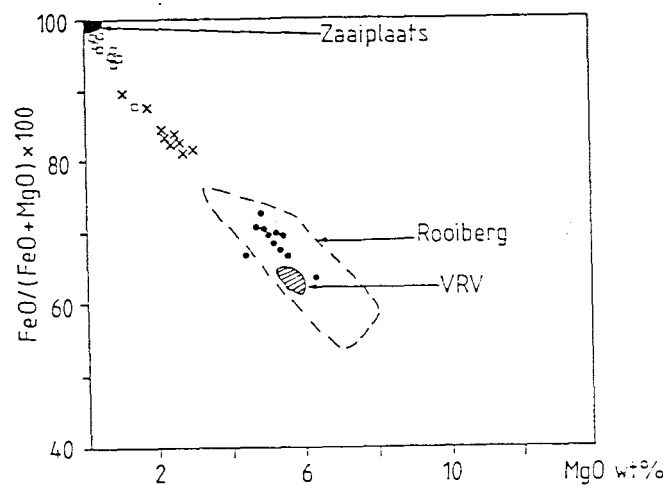


Figure 4.11: The Fe/Mg ratios of tourmaline from the exo-granitic Zaaiplaats and the endo-granitic Rooiberg Sn deposits superimposed on the tourmaline of the Van Roois Vlei Sn-W-F Province as a function of distance. The diagram shows the increase of Mg with increased distance from the granitic source rocks (from Smithers & Pirajno, 1989).

- Open squares = McTaggart's Camp leucogranites.
- Crosses = McTaggart's Camp Selvages.
- Closed circles = Bokseputs tourmalines.
- VRV = Field of Van Roois Vlei tourmalines.

These leucogranitic dykes could be connected at depth with a larger granitic body similar to the post-tectonic Cnydas Granite some 25 km northeast of the Van Roois Vlei deposit. The Cnydas Granite, which forms part of the Keimoes Suite, is of calc-alkaline affinity and displays a fractional differentiation series from granodiorite to alkali-granite (Jankowitz, 1987). Some of these granites are Sn specialised and the presence of tourmaline orbicules attests to high B values which could be considered as a reflection of the early stages of a developing magmatic hydrothermal system during the latter phases of fractionation (Jankowitz, 1987). An I-type and magnetite-series geochemical signature for this granite with anomalous amounts of W (up to 375 ppm) and Sn (up to 21 ppm), as well as an enrichment in B, was demonstrated for these granites (Jankowitz, 1987). Higher anomalies of W, in the form of scheelite are found in the granites enriched in tourmaline (higher B values) and Jankowitz (1987) noted a depletion of Sn in the older phases of the Cnydas granitoid intrusions.

4.7.3.1.2 Anatectic remobilisation.

The presence of tourmalinites east of Kakamas, which lie within metasediments of the Korannaland Sequence, and the spatial association of many of these deposits with tourmaline-bearing pegmatites as well as similar circumstances surrounding other Sn-W mineralisation elsewhere in the NMC, led Bowles (1988) to conclude that the deposits on Van Roois Vlei are a product of remobilisation and reconcentration, and that the tourmalinite horizons represent the protore. The origin of the tourmalinite horizons, although highly speculative, could represent volcanic exhalatives (Bowles, 1988). The presence of B (in tourmaline) in metasediments during high grade metamorphism could promote the generation of magma by lowering the melting temperature and concentrating metals within the partial melt. Ideally, these volatile-rich fluids would then be transported along the pressure gradient and be deposited as veins along shear zones and faults.

In examining a tourmalinite horizon east of Kakamas, Bowles (1988) found that it contained consistently high Sn values (up to 308 ppm) and some of the samples contained up to 33 ppm W. This would support the hypothesis that the tourmalinites

represent a possible source from which the metals could have been remobilised. Tungsten deposits hosted by tourmaline and the general enrichment of tungsten within exhalites, for example in iron formations of Greenland (Appel, 1986) indicate that tungsten can be a component of some submarine exhalative systems.

4.7.3.1.3 Discussion.

Both hypotheses for the origin of the hydrothermal fluids which formed the Sn-W mineralised veins in the Van Roois Vlei Metallogenic Province are acceptable, and neither can be disproved outright. The spatial relationship of tungsten deposits with tourmalinite horizons throughout the NMC as described by Bowles (1988) is without doubt noteworthy. Bowles (1988) also described and suggested that, among others, the tungsten deposits around Okiep are associated with tourmalinite horizons of submarine exhalative origin. Raith (1991), however, concluded that these deposits near Okiep were derived from hydrothermal fluids which originated from the highly evolved Spektakel Suite of granitoids. These granitoids produced highly differentiated pegmatites and aplitic leucogranites in the area. Geochemistry of these granitoids reveals strong similarities with other W- and Sn-W specialised granites (Stemprok, 1984).

The suggestion by Smithies & Pirajno (1989) that the tourmalinites near Van Roois Vlei could be a product of selective replacement of sedimentary horizons by the degassing of a B-enriched granite was refuted by Plimer (1988) who noted that "unreplaced" equivalents of the same horizon have the same composition, suggesting a lack of chemical control for any replacement.

Although Slack (1982) suggested that tourmalines associated with submarine exhalative ores are Mg-rich and those associated with felsic plutonic rocks are Fe-rich, Plimer (1983) and Taylor & Slack (1984) found that tourmalines from tourmalinites of exhalative origin have a compositional range from Mg-rich to Fe-rich and occasionally compositions overlap those from greisens and granites. The compositional range of tourmalines can therefore not indicate conclusively whether they originate from

exhalatives or from felsic magmatism.

The association of numerous granophile elements such as F and Mo suggest, though not conclusively, that the origin of the mineralising fluids for the deposits in the Van Roois Vlei Metallogenic Province originated from a specialised granite. Exhalative Sn-W deposits at Felbertal, Austria (Höll, 1977) and in Argentina (de Brodtkorb & de Brodtkorb, 1977), however, are accompanied by minor concentrations of Mo, Cu, Bi, Au, Ag and Be (Austria) and F, Fe base metal sulphides, Mo, Be and Au (Argentina). The overwhelming majority of Sn-W-deposits directly related to granitic magmatism throughout the world are associated with S-type or ilmenite- series granites, yet the Cnydas Granite, which has been proposed as the origin of the mineralising fluids, is an I-type/magnetite- series granite. Only a very small component of the Cnydas Granite contains a Mol. $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$ ratio of > 1.1 and only a small percentage of samples (6%) taken contained more than 1% CIPW-normative corundum. Fluorine and pegmatite generation is also largely considered to be associated with the ilmenite- series granites.

At the Zaaiplaats Sn mine Coetzee & Twist (1989) have, however, shown that the Nebo Granite of the Bushveld Complex, which has an I-type signature, can enrich Sn enough to form economically viable concentrations. They showed that the upper portion of the Nebo Granite (Bobbejaankop Granite) was formed as a result of fractionation from an unspecialised Nebo granite which contains < 5 ppm Sn (Kleemann & Twist, 1989) and therefore did not geochemically inherit Sn through fractional differentiation (the Cnydas Granite contains up to 21 ppm Sn and up to 375 ppm W). They furthermore concluded that there is no evidence for upper crustal contamination and assimilation. Coetzee & Twist (1989) pointed out that fractional crystallisation enhanced the volatile content, thereby promoting the conditions to facilitate the exsolution of a fluid phase containing F and Cl that would scavenge Sn. This model requires a closed system over an extended time period. It is therefore possible that similar processes were active during the latter stages of a portion of the intruding Keimoes Suite granite. Finally, cooling and solidification of the pluton would enhance pressure build-up and subsequent fracturing of the carapace releasing the

metal-bearing volatile-rich hydrothermal fluids into the surrounding Toeslaan Formation.

A further possibility, however, is that the ascending Cnydas Granite incorporated some of the ore metals as it passed through the tourmaline horizons. Jankowitz (1987) pointed out the increase of the molar $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$ ratio towards the younger, more leucocratic phases of the batholith. This suggests a propensity towards the S-type granite during the latter phases of the intrusive and the possible synchronous assimilation of elements such as Sn, W, B, F, Be, Li, Mo, Bi and base metals which would readily partition into the melt, but yet would remain until the final stages of crystallisation. Concentration of these elements into the residual melt and late stage aqueous fluids, rich in volatiles, would then lead to the concentration of Sn and W from the uneconomical (exhalative) tourmalinite horizons into fractures leading to the formation of the relatively more economical deposits in the Van Roois Vlei Metallogenic Province.

4.7.4 The Riemvasmaak W-(Sn) Metallogenic Province.

The Riemvasmaak W-(Sn) Metallogenic Province is situated in the north western portion of the area under investigation (Fig. 4.12). The following represents a brief summary of the more important aspects of these deposits from Bowles (1988). The stratabound W-bearing veins are hosted by pelitic gneisses, schists and quartzofeldspathic granulites of the Koelmanskop Metamorphic Suite. In the central portion of the Metallogenic Province, a group of quartz-wolframite veins with minor amounts of scheelite, known as the Collinskop deposit, occurs within the partially migmatized garnetiferous biotite granulite and biotite-garnet-sillimanite rock (kinzigite) of the Bok-se-puts and Collinskop Formations respectively. Some of the mineralised veins lie adjacent to, and subparallel to the contact between the host rocks and the biotite-rich granite gneiss of the Donkieboud Granite Gneiss, which forms part of the Eendoorn Suite, and are generally conformable to the foliation. Sulphides within the veins are pyrite, chalcopyrite and locally arsenopyrite.

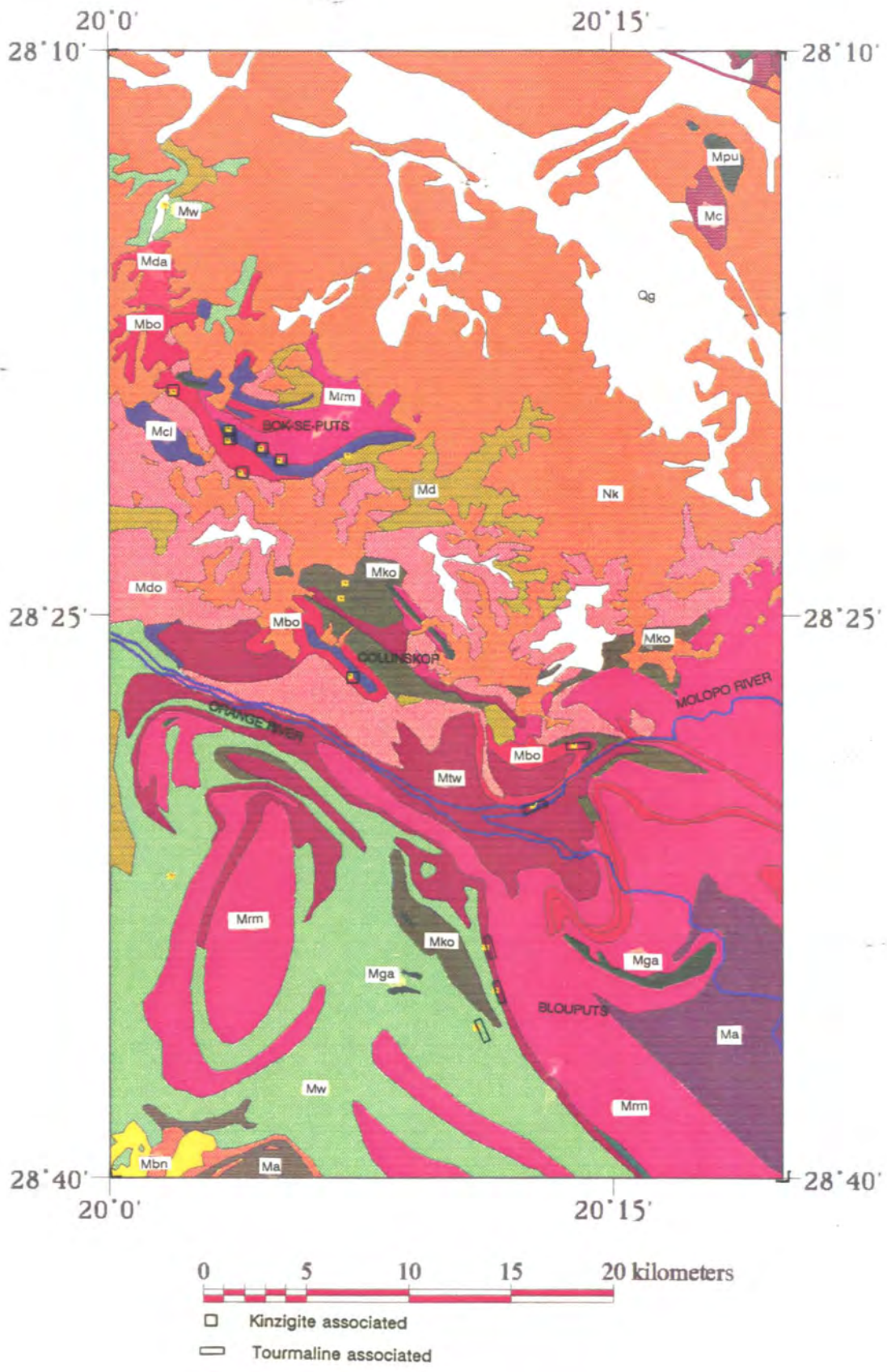


Figure 4.12: Distribution of W mineralisation in the Riemvasmaak area.

Further northeast, the Bok-se-puts deposits consist of wolframite-scheelite-bearing quartz veins which include chalcopyrite, pyrite, loellingite, bornite and azurite as well as feldspar, garnet, epidote, sericite, biotite and tourmaline. Towards the east of the Metallogenic Province, the Molopo W deposits are associated with thin bands of quartz-tourmaline rock which is interbedded with garnetiferous biotite granulite of the Bok-se-puts Formation. South of the Orange River, W is hosted by feldspathic quartz veins within schistose portions of the garnetiferous biotite gneiss of the Twakputs Gneiss which also forms part of the Koelmanskop Metamorphic Suite. According to Moen (quoted by Bowles, 1988), the mineralisation in this area shows a close association with a narrow stratiform tourmalinite unit which lies interbedded with the pelitic gneiss.

With the exception of the Renosterkop deposit, which is handled separately (see Section 4.7.5), this Metallogenic Province is characterised by the general lack of Sn and only locally a minor tourmaline component, the latter indicating little or no boron in some of the mineralising systems (eg. Collinskop). There is also a noticeable absence of fluorite in any of the W mineralised deposits of this Metallogenic Province as opposed to those in the Van Roois Vlei Metallogenic Province. In the Riemvasmaak area there is no indication at the surface of any highly evolved or differentiated late-stage granites and there is little evidence for the association of these mineralised veins with the granites. The close association of some of the mineralised veins with the biotite-rich (possibly S-type) pre to syn-tectonic Donkieboud Granite Gneiss may, however, contradict the above conclusion, but further work would be necessary in assessing the potential of this granite to produce W mineralisation.

Generally, the W mineralisation in the north-western portion of the Riemvasmaak Metallogenic Province (Bok-se-puts - Collinskop area) is associated with kinzigites and the south-eastern portion (Molopo to south of the Orange River) is associated with tourmaline. It is therefore possible that the processes leading to the concentration of certain metals under conditions of high-grade metamorphism and anatexis in a B "dominated" system described in the section dealing with the Van Roois Vlei Metallogenic Province could have led to the formation of the deposits in the

Riemvasmaak Metallogenic Province that are associated with tourmaline. Bowles (1988a) likened these deposits and tungsten deposits associated with tourmaline-rich rocks at Yanco near Broken Hill, N.S.W., Australia (Barnes, 1983).

As far as the association with kinzigite is concerned, Moen (quoted by Bowles, 1988) suggested that the kinzigites represent an aluminous-rich variety of restite, formed during ultrametamorphism or granitisation of the surrounding garnetiferous biotite granulite. Bowles (1988) suggested the close association of W metallogenesis and the formation of kinzigite. During this process granophile elements such as quartz and feldspar become mobile and were removed from the system while the granophobe elements, rich in Mg-Fe-Al remain immobile. The large, highly charged tungsten ion, generally incompatible with silicate minerals, enters the fluid phase of the partial melt thereby concentrating the tungsten and precipitating it in nearby structures at sites of favourable physio-chemical environments.

4.7.5 The Renosterkop Sn-W-Zn deposit.

Localised patches of coarse-grained wolframite, heterogeneously disseminated cassiterite as well as pyrite, sphalerite and chalcopyrite are distributed within a quartz-biotite-topaz rock known as the Renosterkop Formation, which is underlain by quartzo-feldspathic gneiss of the pre to syn-tectonic Riemvasmaak Gneiss. Minor amounts of fluorite are also present. The mineralised zone is surrounded by a pervasively chloritised alteration halo followed by a biotite-rich alteration zone.

Saad (1987) suggested that the mineralisation formed as a result of the influx of metasomatic F-rich fluids during the final stages of deformation when a late stage thrusting event caused a weakening of the host rock to allow for conduits, along which the fluids migrated. The Riemvasmaak Gneiss, host to the mineralisation, became greisenised *in situ* by these metasomatic fluids forming the Renosterkop Formation. Saad (1987) furthermore proposed a multiphase invasion of metasomatic fluids into the conduit system which probably resulted in the remobilisation and concentration of the metals precipitated in previous events. He used this to explain the lack of metal

zonation between Sn, W and Zn within the biotite greisen. The origin of the mineralising fluids is unknown, but a highly fractionated granitoid body (or pegmatite) at depth could have been a likely source.

4.7.6 The Lutzputs Fe-Cu-Ag Metallogenic Province.

The Fe-Cu-Ag vein deposits in the Lutzputs area (Fig. 4.13), as described by Bicker & Ralston (1986), occur in northerly and north-easterly-trending shear zones mainly within metasediments of the Toeslaan Formation, Biesje Poort Group. Metasediments include banded garnetiferous, biotite gneiss with quartz granulite intercalations, biotite-garnet-sillimanite gneiss and pegmatitic gneiss. According to Bicker & Ralston (1986), there are two main shearing directions, namely, an older north-easterly and a younger northerly trend. Mineralisation is predominantly within the northerly trending shears. The host rocks to the mineralised veins display extensive propylitisation (chloritisation, kaolinisation and silicification) which rapidly decreases in intensity away from the mineralised zone and is probably related to the period of sulphide mineralisation (Bicker & Ralston, 1986).

Mineralised shear zones are largely associated with magnetite-hematite concentrations and vary considerably in grade, width and strike length. Mineralisation consists essentially of magnetite, hematite, arsenopyrite, chalcopyrite, pyrite, pyrrhotite and silica gangue together with freibergite, sphalerite, bismuthinite, gold and molybdenite (trace) which do not necessarily occur together. Within a portion of the Lutzputs Metallogenic Province there is a possible northward zonation from chalcopyrite-freibergite association in the south to a chalcopyrite-pyrite association to almost exclusively arsenopyrite in the north. Bicker & Ralston (1986) noted a paragenetic sequence as follows: magnetite, hematite, manganese oxide, arsenopyrite, pyrite, pyrrhotite, freibergite, chalcopyrite, sphalerite, bismuthinite and gold.

The origin of the mineralising fluids which led to the precipitation of these deposits is an enigma. Due to the limited nature of the knowledge available on these deposits, the following hypothesis on the origin of the metals is highly speculative. The wall rock

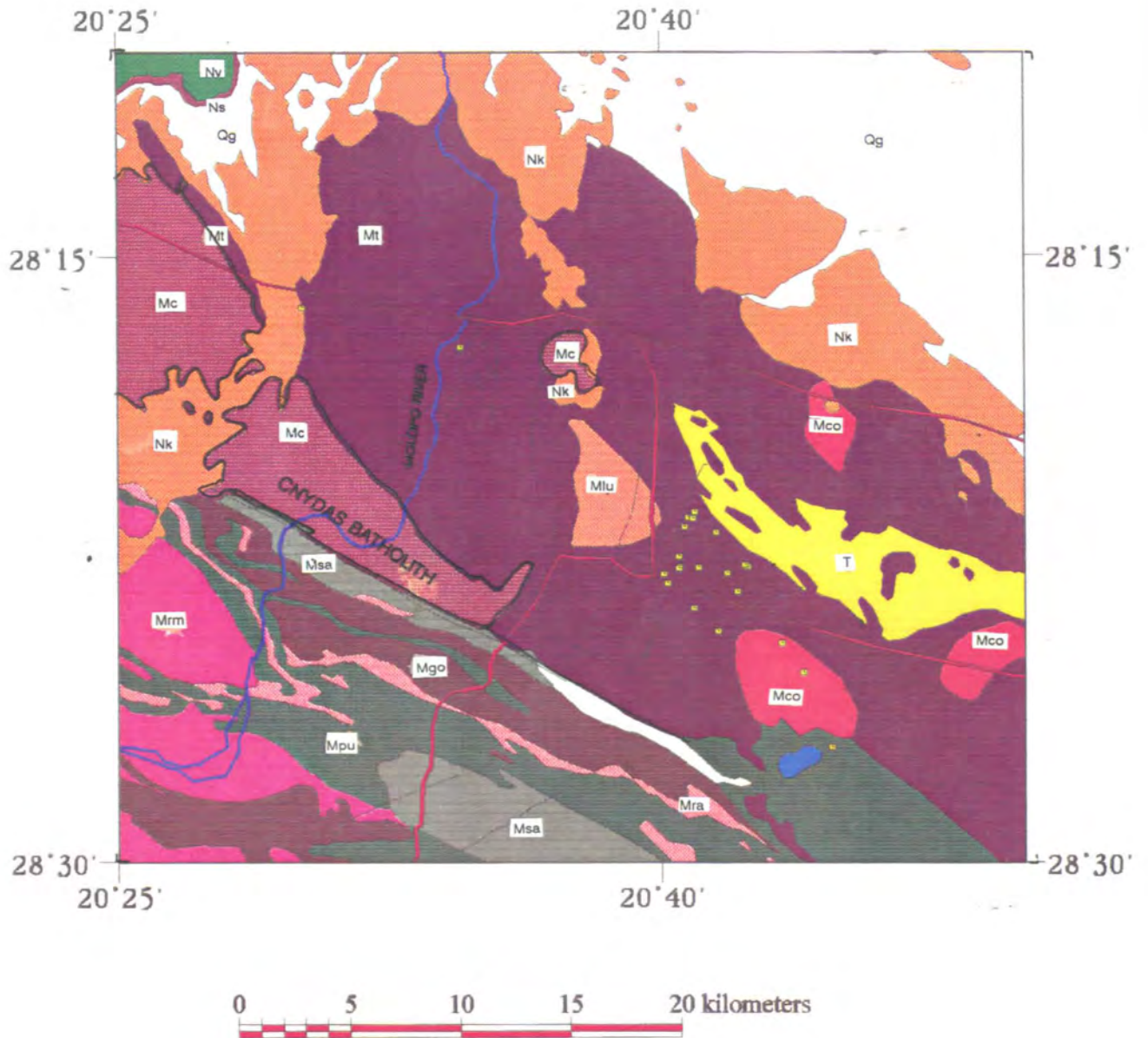


Figure 4.13: Distribution of the mineralisation in the Lutzputs Fe-Cu-Ag Province.

alteration associated with the mineralisation leaves little doubt as to the hydrothermal nature of the ore-bearing fluids and the preference of these deposits to form in shear zones of a certain direction attests to the influx of these fluids during a specific time interval of late stage deformation.

If a granitic origin is assumed for the Van Roois Vlei W-Sn Metallogenic Province, the proximity of the Lutzputs Province to the Van Roois Vlei W-Sn Metallogenic Province and the Cnydas Granite could indicate a possible granitic source for the mineralising fluids, and suggest that a possible broad scale zoning, from a proximal Sn-W-B-F-dominated system to a more distal chalcophile-dominated system is present (Section 4.7.2 ; Strong, 1990 and Fig. 4.10). The mineralised quartz veins were then deposited from hydrothermal solutions emanating from the Cnydas Granite and emplaced along dilatant shear zones during the last stages of intrusion. The Cnydas Granite shows a systematic depletion of Zn and to a lesser extent Cu from the oldest phases to the younger phases (Jankowitz, 1987).

It is also possible that the intruding granites of the Keimoes Suite or, more specifically, the Cnydas Granite could only have provided the heat source necessary to drive connate or meteoric waters within the metasedimentary Toeslaan Formation. The water and CO₂, driven by the prevailing thermal gradient, would percolate through the metasediments thereby mobilising elements such as SiO₂, S, As, Zn, Ag, Au etc., and draw them into vein faults in a type of epithermal system where they would precipitate in the more dilatant parts. An increase in the CO₂ content of the percolating fluids would depress the mobility of silica and, together with sulphur, enhance the mobility of Fe, Mn and the chalcophile elements (Boyle, 1965) found in the mineralised veins. Magnetite, pyrite, Cu, Mo, Bi, Ag, Au mineralisation is typical of epithermal systems driven by I-type granites.

4.7.7 Other deposits potentially associated with granitoids.

Numerous vein related deposits are distributed mainly within the Kakamas Terrane and lie predominantly peripheral to the Van Roois Vlei Metallogenic Province in the

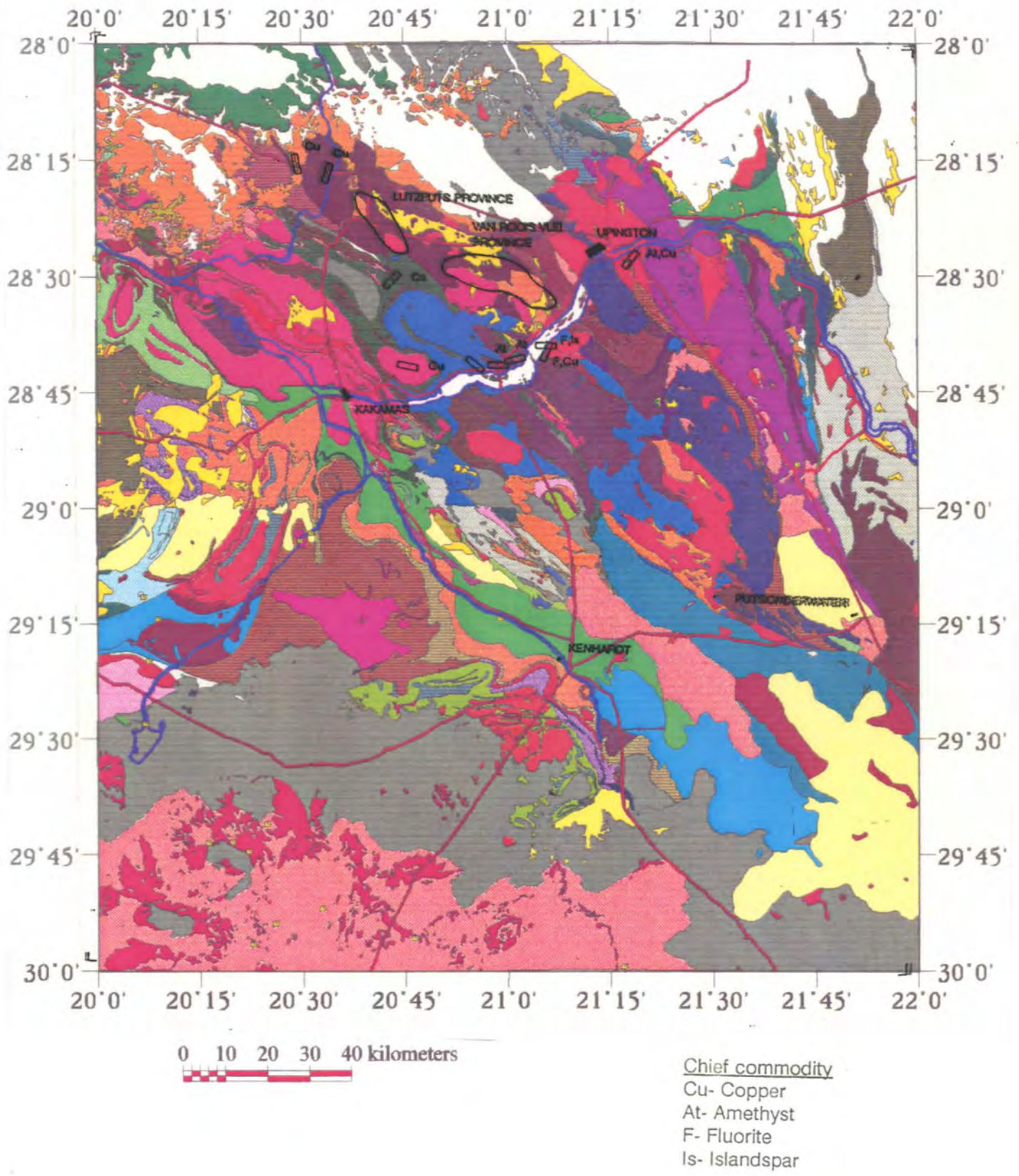


Figure 4.14: Distribution of vein deposits peripheral to the Van Roois Vlei Province.

south (Fig. 4.14). These veins are host to mainly Cu ± amethyst ± fluorite. The proximity of these mineralised veins to the Van Roois Vlei Sn-W-F Metallogenic Province is remarkable, and one is tempted to relate these deposits to a similar granitic source, if a granitic source for the Sn-W-F vein deposits at Van Roois Vlei is accepted. These veins could then represent a continuation of the outward zonation of the more chalcophile elements from a common granitic source, but in the opposite direction to the Lutzputs deposits (Fig. 4.10).

The veins are mainly hosted by the syn to late-tectonic granitoids of the Keimoes Suite such as the Vaalputs and Kanoneiland Granites and Friersdale Charnockite in the vicinity of the Orange River. The possibility also exists that these veins represent late-stage fluids emanating from the host granite themselves. However, the common association of the mineralised veins with shear zones and the association of some of the veins with charnockite, a traditionally "dry" granite which rarely produces a residual aqueous phase, thereby reducing the solubility of chalcophile elements, militates against an endo-granitic source for these veins.

4.7.8 The Pypklip West F vein deposits.

These veins constitute one of the few mineralised quartz-vein systems within the Bushmanland Subprovince in the portion under investigation (Fig. 3.3). The veins are hosted within northeast-trending shear zones in biotite gneisses and amphibolites of the De Banken Formation. In the main Pypklip West deposit, fluorite is associated with calcite, galena and pyrite, while the other fluorite-mineralised veins in the vicinity are not known to contain any sulphides. The paragenetic sequence as determined by Hugo (1962) is from epidotisation, quartz, feldspar, calcite, fluorite, galena and pyrite which precipitated during three phases of influx of hydrothermal fluids.

Within the region, there are numerous shear zones, many of which contain quartz veins, which exhibit similar dip and strike attitudes and are younger than the igneous rocks in the vicinity. Apart from the young, unfoliated Uitkoms Granite to the southeast of these veins on the farm De Banken 130 which may have an association

with the mineralisation, the lack of any late-stage felsic plutonism argues against any granitic source for the hydrothermal solutions and another source must be found. The association of these brecciated mineralised veins with relatively young shear zones indicate that the hydrothermal fluids were formed during a tectonically active period at the final stages of deformation. Harris (1992) pointed out that there this area experienced significant northwest-southeast extension late-stage dextral shearing. It is possible then, that there is a relationship between late tectonism and regional metamorphism.

During metamorphism, H₂O and CO₂ would be mobilised as a result of an increase in the geothermal gradients within the sedimentary piles. The most mobile constituents of the country rock such as SiO₂, Ca and CO₂ would be drawn from the rocks and precipitate in the dilatant structures of lower pressure, representing the first and second stages described by Hugo (1962, see Section 3.6.2.2.1). The precipitation as a result of either the decrease of pressure or temperature would lead to throttling of the vein-system and renewed pressure built-up. The increased hydrostatic pressure could lead to fracturing and subsequent pressure loss as well as possible retrograde boiling. During the boiling process, volatiles may escape into the vapour phase leaving the hydrothermal solution less capable of metal transport. At this stage Fe and Pb would be precipitated in the form of pyrite and galena within the shear zones, thereby representing the final stage of deposition.

4.7.9 The Uitkyk Boven De Kalkgaten W vein deposit.

The 50 m wide zone of quartz veins containing W, Cu and Mo lies within the quartzofeldspathic gneiss of the Kokerberg Formation and is situated along the contact with the intruded syn to late-tectonic biotite gneiss of the De Bakken Granite (Fig. 3.3).

The spatial association of these veins with the De Bakken Granite points towards a possible relationship between the ore-forming hydrothermal fluids and the intruded granite. Frick (1986) demonstrated that the De Bakken Granite is of calc-alkaline affinity and that it is slightly fractionated but that it cannot be regarded as a Sn-

specialised granite. The De Bakken Granite, however, was shown by Frick (1986) to contain elevated values of Sn, but he did not apparently analyse for W. It is therefore possible that the partially fractionated De Bakken Granite, which is not considered to be part of the Keimoes Suite, could well have been the source for the mineralising fluids. This isolated deposit furthermore, lies in the Bushmanland Subprovince and is therefore not considered to be related to the Van Roois Vlei W-Sn Metallogenic Province.

4.7.10 Discussion.

If a granitic source is accepted for the vein deposits peripheral to the Van Roois Vlei Metallogenic Province, as well as the Van Roois Vlei Metallogenic Province itself and the Lutzputs Metallogenic Province, either directly or indirectly (heat source), it is possible to include these metallogenic provinces into one regional metallogenic province where all these deposits are related to fluids emanating from a common granitic source. The spatial correlation between the Keimoes Suite and the above mentioned metallogenic provinces, both lying within the Kakamas Terrane, cannot be ignored. Geringer (1988) demonstrated that the Keimoes Suite is a composite batholith comprising of predominantly I-type granitoids with some S-type components, and that the granitic plutons in the vicinity of the Van Roois Vlei Metallogenic Province are largely mesozonal.

Most Sn deposits are associated with highly differentiated biotite granites with a noticeable absence of hornblende or other calcium silicates such as amphiboles, pyroxenes and sphene (S-type, magnetite-series), which may be a prerequisite for the enrichment of Sn as an incompatible element in the residual melt (Ishihara, 1977). Tin-bearing granites commonly contain F, Li and/or B (Heinrich, 1990). Tungsten-only deposits are generally associated with I-type, ilmenite-series granitoids, encompassing a much wider compositional range from granodiorite to alkali-feldspar granites (Kwak, 1987).

Many cassiterite-bearing veins contain quartz and muscovite or Li-micas and are

frequently associated with fluorite, chlorite, arsenopyrite and/or pyrite (Heinrich, 1990) as is the case in the Van Roois Vlei Metallogenic Province. Heinrich (1990) pointed out that in many composite or crosscutting veins, this oxide phase is followed by a sulphide stage precipitating minerals such as chalcopyrite, pyrrhotite, sphalerite, galena, stannite and other sulphides. This stage is also evident in the Van Roois Vlei Metallogenic Province. The presence of F and B in the system indicates the presence of a highly evolved granite at depth as the B and F concentrated in the aqueous phase during differentiation which greatly enhance the mobility, and lowered the solidus of the hydrothermal fluids. This allows the escape of these fluids into the country rock following the build-up of lithostatic pressure. The W-Sn-(F, B) veins of the Van Roois Vlei Metallogenic Province as well as the F-bearing veins peripheral to the W-Sn province is related to a highly evolved component of the Keimoes Suite batholith. The search for highly evolved S-type or I-type granites at depth is therefore pertinent with respect to exploration. As Coetsee & Twist (1989) pointed out, fractionated, volatile-rich granites are of greater importance in the search for, example Sn, than the actual Sn content of the granite. Likely trap sites for these types of deposits are pre-existing faults/fissures, contacts between various lithologies (i.e. zones of structural weaknesses where fluids under pressure may flow down pressure gradients). Other trap sites may be cracks formed by the sudden release of lithostatic or hydrostatic pressure. It is also likely that W-Sn-rich greisen deposits could be found at the cupolas of these evolved mesozonal granites (Renosterkop).

The mineralogy of the Lutzputs Metallogenic Province (magnetite, pyrite, Cu, Mo, Bi, Ag, Au) is more indication of an I-type granite. These veins deposits were probably derived from an I-type component of the Keimoes Suite batholith with or without an influence of connate/meteoric waters which leached metals from the Toeslaan Formation metasedimentary host.

4.8 PEGMATITES.

4.8.1 Introduction.

The pegmatite belt in the area under investigation generally follows the regional trend of its host rocks and is largely confined to the Kakamas Terrane between the Hartbees River Thrust in the west and the Cnydas-Bovenrugzeer Shear in the east. From Figure 3.3 it can be seen that most of the pegmatites are concentrated in the western portion of the Kakamas Terrane, west of the Cnydas Shear, and may be genetically associated with the tectonic boundary between the Gordonia and Bushmanland Subprovinces.

The pegmatites are host to a variety of pegmatite minerals, the most economically important being microcline-perthite, muscovite, beryl, spodumene, tantalum-niobium and bismuth. The pegmatites were formed during the latter stages of the Namaqua Orogen as indicated by Rb-Sr model ages for muscovite and Pb-Pb ages from zircon which show an age of between 1000 and 950 Ma (Hugo, 1986). The time period 1200-800 Ma is characterised by the development of numerous rare-element-bearing pegmatites (see Section 4.8.2 for explanation of this pegmatite type), notably those of the Grenville province in Canada, most of the pegmatites in southern Norway, the Bihar province in India and the Georgetown and Musgrave provinces in Australia (Cerny, 1982a). In the Northern Cape there appears to be a range of pegmatites from replacement, in the case of the homogeneous types, through to emplacement, in the case of the inhomogeneous pegmatites.

4.8.2 A review of pegmatites.

Pegmatites have been classified by Ginsburg et al. (1979) according to depth of formation, mineralisation and their relations to igneous processes and metamorphic environment, and are summarised as follows:

- A) Shallow: 1.5-3,5 km. **Miarolitic pegmatites.** Intrusive into low-grade metamorphic rock. Often contain cavities bearing optical fluorite, gem-quality beryl, topaz etc.

- B) Intermediate: 3.5-7 km. **Rare-element pegmatites**. Enriched in granophile elements (Li, Rb, Cs, Be, Ta, Nb and minor Sn). Fill fractures in cordierite-amphibolite facies rocks and generated from differentiated granites which tend to be allochthonous.
- C) Great depth: 7-11 km. **Mica-bearing pegmatites**. Hosted by upper amphibolite facies metamorphic rock and commonly carry extensive mica reserves and minor, if any, REE. Products of anatexis or separated from anatectic, more or less autochthonous (S-type) granites.
- D) Extreme depth: > 11 km. Generated in granulite facies terrains. Usually barren, but locally contain allanite, monazite and corundum. They commonly grade into migmatites and have no obvious granitic source.

The above calculation of depth was based on geobarometric characteristics of the metamorphic grade of the surrounding host rock and have no direct bearing on the depth from the palaeosurface. In terms of the direct link between granites and pegmatites, the classification proposed by Ginsburg et al. (1979) indicates a decreased connection from the shallow-level pegmatites downwards. There could therefore be a continuum between the two extremes from a direct granite origin to a product of anatexis/ultrametamorphism. The classification of pegmatites according to depth of formation (geological setting) as proposed by Ginsburg et al. (1979) somewhat restricts the granite/metamorphic origin controversy which has been prevalent for many years. The following summary of Ginsburg et al.'s (1979) classification is largely based on work done by Cerny (1982b):

Miarolitic, shallow-level pegmatites are found within, or in close proximity to, granites and their direct relation is not questioned. In cases where they intrude the country rocks, they intrude them to very limited distances only. Ginsburg et al. (1979) distinguished the following four groups of granites which provided the source for miarolitic pegmatites:

- * Rapakivi and rapakivi-like granites.
- * Subalkalic, porphyritic subsolvus granites, leucocratic, biotite-bearing granites.
- * Alkalic, hypersolvus, leucocratic to alaskitic granites with riebeckite or rarely

aegirine.

* Porphyritic biotite or two-mica-bearing granites of the more or less calc-alkaline series.

The first three granites exhibit some similar characteristics and are of late orogenic affinity, the alkalic group often being anorogenic. The compositionally increasing alkalinity of the first three classes is the main feature which distinguishes them from granites parental to the other pegmatite types as defined by Ginsburg et al. (1979). The fourth granite type is compositionally similar to granites producing the rare-element-bearing pegmatites, but are intruded to shallower depths.

Rare-element pegmatites of intermediate depth are often associated with granites where the pegmatites have separated from their granitic source as a result of their intrusion into tectonically active regimes along deep fault systems (Cerny, 1982a). The genetic relationship of these pegmatites to their granitic source is often demonstrated (for example Beus, 1984, from the Aksu-Pushtiru pegmatite field in Turkestan). Many pegmatite fields of this type, however, show no exposed granitoid parent. If a metamorphic origin for the pegmatites is sought, then geochemical and petrogenic consistencies of the pegmatites from such diverse modes of origin would be unlikely. The potential granitic source could probably then still lie at depth, below the current level of exposure. Pegmatite melts are highly mobile in tectonically active regions.

Granites that generate rare-element-bearing pegmatites of intermediate depth are usually late to post-tectonic granites of calc-alkaline affinity where they are emplaced along previously formed faults and fracture systems during the waning stages of tectonic activity. These granites are typically leucocratic biotite, two mica or muscovite-bearing granites and contain accessory garnet, tourmaline, cordierite and/or andalusite.

The igneous/metamorphic origin for mica-bearing, deep-level pegmatites is still unresolved and they could be a product of both anatexis and fractionated felsic magmas at depth (Smakin & Makagon, 1972). These pegmatites often are dispersed

over large areas where granitoid plutons are lacking. Many show features described by Jahns (1955) as typical of metamorphic-anatectic origin, including the following:

- * lack of correlatable igneous rocks;
- * occurrence in high-grade terranes where differential fusion and recrystallisation has taken place;
- * correlation between pegmatite and host rock composition;
- * presence of metamorphic index minerals in pegmatites and host rocks;
- * a deficiency of rare elements usually associated with magmatic differentiation.

There are, however, some mica-bearing pegmatite fields, for example Mama-Tchuya in northern Baikal, where there is evidence for the derivation of the pegmatites from a granitoid source (Cerny, 1982b).

Granitoids that form the mica-bearing pegmatites of great depth vary considerably in composition from trondhjemites through granodiorites to granites and are mostly two-mica granites with accessory garnet, tourmaline, kyanite, zircon, apatite and monazite.

The pegmatites of maximum depth display no direct relation with granitic bodies and are considered to have formed as a product of ultrametamorphism, mostly anatexis in migmatite terrains of granulite facies (and upper amphibolite facies) metamorphism. The pegmatites and granites in these terrains show no spatial relationship with one another and are usually conformable to the host rocks, often grading into migmatites.

From the above, it can be seen that regional zoning can often be inconspicuous or even absent especially with pegmatites associated with increasing depth of formation. In general, however, pegmatitic fluids richer in volatiles would be most mobile and crystallise at lower temperatures, allowing them to migrate further into relatively shallower levels of emplacement along the lithostatic pressure gradients. According to Cerny (1982a), reactions of pegmatitic fluids with wall rocks are limited, even in the most reactive, volatile-rich pegmatites and it is often difficult to distinguish between contaminants from the wall-rocks and the initial source fluids.

The pegmatites in the area under investigation are predominantly of the rare-element-

bearing type. Cerny (1982a) noted that pegmatites are often emplaced following a period of regional metamorphism either in a retrogressing environment, in the case of mica-bearing pegmatites, or during a period affected by granitic intrusions, where the thermal effect is high, in the case of the rare-element-bearing pegmatites.

4.8.2.1 Evolution of granites and the formation of rare-element pegmatites.

According to Cerny (1982b), only a small portion of a calc-alkaline batholith is responsible for the formation of mineralised pegmatites. Intrusive batholiths may undergo gradual differentiation as is demonstrated by Jankowitz (1987) on the relatively small Cnydas Granite, where later, more evolved phases showed a propensity towards S-type granites as defined by Chappell & White (1974). The degree of fractionation of an intruding batholith, or of anatexis of supracrustal sequences, would have an effect on its capacity for pegmatite formation. The compositional variation of major elements within different facies of a single intrusion is rather limited compared to more dramatic variations of the trace elements, such as increases in Li, Rb, Cs, Be, Ga, Sn, Mn, Y and depletion of Mg, (Fe), (Ca), Sr, Ba, Ti and Zr (Blockley, 1980; Cerny et al., 1981). LREE and HREE are also subject to change with depletion of the former and enrichment of the latter, with a pronounced negative Eu anomaly in the early stages followed by a general decrease in all REE during the pegmatitic stage (Cerny, 1982b).

Most granitoids parental to rare-element-bearing pegmatites are difficult to evaluate because they have been intensively modified by processes other than fractionation and host rock reactions (Cerny et al., 1981; Goad & Cerny, 1981; Longstaffe et al., 1981). Cerny (1982b) summarised this as follows: "it appears that the 'fertile' granites are genetically separate members of batholith complexes, rather than products of their bulk differentiation".

The depletion of REE can be explained by removal of allanite and/or monazite (Miller & Mittlefehldt, 1982). The changes observed in some of the other trace elements have been proposed by Cerny (1982) to be related to thermo-gravitational

convection diffusion (Shaw et al., 1976; Hildreth, 1979). Repeated pulses of pegmatitic fluids leaving the diffusion column during the process of diffusion would result in the pegmatites becoming more fractionated with time and the accompanied increase in volatiles would allow for the pegmatites to migrate further from their source. This process would allow for the regional zonation often observed around the granitic source (Cerny, 1982b).

The geochemical signature of twenty fertile granites, with respect to rare-element pegmatites, were investigated by Cerny (1991). He divided these into two groups, namely, those containing Li, Rb, Cs, Be, Ga, Sn, Nb < Ta, B, P, F (LCT) and Nb > Ta, Y, REE, Sc, Ti, Zr, Be, Th, U, F (NYF). Conclusions drawn from Cerny's (1991) research are summarised in Table 4.8.

Table 4.8: Characteristics of LCT and NYF-bearing granites.

LCT GRANITES	NYF GRANITES
Tendency toward syn to late-tectonic origin (although poor correlation).	Tendency toward late to anorogenic origin (although poor correlation).
Largely, though not exclusively associated with peraluminous granites.	Associated with metaluminous to peralkaline granites.
Derived from undepleted upper-crustal lithologies suffering first anatectic melt. Protoliths to this melt are supracrustal sequences and ortho and para-lithologies of their basement.	Derived from melting of depleted lower-crustal sources, residual after preceding anatectic events (A-type granites), or melting of juvenile lithologies with short crustal residence.

After Cerny (1991)

The generation of the LCT-bearing systems are the result of these rare-elements being highly volatile and easily mobilised (Cerny, 1991). They act as fluxes depressing the solidus of the magmas after being extracted into the first melts from progressively metamorphosed lithologies and subsequently enhance fractionation. The NYF-bearing systems are generally depleted in volatiles, F probably being the main volatile component, and are thought to be derived from a second melting event of middle to lower-crustal, high pressure lithologies which were dehydrated and generally depleted during the previous anatectic event (Cerny, 1991). A second potential source for the NYF-bearing system proposed by Cerny (1991) is a combination of direct differentiation of a new mantle-derived basic magma which provides the fluids necessary for anatexis of the lower crust.

In a mixed NYF-LCT-system Cerny (1991) proposed three possible alternatives, namely: "the crust could have been partially depleted, or the anatexis affected a mixed range of depleted plus undepleted protoliths, or a pristine NYF magma from depleted crust became contaminated by digestion of undepleted lithologies".

4.8.3 Distribution of pegmatite types and their commodities in the Upington-Kenhardt pegmatite belt.

A vast majority of the homogeneous pegmatites in the area under investigation are barren, and are of little economic significance containing mainly quartz, feldspar (albite-oligoclase, microcline-perthite) and some accessory minerals (muscovite, biotite, schorl, magnetite, ilmenite and zircon). All rare-element-bearing pegmatites and those containing mineable concentrations of other minerals such as feldspar, are considered to be of the inhomogeneous type. Hugo (1969) pointed out that of the approximately 11 000 pegmatites in an area of similar extent to the one currently under investigation, about 10 000 of these are inhomogeneous. For practical reasons, only those pegmatites containing reasonable concentrations of commodity related minerals, or visible mineral constituents, were investigated during the metallogenic mapping program. Although an attempt was made to investigate as many pegmatites as possible, not all of them were investigated. In addition, many pegmatites remain unexposed (covered by surficial deposits, at depth or lack of trenching). The following analysis of the pegmatites could be considered to be representative of the pegmatites as a whole and is concerned largely with the inhomogeneous (simple and complex) pegmatites.

The inhomogeneous pegmatites are, apart from a few exceptions, discordant with respect to their host rocks. They have been interpreted by Hugo (1969) to be of emplacement origin. The simple pegmatites commonly consist of quartz, perthite and albite-oligoclase and show a well developed internal zoning of no more than four zones. The border and wall zones contain essentially microcline-perthite, quartz and plagioclase in varying proportions and grain sizes. The intermediate zone contains perthite with plagioclase in subordinate amounts and accessory andalusite and rare-earth minerals. The cores of these pegmatites are composed of quartz. REE-bearing

minerals are usually concentrated with perthite along the outer margins of the intermediate zone or cores. Biotite is also a common constituent in these pegmatites and is found in the border, wall and intermediate zones. Andalusite occurs only in the simple pegmatites, whereas tourmaline (mostly schorl) occurs mostly in the homogenous and complex pegmatite varieties. The quartz core of simple pegmatites containing REE-bearing minerals often has a pinkish tinge (Hugo, 1969). REE-bearing minerals are largely, though not exclusively, confined to the simple pegmatites

The complex pegmatites are generally larger and more irregular in shape than the simple varieties and commonly contain three or more zones. The border and wall zones usually consist of plagioclase, quartz, perthite, muscovite with minor schorl, biotite, garnet and magnetite. The intermediate zone commonly contains albite (mostly cleavandite), quartz and muscovite with accessory amounts of beryl, perthite, apatite and rarely schorl, topaz, triplite, zircon and REE-bearing minerals. Additional intermediate zones are often present, and contain *inter alia*, quartz, cleavandite and muscovite with accessory beryl, spodumene and columbite-tantalite with replacement bodies of lithia mica, cleavandite and quartz, or albite with accessory cassiterite, schorl, garnet, quartz and muscovite. Muscovite is far more abundant in the complex pegmatites than in the simple pegmatites, where this mineral is of minor importance.

From Fig. 4.15 it can be seen that the pre-tectonic supracrustal rocks host the majority

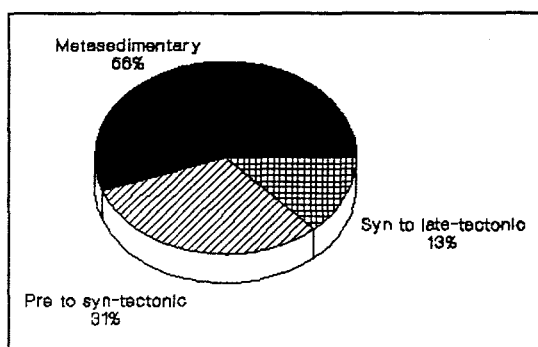


Figure 4.15: Pie graph showing the percentage of pegmatites within metasedimentary sequences, pre to syn-tectonic intrusives and syn to post-tectonic intrusives.

of the pegmatites in the area. There is, however, no discernible preference of pegmatite development in any particular hostrock. The amount of pegmatites within the various formations are depicted on the graphs in appendix B and C and the distribution of these pegmatites according to commodity is shown on maps in appendix D.

Pegmatite distribution is wide spread along the western part of the Kakamas Terrane

but broadly consists of various clusters or groups as defined by Cerny (1982a). These groups are termed the Riemvasmaak, Kakamas, Middel Post, N'Rougas, Wolfkop, Steynspuits and Rok Optel Groups (Fig. 4.16). The groups of N'Rougas, Wolfkop, Steynspuits and Rok Optel are not as well defined and are situated to the east of Kenhardt. In the northern portion of the pegmatite belt the groups are well defined where numerous pegmatites are highly concentrated in relatively small areas.

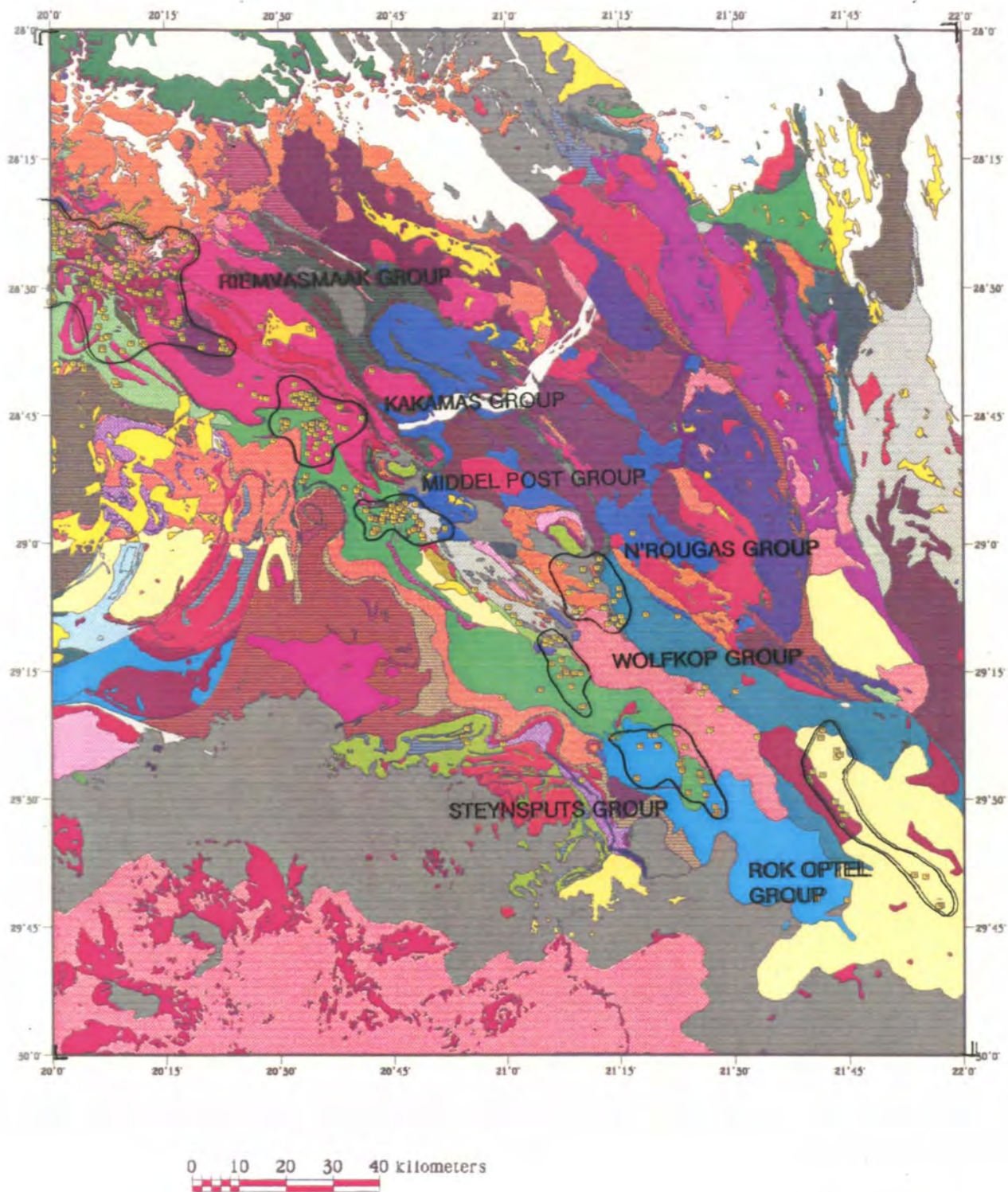


Figure 4.16: Distribution of all pegmatites in the area under investigation and the localities of the various pegmatite groups.

Beryl: Beryl occurs in almost all types of pegmatites except in simple, andalusite-bearing pegmatites (Hugo, 1969). It is wide spread throughout the belt and present in all the pegmatite groups although only three occur in the Steynsputs group. Although all the pegmatite-bearing formations in the Kakamas Terrane contain beryl-bearing pegmatites, the highest concentration of beryl-bearing pegmatites is in the Riemvasmaak Gneiss and Kenhardt Migmatites. Beryl-bearing pegmatites are therefore not generally exclusive to any particular formation.

Rare earth element-bearing minerals (REE): REE-bearing pegmatites are characteristic of simple pegmatites, although Hugo (1969) noted that they occasionally occur in complex pegmatites. The majority of these pegmatites are situated in the Riemvasmaak and Kakamas groups while some occur in the Steynsputs group. The Middel Post, N'Rougas, Wolfkop, and Rok Optel groups are virtually barren of REE-bearing pegmatites. These pegmatites are abundant in the pre to syn-tectonic intrusives, especially the Donkieboud Granite Gneiss and Riemvasmaak Gneiss. Generally, however, a wide variety of pegmatite-bearing formations, including three in the Bushmanland Subprovince, host REE-bearing pegmatites.

Mica: Mica-bearing pegmatites occur in all the various types of pegmatites, although rarely in the simple pegmatites. The figure in appendix D showing the distribution of mica-bearing pegmatites only shows the deposits that were mined for mica and is therefore not a true reflection of the distribution of mica-bearing pegmatites.

Lithium: Lithium-bearing pegmatites occur in complex pegmatites and are confined to the Rok Optel group and associated with the syn to late-tectonic Rok Optel Granite of the Keimoes Suite. Hugo (1969), however, noted two further pegmatites in the N'Rougas group which contained spodumene and they are the only pegmatites to host cassiterite.

Phosphate: Phosphate, in the form of apatite occurs in nearly all complex and poorly zoned (homogeneous) pegmatites (Hugo, 1969). Apatite-bearing pegmatites are found particularly in the northern portion of the pegmatite belt (Riemvasmaak group) but

are also sparsely scattered throughout the belt. The Twakputs Formation contains the highest proportion of apatite-bearing pegmatites followed by the Witwater Gneiss and the Riemvasmaak Gneiss.

Tantalum/Niobium: Tantalum/niobium is present in the columbite-tantalite series of minerals and occurs in complex pegmatites. These are found exclusively in the Rok Optel group and is associated with the Rok Optel Granite. According to Hugo (1969) the Nb:Ta (Nb > Ta) ratio remains fairly constant from pegmatite to pegmatite.

Andalusite: Andalusite-bearing pegmatites are all of the simple type (Hugo 1969) and are concentrated mainly in the northern portion of the pegmatite belt in the Riemvasmaak, Kakamas and Middel Post groups. Other occurrences are sparsely distributed throughout the pegmatite belt. A few isolated andalusite-bearing pegmatites also occur in the Bushmanland Subprovince. The Kenhardt Migmatite and the Twakputs Formation are host to the majority of these pegmatites while the Donkieboud Granite Gneiss contains most of these pegmatites among the intrusive rocks. These pegmatites do not occur in any syn to late-tectonic intrusive rocks.

Fluorite: Fluorite-bearing pegmatites are concentrated mainly in the northern portion of the pegmatite belt. These pegmatites are found almost exclusively in the Riemvasmaak and Kakamas groups. The majority of the fluorite-bearing pegmatites are hosted by pre to syn-tectonic intrusives, namely the Riemvasmaak Gneiss and Augrabies Gneiss.

Rose Quartz: Rose quartz-bearing pegmatites are rather evenly distributed throughout the western portion of the pegmatite belt. The only groups that do not appear to contain rose quartz-bearing pegmatites are the N'Rougas and Rok Optel groups. The Kenhardt Migmatite, Riemvasmaak Gneiss and unnamed basic rocks are host to the majority of these pegmatites.

Corundum: A few scattered corundum-bearing pegmatites are situated almost exclusively in the northern portion of the pegmatite belt, mainly in the Riemvasmaak

and Kakamas groups. They are found in simple pegmatites and are usually associated with andalusite-bearing pegmatites (Hugo, 1969). Apart from the unnamed basic rocks, these pegmatites are hosted by metasediments.

Magnetite: Magnetite-bearing pegmatites are widespread throughout the pegmatite belt and according to Hugo (1969), are most abundant in the homogeneous and REE-bearing simple pegmatites.

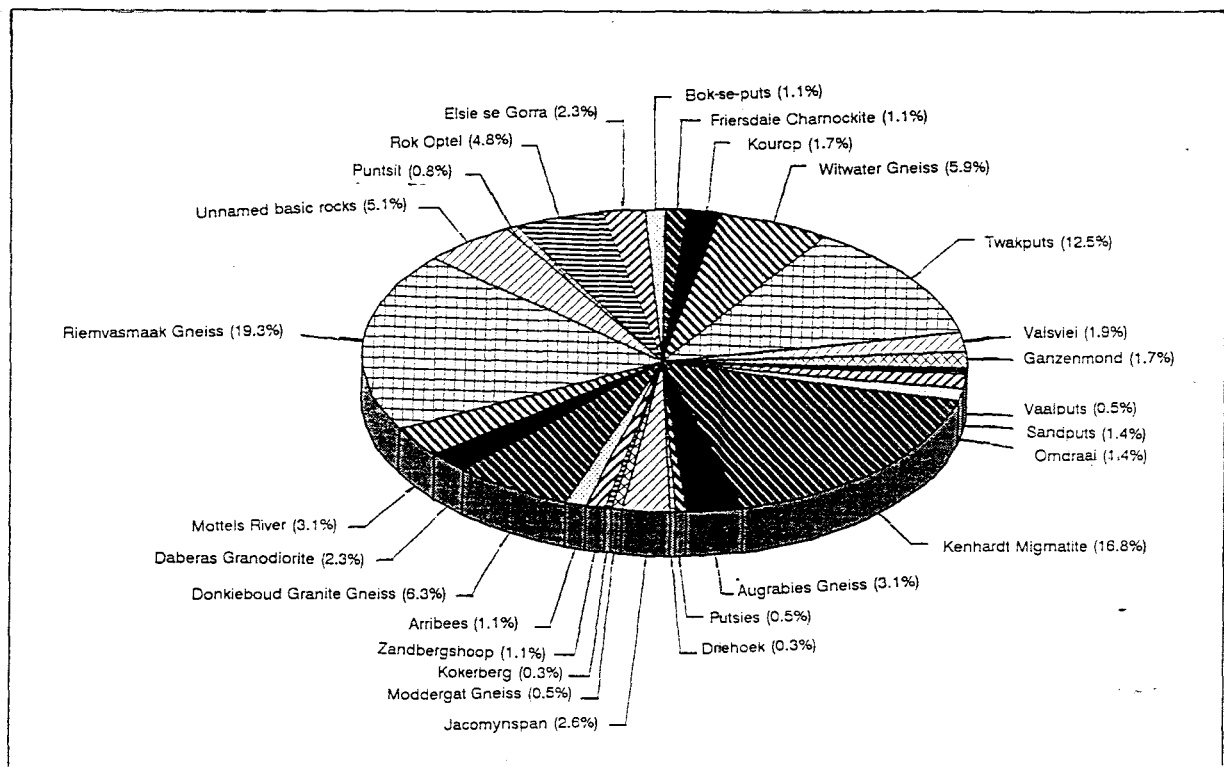


Figure 4.17: Pie graph showing the percentage of pegmatites within the various pegmatite-bearing formations. Further details of what each formation contains with respect to pegmatites are shown in appendix C.

If a comparison is made between Figure 4.16 and Figure 4.17, it can be seen that there is a broad correlation between the surface area exposure of the formations and the amount of pegmatites contained within them. For example, the Riemvasmaak Gneiss (pre to syn-tectonic intrusive) and the Kenhardt Migmatite (mainly metasediments) contain the highest percentage of pegmatites and also cover a large percentage of the surface area within the pegmatite belt. There are, furthermore, large areas within these formations which do not contain significant concentrations of pegmatites. Exceptions to this generalisation are, for example, the Twakputs Formation (metasedimentary) and Unnamed basic rocks (syn to late-tectonic intrusive) which contain large numbers of pegmatites relative to their surface area, and the Friersdale Charnockite which contains few pegmatites relative to its surface area. The Rok Optel Granite is the only rock type to exclusively contain specific commodities (Ta/Nb and Li). The reason for this is unclear, as no research has been conducted on this granite. It can therefore be generalised that there is little correlation between pegmatites, their contents, and the formations which host them.

4.8.4 Discussion.

From Figure 4.18 it can be seen that the homogeneous pegmatites are largely concentrated in the northern portion of the pegmatite belt in the Kakamas and Riemvasmaak areas. These homogeneous pegmatites appear only in the Riemvasmaak, Kakamas and Middel Post groups. The distribution of the inhomogeneous pegmatites are fairly evenly spread throughout the pegmatite belt, although the N'Rougas, Wolfkop, Steynspuits and Rok Optel groups are exclusively inhomogeneous. It has been accepted previously, that the inhomogeneous pegmatites are a reflection of greater distance from their source due to higher levels of volatiles which would decrease their solidus temperature and increase their mobility. In other words, the most fractionated pegmatites are located furthest away from the granitic source. This phenomena is demonstrated by Trueman & Cerny (1982). It is therefore likely that the granitic source emplaced at higher levels in the crust in the northern portion of the belt. This could also point to the possibility that the southern portion of the Kakamas Terrane is representative of a deeper crustal level. Trueman & Cerny (1982) pointed out that

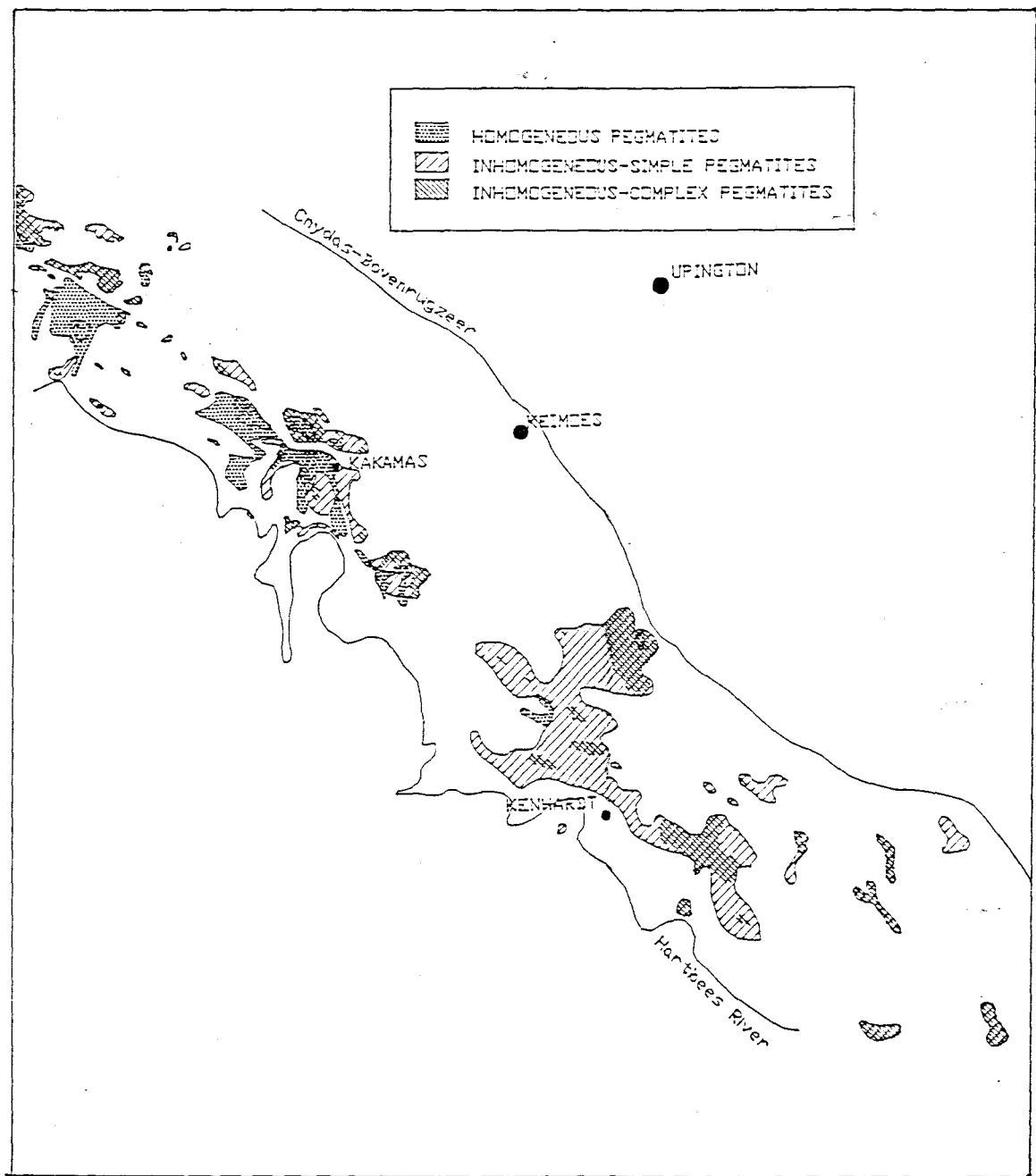


Figure 4.18: Distribution of homogeneous, simple and complex pegmatites within the area under investigation (after Hugo, 1969).

Ta/Nb and Li-bearing pegmatites are developed further from the source granite. In the Upington-Kenhardt area Ta/Nb and Li is only present in pegmatites in the southern portion of the pegmatite belt, which is a further indication of the possibility that the source granites lie at greater depth in the south. The REE-bearing pegmatites are predominantly simple pegmatites reflecting an intermediate level of intrusion. These pegmatites are characteristic of the northern portion of the pegmatite belt and may also indicate a shallower level of granitic intrusion in the north.

The pegmatites in the area under investigation contain both LCT (Li, Be, Sn, B, P and F) and NYF (Nb>Ta, REE, Ti, Zr, Be, Th, U, and F) geochemical signatures as defined by Cerny (1991) and it would seem that some of the LCT mineral assemblages crystallised in the more fractionated pegmatites. If Cerny's (1991) model of a mantle-derived NYF granitic magma is accepted, then some form of extraction of LCT-elements must be sought. Although only speculative, contamination from undepleted crustal material must have occurred (see Section 4.8.2.1) at deeper levels.

The various pegmatite groups may be a manifestation of specific pegmatite generating granites at depth, although no obvious zoning exists within the groups themselves. A more detailed mineralogical and geochemical study of the pegmatite groups along the lines suggested by Trueman & Cerny (1982) would probably reveal such zoning.

4.9 DISCUSSION ON THE MINERALISATION RELATED TO FELSIC MAGMATISM (IN THE KAKAMAS TERRANE).

The overwhelming abundance of syn to late-tectonic granitoid intrusions of the Keimoes Suite as well as the presence of numerous mineral deposits, traditionally related to granitoids, within the Kakamas Terrane is certainly not coincidental. The styles of mineralisation in the Kakamas Terrane change abruptly over the Cnydas-Bovenrugzeer shear zone from vein-related deposits in the east, to pegmatite deposits to the west of this shear zone. In Figure 4.19 it can be seen that these two contrasting styles of mineralisation reflect two differing depths of formation, the pegmatites having formed at much greater depths than the vein-type deposits. As suggested by Varlamoff (1978), the accumulation of residual magmas and fluids are only possible during slow

cooling in a more or less closed system at great depths. Under conditions of shallower depths the magma chamber cools more rapidly, which provides insufficient time for the accumulation of enough volatiles in the residual fluid for the formation of pegmatites, while Sn and W may still separate in sufficient quantities and intrude into the country rocks. Humphreys & Van Bever Donker showed differing tectonic depths across the Bovenrugzeer-Cnydas shear of 1.6-2.4 Kb compared to 3.5 Kb to the west of the shear zone.

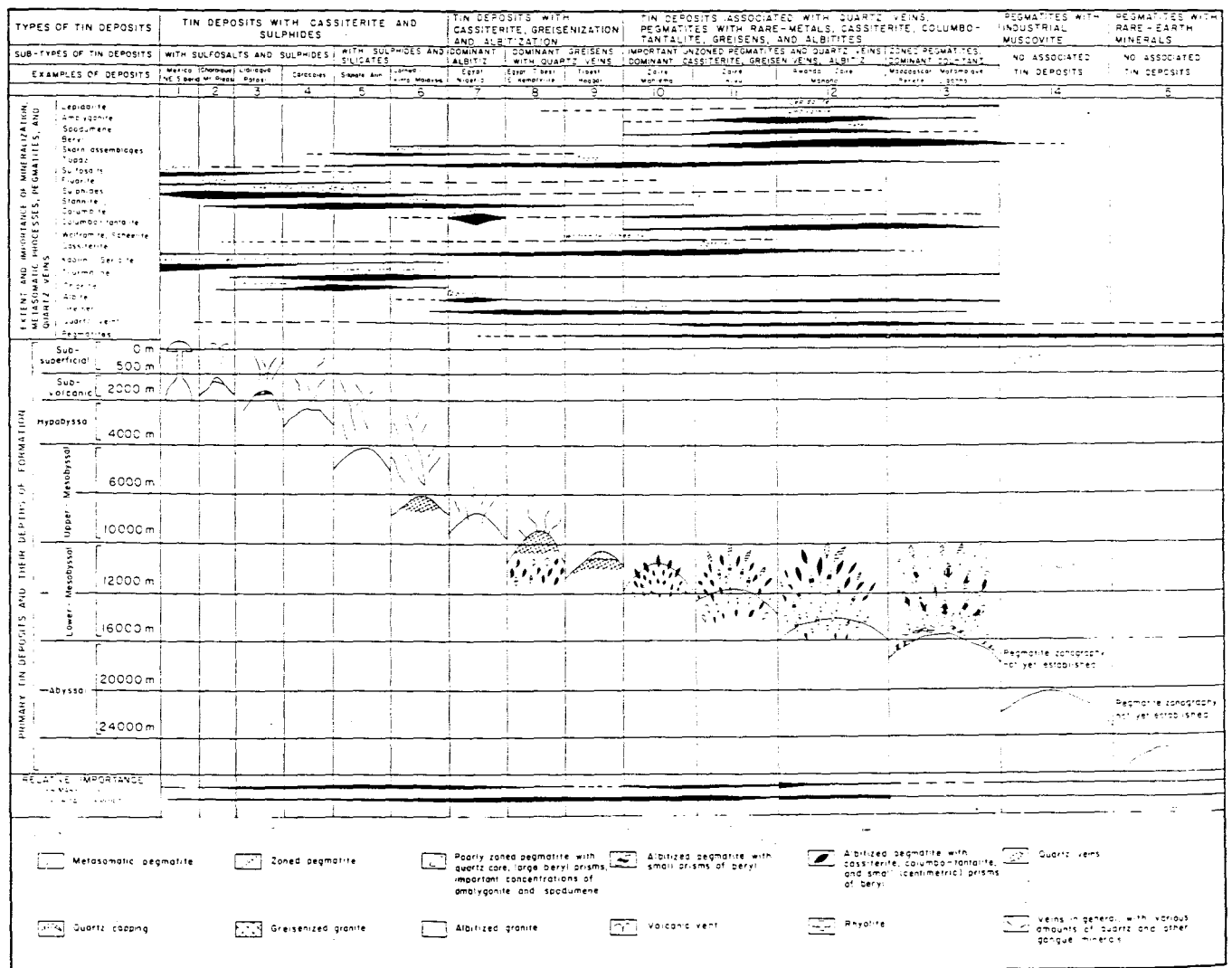


Figure 4.19: Figure showing the various styles of granite-related mineralisation and their mineralisation in relation to depth of formation (from Strong, 1990).

Geringer et al. (1988) indicated that there is an increase in the potassium index of the Keimoes Suite granitoids from east to west which has been interpreted as a manifestation of deeper katazonal granitoids in the west and shallower epizonal granitoids in the east. It was further suggested by these authors that this might represent differing crustal levels, with deeper crustal levels in the west and higher-level intrusions in the east. This general trend may indicate a continuation of even deeper level granites, not necessarily associated with the Keimoes Suite *sensu stricto*, which produced the pegmatites.

These observations of Geringer et al. (1988) and Humphreys & Van Bever Donker (1990) correspond well with shallower-level vein-type mineralisation to the east of the Cnydas-Bovenrugzeer Shear and deeper-level pegmatite mineralisation to the west. From this evidence, one would like to deduce that the Cnydas-Bovenrugzeer Shear could possibly represent a line of differing crustal levels. Geringer & Botha (1977) did not, however, find a distinct variation in metamorphic grade over this shear zone. The regional structure, however, changes remarkably from a rather simple fold pattern northeast of the shear to a complex superimposed fold system southwest of the shear (Geringer & Botha, 1977).

Geringer et al. (1988) pointed out that the geochemistry of the Keimoes Suite granitoids resembles that of a destructive margin-type granite with a tendency towards within-plate granites. If a subduction zone-type setting is accepted for the Areachap Terrane, further subsidence of the colliding continental plate with the Kaapvaal Craton, following the closure or choking of the "Areachap ocean", could have produced the dominantly I-type granites of the Keimoes Suite as proposed by Geringer et al. (1988) as well as the mantle-derived, contaminated granites necessary to produce the pegmatites. Underriding of a continental plate may result in back-thrusting, exposing various levels of the underriding crust as it was being "scraped off" in the process of crustal thickening.

4.10 THE KORAS GROUP.

The 1260-1050 Ma (Botha et al., 1979) Koras Group rocks (Fig. 4.7) consist of interbedded mafic and felsic lavas with sedimentary sequences which were deposited in three separate graben or half-graben structures in the Upington Terrane along the eastern margin of the Namaqua Tectonic Province. The angular nature of the clasts within conglomerates and the immature character of the clastic detritus is indicative of a local source for these sediments. The grabens are associated with the Brakbosch and Blaauwbospan shear zones (Fig. 4.7) which are large right lateral shear zones formed during the waning stages of the Namaqua event (Moen, 1987). These deep-seated faults caused structural weaknesses which controlled the outpourings of the volcanic sequences.

The volcanic rocks vary in composition from basalt through K-rich andesitic dacite, dacite, rhyodacite and rhyolite, forming part of a calc-alkaline suite (Grobler et al., 1977). The more mafic members of the volcanic sequences show some shoshonitic affinities. Moen (1987) suggested that the basic magma was derived from the mantle, while the abundant felsic magma was derived from partial melting of the crust. Assimilation of crustal material with basaltic magma produced a wide spectrum of intermediate rocks.

Mineralisation in the form of malachite staining and chrysocolla occurs mainly within faults, fold hinges and volcanic breccia, associated with the basic lava units (Fig. 4.7). It is therefore possible that there were two periods of mineralisation, one related to the syn-tectonic volcanic breccia and the other related to a later phase of influx of metal-carrying fluids, possibly related to post-tectonic felsic intrusives related to the Koras Group (Blauwbosch Granite and Rooiputs Granophyre, Fig. 2.6). In the latter case, the fluids could either have been derived directly from the felsic intrusives, or the intrusives could have provided a heat source, creating a convective cell which resulted in mineralised fluids escaping along structural weaknesses such as faults and fold hinges. The high heat flow provided by the convective cell would then promote leaching of metals from the underlying rocks within the basin. Basalts generally

contain high background concentrations of Cu and other metals such as Ag, Co and Au from which the circulating fluids may have leached the metal as proposed by Jolly (1974). Bischoff & Seyfried (1978) showed that trapped sea-water within basalt becomes increasingly acidic at temperatures exceeding 300°C and, under these conditions, is able to leach metals from the basalt. Other possible sources may have been basin sediments (White, 1971) or buried basement. On the basis of average metal content in various rock types, Wedepohl (1974) concluded that almost all rock types are able to produce ore forming concentrations as a result of persistent leaching by circulating acid solutions.

The precipitation of the metals requires a sulphur source and a reducing depositional environment. Situations which may provide these conditions are: pre-existing pyritic beds, organic-rich beds, mixing of reduced metal-bearing fluids with sulphate-bearing meteoric water, sulphate-reducing bacteria, evaporite beds and the introduction of hydrocarbon reductants (Maiden et al., 1986). There is little evidence of these requirements being present in the Koras Basin, and along with the limited size of the basin, from which the metals could have been leached, could account for the restricted nature of the deposits.

Alternatively, the triviality of these occurrences may indicate that these deposits are merely related to heated formational waters which locally concentrated the ore-forming elements during structural events. The Koras Group has been correlated with the Sinclair-Ghansi series of fault-bounded sequences in Namibia and Botswana (Borg, 1988). Borg (1988) suggested that these sequences were deposited in a Kibaran-aged extensional rift system that propagated northwards, while Jacobs et al. (1993) argued for a series of strike-slip pull-apart basins, formed during the post-tectonic transpressional shearing which is represented, among others, by the Brakbosch Fault in the area under investigation. In these sequences, Borg & Maiden (1986) indicated that Cu mineralisation occurs locally in highly altered basalts, and Au mineralisation is hosted by felsic porphyries. Stratabound Cu-Ag-Au-(Pb)-(Zn)-Mo-Co deposits are hosted by fine clastic sediments. The mineralisation occurs in pyritic slate, metasiltstone, pyritic sandstone, calcareous slate and detrital and algal limestone. A

diagenetic origin of the mineralisation is proposed by Borg & Maiden (1986) where fluids, guided by basement faults, precipitated the metals in favourable host rocks after leaching the metals out of the volcanic rocks. This scenario may well have been present in the Koras Group.

The known mineralisation is confined to the northern domain as defined by Moen (1987). The stratigraphy of the three domains can all be correlated to some extent which led Moen (1987) to deduce that the sequence of events was similar in all three domains. It is therefore highly likely that similar mineralisation could be found in the central and southern domains, particularly along faults and fold hinges. Further targets for exploration may be relatively permeable rocks along which metalliferous fluids may have been transported and deposited, as well as oxidation/reduction interfaces which may have formed chemical trap sites for the precipitation of the metals.

4.11 URANIUM.

The discovery of the Yeelirrie deposit in Australia (Haycraft, 1976; Cameron, 1990) in 1972 resulted in renewed interest in the so-called Calcrete-type of uranium deposit (Dahlkamp, 1978). According to Dahlkamp (1978), these deposits form in arid climatic regions where the uranium is concentrated in irregular lenticular forms within flat channels which have cut into Archaean basement rocks and which subsequently filled with clay, sand and calcrete. The uranium usually occurs as a uranyl-vanadate called carnotite. The term "calcrete-type" is considered by Toens & Hambleton-Jones (1980) to be misleading as many of these deposits occur in fluvial sediments ranging from boulder beds to silts which may or may not be highly calcified or be interbedded or overlain by pedogenic deposits such as calcretes, dolocretes or gypcretes. These authors preferred to name these types of uranium deposits "surficial uraniferous deposits", and subdivided them into fluvial, lacustrine/playa and pedogenic. These terms are not necessary absolute and may overlap somewhat. The following is a summary of this classification scheme as devised by Toens & Hambleton-Jones (1980):

Fluvial-type deposits may vary according to the gradient of the fluvial system and

are represented in Figure 4.20. The valley-fill-type has the highest economic potential and is deposited in a high energy regime under flash-flood conditions. The flood-plain-type either overlies or occurs downstream of the valley-fill-type deposits in a slightly lower energy environment but the composition of the sediments is essentially the same. Lateral extent of the ore body is more extensive. Further downstream the deltaic-type deposit is formed as the flood plain enters a playa lake. Sediments are finer grained and are often interbedded with lacustrine deposits.

Lacustrine/playa-type deposits are either topographically controlled or drainage controlled as a result of choked drainage channels. Sediments within these pans are usually clay and silt which have been cemented by calcite, gypsum and salt with associated hypersaline water. Evaporation is the dominant process.

Pedogenic deposits are generally considered to consist of the various forms of "cretos" such as calcrete, gypcrete, dolocrete, etc. Authigenic pedogenic uranium deposits overlie decomposed basement rocks and are considered to have formed entirely *in situ* and may have been cemented by calcite and gypsum with associated carnotite. Allogenic pedogenic uranium deposits occur in "cretos" having been transported in their host material to their present position.

Valley-fill-type deposits, such as the Langer Heinrich deposit (Hambleton-Jones, 1976) in Namibia, are a result of uplift along the escarpment during the mid-Tertiary which caused incision and erosion of the old African erosion surface (Hambleton-Jones et al., 1986). During periods of heavy rainfall the deeply incised valleys were filled and choked forming potential uranium ore bodies. In the area under investigation, there is no known deposit of this type as most of the surface is a peneplain with no deeply incised paleo-valleys. The only area amenable to the valley-fill-type deposits is in the northeast of the area along tributaries of the Molopo and Bak Rivers.

Under slightly lower energy conditions choking of the river valleys or flood plains prevailed. The Brulkolk uranium deposit is of this type where a paleo-river system has been choked with red alluvium. The uranium, in the form of carnotite, is associated

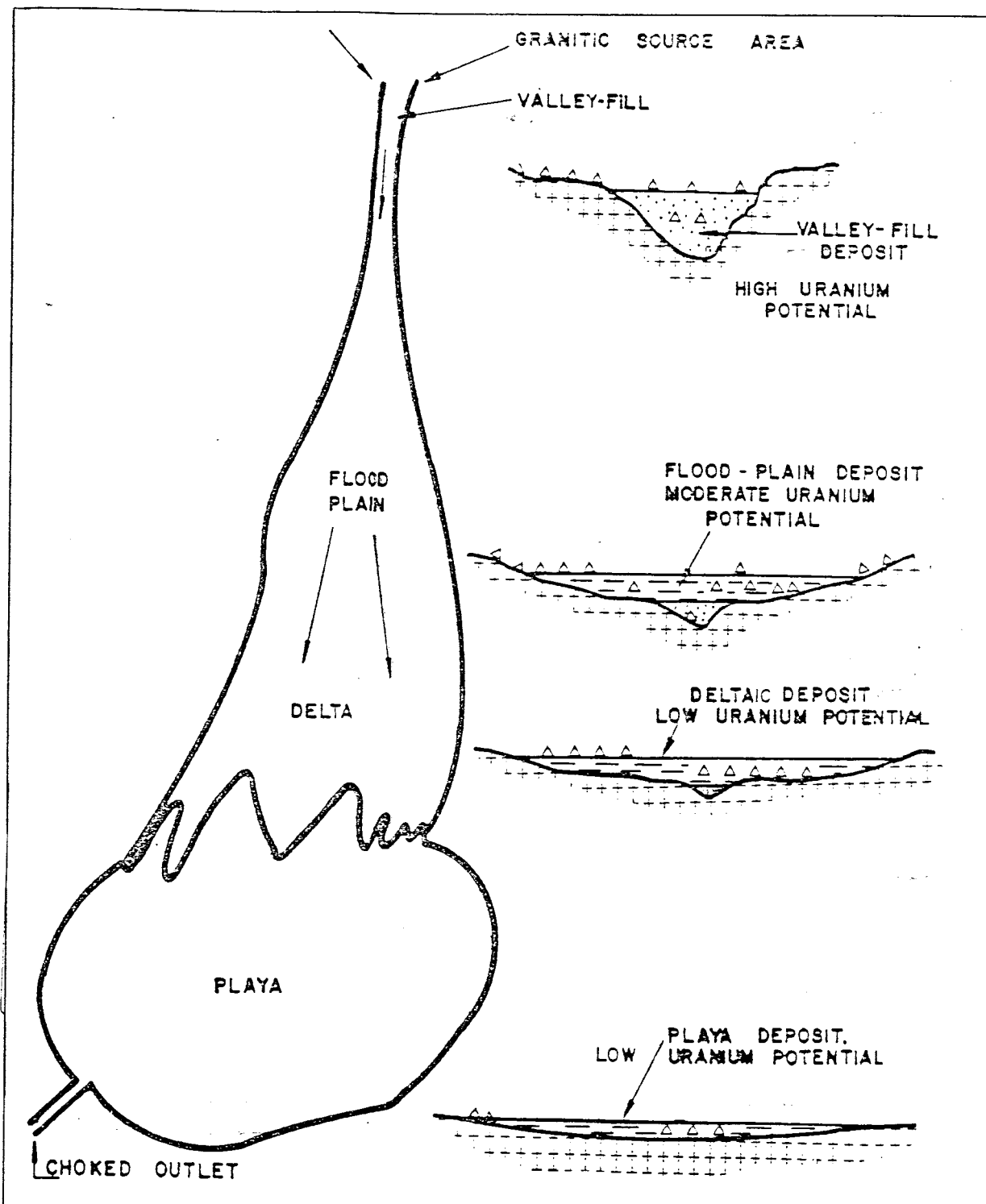


Figure 4.20: Schematic representation of the location of the fluvial-type surficial uranium deposits and their economic potential (from Toens & Hambleton-Jones, 1980).

with gypsum which precipitated in the red alluvium. The present surface is composed of red fluvial sands and gravels. A similar occurrence to the Brulkolk uranium deposit is the Kleinbegin uranium deposit where carnotite is found within a gypsiferous sand in a stream valley.

Peneplanation of the land surface caused ponding and the formation of pans with the development of lacustrine/playa-type surficial uranium deposits, of which the Geelvloer uranium-gypsum deposit is an example. Carnotite is found mainly within weathered Dwyka shale, basement granitic material and red aeolian sand which often contains calcareous nodules and is cemented by calcium carbonate in the upper layers (Hambleton-Jones et al., 1986). The mineralisation is intimately associated with the contact between the basement granites, Dwyka shales and porous sand overlying fractures in the granites. The carnotite is also found to a lesser degree in the pan sediments themselves.

The McTaggart's Camp and Dyasons Klip as well as the Arribees & Osvlei uranium occurrences are all intimately associated with calcretes and can be classified as pedogenic-type deposits. At McTaggart's Camp and Dyasons Klip they contain both authigenic and allogenic components whereas, at Arribees & Osvlei they probably represent allogenic pedogenic precipitates (Treasure, 1977, see Section 3.10.3).

These deposits have been preserved due to the prevailing dry climate. Wetter climates would result in erosion of the sediments and/or dissolution of the carnotite (Toens & Hambleton-Jones, 1980). The preservation potential of the ore body, in this regard, should be borne in mind when exploration for these deposits are carried out. Exploitation would be relatively cheap due to the surficial nature of the ore bodies. In the area under investigation, there are no true valley-fill-type deposits such as Yeelirrie in Australia and Langer Heinrich in Namibia. These deposits have, according to Toens & Hambleton-Jones (1980), the highest economic potential of all the above-mentioned deposit types. This would suggest that there is little chance of finding a large uranium deposit in the area under investigation, and only modest deposits of the flood plain, lacustrine/playa and pedogenic-type are probable.

The ultimate source of the uranium is probably from certain granites in the area, although precisely which granites are potential sources is not known. Airborne radiometric surveys only cover the central portion of the area under investigation between the 28°30" and 29°15" south latitudes. These indicate that the Riemvasmaak Formation shows the largest anomalies, providing a potential source for valley-fill-type deposits in the deeply incised valleys in the Riemvasmaak area or pedogenic deposits overlying the Riemvasmaak Formation in the area north of Kakamas. The De Bakken Granite is a further possible source where Frick (1986) has shown leaching of considerable amounts of uranium (see Section 3.6.1.). In the area surrounding the De Bakken Granite, pedogenic or flood-plain-type precipitation maybe found. In the area surrounding the Kleinbegin occurrence, no radiometric anomaly is indicated.

4.12 GYPSUM.

Gypsum deposits are found in arid regions where the mean annual potential evaporation exceeds the mean annual precipitation. Only rarely are gypsum crusts developed in areas where the mean annual rainfall exceeds 250 mm. Gypsum is highly mobile, which means that if the moisture availability exceeds the evaporation rate at any given time, gypsum will be dissolved and transported into the soil zone, and ultimately into the groundwater if the soil moisture capacity is exceeded.

Gypsum in the area under investigation is present in a variety of forms from horizontally bedded gypsum with or without associated pans to gypsum associated with river beds. The gypsum deposits in the Aries area, in the north-western extremities of the area under investigation (Fig. 3.5), comprise pan-type evaporitic deposits as well as subsurface gypsum crusts related to lateral movement of near-surface groundwaters, where the deposits lie at about 9m above the level of the pan to the east. The gypsum was derived from the surrounding Nama sediments (see Section 3.7.1).

To the south of the study area, in the Brandvlei region, as well as in parts of Verneukpan, in the southern portion of the area under investigation, there are numerous, extensive areas underlain by near-surface gypsum layers. The gypsum

consists of hard crystalline, nodular crystalline and powder varieties. The gypsum powder probably represents residual material from an older subsurface crystalline gypsum horizon. These deposits are related to peneplains which are well developed in this region.

In the Rietput deposit, gypsum is deposited on the banks of the Sout River in the form of selenite crystals as well as powder gypsum which grades into a hard crystalline variety. Much of the gypsum in this deposit has a high clay and gravel content.

In conclusion, gypsum is not confined to a specific environment, although the deposits are generally confined to the northern and southern parts of the area under investigation, close to, or within the areas covered by the Nama and Karoo sediments. It is therefore highly probable that these rocks provide the source for the calcium and sulphate which is essential for the development of gypsum, as suggested by Visser et al. (1963).

Chapter Five

5. PLATE TECTONIC MODEL.

5.1 INTRODUCTION.

Several tectonic models have been proposed by various authors to synthesise the geological observations that have been made. Some of these are briefly summarised below.

5.1.1 Mantle plume and/or Rift models.

Kröner (1979), Botha & Grobler (1979) and Van Zyl (1981) proposed models for an ensialic and intracontinental mobile belt on which lateral spreading of a hot asthenosphere would result in rifting, heating of the lower crust due to subsidence, and the generation of magma. Botha & Grobler (1979) suggest that the evolution of the eastern Namaqua front occurred in four stages, commencing with the rise of a mantle plume at the edge of the Kaapvaal Craton which caused doming and rifting. This first stage resulted in the deposition of the Matsap, Kheis and certain Namaqualand volcano-sedimentary sequences. The second stage of rifting developed more rapidly and resulted in further deposition of the Namaqua volcano-sedimentary sequences as well as the Wilgenhoutsdrif sequences. During this stage the Kheis and Namaqua basins merged and formed a basin with geosynclinal characteristics. The third stage was characterised by the temporary cessation of mantle upwelling, resulting in compressive folding and subsequent anatexis which led to the generation and emplacement of the Keimoes Suite granitoids. Reactivation of mantle upwelling and rifting led to the grabens into which the Koras volcano-sedimentary sequence was deposited during the fourth stage.

Van Zyl (1981) envisaged the formation of a triple junction rift system as a result of

mantle plumes. The north-trending arm failed while the remaining two continued spreading to form a roughly east-west elongated ocean into which shallow-water marine and continental sediments were deposited. This model implies broadly synchronous deposition of the Bushmanland and related rocks with the Matsap, Vaalkoppies and Groblershoop rocks. Reversal of spreading led to subduction of the oceanic lithosphere and the formation of large quantities of calc-alkaline mantle-derived volcanics and granitoids. Isostatic instability, after the period of major orogeny, resulted in the marginal parts being subjected to tensional stress, and the Koras basin being formed.

5.1.2 Plate tectonic models.

Botha & Grobler (1979) envisage the subduction of an oceanic plate from west to east, with the Brulpan-Wilgenhoutsdrif rocks accumulating in a miogeosynclinal back-arc basin and the Namaqua rocks forming in an eugeosyncline in the west. In this model, the Areachap Group is seen as a pre-tectonic volcanic arc, while the Koras Group is interpreted as a post-tectonic continental volcanic arc sequence. A large body of metagabbro with associated amphibolite south of Copperton, which may represent a metamorphosed ophiolite complex, is considered by Botha & Grobler (1979) to be evidence for this subduction. Further evidence includes the calc-alkaline nature of the Areachap and Koras Groups, as well as a narrow kyanite-bearing zone in the Dagbreek Formation along the edge of the Namaqua "domain", which may indicate high-pressure overprinting during subduction. These authors admitted that the main flaw in this model is the fact that the synclinal sediments and the main period of tectogenesis are geochronologically two separate events.

Moen (1988), in explaining the genesis of the Wilgenhoutsdrif Group, suggested a subducting slab from east to west. The Groblershoop Formation was then deposited in the fore-arc region and the Vaalkoppies sediments in the back-arc miogeosyncline. Moen (1988) postulated that the Wilgenhoutsdrif Group may be allochthonous, having been tectonically emplaced to its present position, explaining the basal unconformity as a décollement plane and that the Jannelsepan amphibolites intruded as Andean-

type volcanics associated with an active continental margin. Later isostatic imbalances along a linear zone of structural weaknesses in the newly formed continent were accompanied by large-scale normal faulting and the development of the Koras Group.

Stowe (1983) considered two distinct "shortening" events. The first period of convergence occurred subsequent to the deposition of the Olifantshoek Sequence along the Kaapvaal Craton margin, where eastward thrusting over the continental margin may have caused the development of the serpentinite and kyanite zones in a possible "cryptic" suture zone. The Wilgenhoutsdrif Group was either formed in a back-arc basin or early rift environment. Further to the southwest, the Jannelsepan Formation formed obliquely to the Wilgenhoutsdrif Group in an arc setting. The second period of convergence was directed in a northerly direction with the underriding of the southern "plate" over the northern "plate" and the development of the granitic plutons in a Cordilleran setting. Continued northward movement led to large-scale transcurrent shears oblique to the line of convergence to bring the Namaqua Province in tectonic juxtaposition with the Kheis Province which had been relatively unaffected by the subduction. The development of the transcurrent shearing also led to the formation of the Koras Group of lavas and sediments.

5.2 A TECTONIC MODEL.

The model which best fits the metallogensis of the region is the subduction model based on the models described by Van Bever Donker (1991); Geringer (1979); Geringer & Ludick (1990); Geringer et al. (1986) and Stowe (1986).

The pre-Kibaran orogeny (Section 2.3) is not that well understood due to the excessive overprinting of the Kibaran-aged Namaqua orogenic event. However, it is clear that a widespread basement, overlain by various volcano-sedimentary sequences must have been developed prior to the onset of the Kibaran-aged orogeny. The distinction between these and later volcano-sedimentary sequences of Kibaran-age remains enigmatic. Although highly speculative, it is possible that initial rifting in a triple junction setting, as proposed by Van Zyl (1981), took place to form an elongate east-

west trending intracontinental basin (Moore, 1980) within which the line of sedimentary exhalative deposits defined by the Aggeneys, Putsberg and Adjoining Geelvloer were deposited. These deposits may have formed in third-order basins within the main rift valley. Large (1981) suggests that SEDEX deposits are all located in third-order basins, close to the margins of first or second-order basins and are all in proximity to fault zones (presuming the deposits to be proximal) that are considered to have been active during deposition of the host sediments. The mineralising fluids were then controlled by normal faulting which defined the edges of these basins and precipitated the sulphides synchronously with the influx of sediments, which Moore (1980) defined as having been deposited in continental, lagoon-shallow marine and marine environments. The Rozynbosch deposit could also have been related to this period of mineralisation, but was later remobilised and tectonically displaced during thrusting of the Hartbees River Thrust event and/or the final stages of northward convergence related to the main Kibaran-aged Namaqua event, as discussed later.

Whatever the main underlying cause was for the Bushmanland "microcontinent" being a separate landmass from that of the Kaapvaal Craton, whether it be as a result of rifting or as a separate continent altogether, the basic concept of collision tectonics seems to correspond favourably to explain the geological and metallogenic features that are characteristic of the area under investigation. The Areachap Group has been the subject of much interest over the past twenty years and considerable geological and even metallogenic evidence points towards an arc-related subduction zone from west to east. The Namaqua orogeny is therefore regarded as a destructive plate tectonic event.

The main mineralising event in the area under investigation coincided with the initiation of Kibaran-aged deformation, which commenced with closure of the oceanic basin and subduction of the ocean plate under the Kaapvaal Craton (Fig. 5.1a). Subduction and related calc-alkaline volcanism created an arc environment which is largely represented by the Jannelsepan Formation. These circumstances led to the development of the Besshi-type massive sulphide deposits along the arc in the Areachap Terrane between 1600-1300 Ma ago. According to Slack & Shanks (1989),

Besshi-type deposits occur in several different tectonic environments including:

- * spreading ridges in both continental-margin and back-arc basin settings,
- * within intracontinental rifts, and
- * along rifted continental margins.

These authors noted that there appears to be a continuum in geological settings for Besshi-type deposits from largely clastic metasedimentary sequences to metasedimentary sequences with minor associated felsic metavolcanic rocks to sequences with abundant metavolcanic rocks. The VMS deposits of the Areachap Terrane therefore probably precipitated during the more advanced stages of subduction when back-arc rifting was prevalent. In later stages of subduction, when the continental Kaapvaal Craton was further overriding the oceanic plate, increased sialic crustal material was produced, either by anatexis of crustal blocks underthrust at the trench, or by partial melting of the crust at the base of the thickening continental plate. At this stage, just before arc-continent collision, the tendency of increased felsic volcanism in a more mature arc setting, may have produced the deposits with slightly more Kuroko-type characteristics (Hutchinson, 1973), such as the Prieska ore body. The Kakamas Terrane, at the stage of island-arc development, is represented by the

Fig. 5.1a

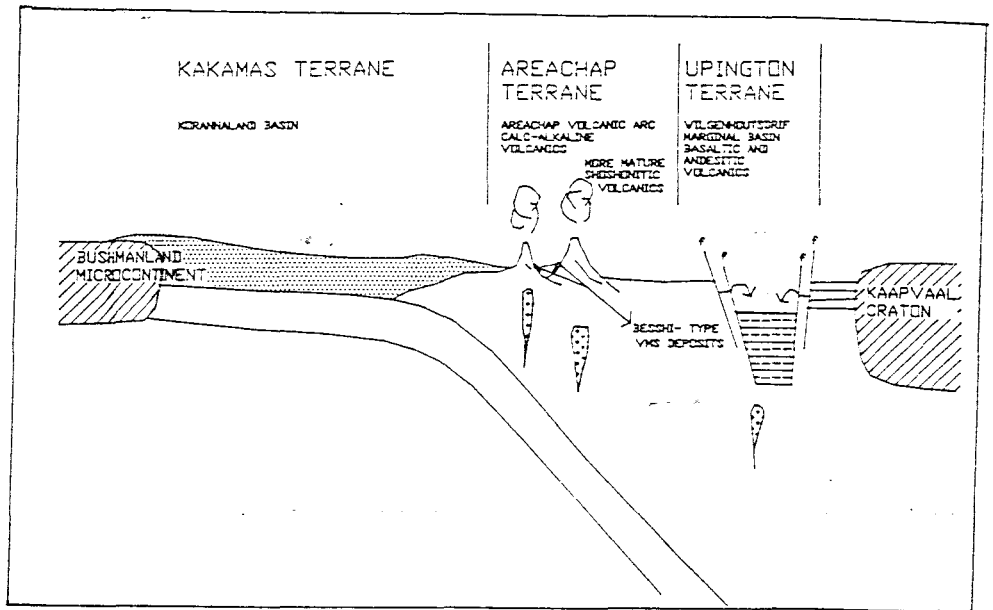


Fig. 5.1b

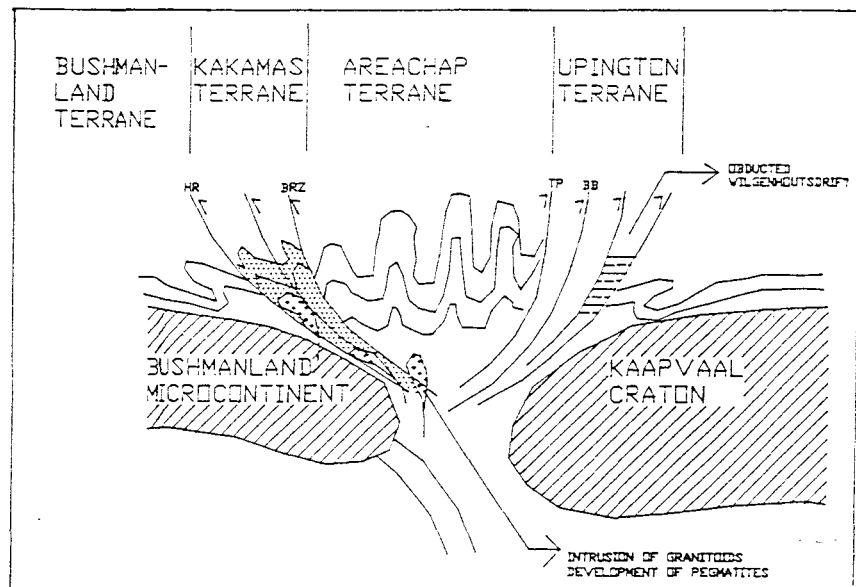


Fig. 5.1c

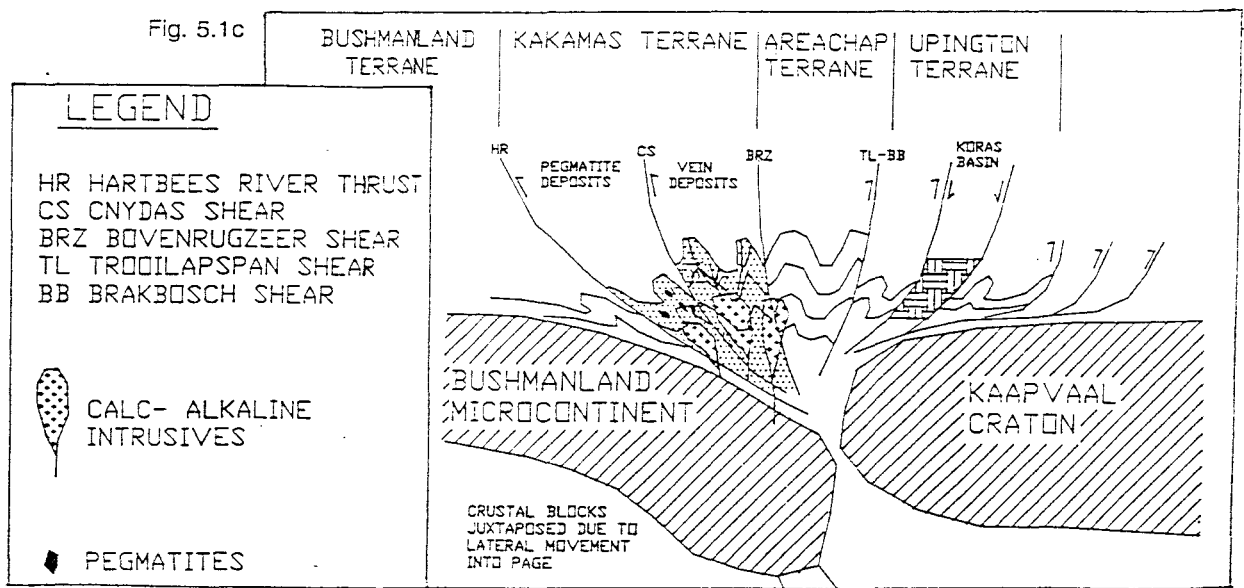


Figure 5.1: Schematic representation of the geotectonic evolution of the area under investigation.

Korannaland Sequence of metasedimentary rocks, and was probably deposited in a western basin-type environment (Fig. 5.1a).

During the more mature stages of the subduction, sinking of the underriding plate caused tension and rifting in the continental side of the arc (Fig. 5.1a). This rifting provided deep-seated faults, along which basaltic and andesitic lavas were extruded, and with the accompanied sedimentation within the marginal basin, formed the Wilgenhoutsdrif Group of rocks.

If the interpretation of the Wilgenhoutsdrif Group being a manifestation of a marginal basin is correct, then it is likely that the descending plate had a steeply dipping Benioff zone (Mitchell, 1973). This could be one of the reasons why there are no Sn, W, Bi, F deposits related to granites on the eastern side of the suture zone, as these deposits do not occur near continental margins along steeply dipping Benioff zones (Mitchell, 1973). Steeply dipping Benioff zone produce porphyry copper deposits (Mitchell, 1973) which are not evident in the region of investigation, but may already have been eroded away as these deposits form at very shallow depths. It is possible that ascending magmas from the melting subducted slab may have provided heat and fluids necessary to produce carbonate hosted Pb-Zn deposits (Mississippi Valley-Type) on the continental side of the subduction zone which could be found in the Transvaal dolomites, to the east of the Namaqua suture zone.

Further closure of the ocean led to the collision of the Bushmanland microcontinent against the Kaapvaal Craton at about 1300 Ma and the commencement of the intrusion of predominantly I-type granitoids and extensive thrust faulting (Fig. 5.1b). It is possible that at this stage the Wilgenhoutsdrif Group was obducted and that no complete ophiolite sequence is present (Stowe, 1986).

The calc-alkaline character of the Keimoes Suite granitoids (Geringer et al., 1988) suggests that they intruded into a destructive plate margin environment, although calc-alkaline magmatism does not indicate whether a volcanic arc, active continental margin or continent-continent collision environment was active at the time (Hall,

1987). The chemical characteristics of major, trace and rare earth elements (Geringer et al., 1988) point towards a destructive-margin granite with a tendency towards within-plate granites which intruded into a crust-dominated environment. In trying to accommodate the calc-alkaline volcanic arc of the Areachap Terrane, with which parts of the Keimoes Suite are intimately associated, Geringer et al. (1988) postulated that the granites intruded after subduction of a small oceanic plate which resulted in the formation of the Areachap volcanic arc. Further compression led to the underthrusting of the Namaqua plate beneath the Kaapvaal Craton, accompanied by considerable crustal thickening and the generation and emplacement of the Keimoes Suite. Because of its relatively lower density, the continental crust is unlikely to have been subducted to great depths. In general, syn to late-collision thrusting is largely concentrated in the underthrusting foreland, and is less intense in the adjacent former arc system and hinterland of the overriding continent (Mitchell & Garson, 1981). This leads to successive thrusting further towards the foreland interior, defined in the area under investigation by the Kakamas Terrane. This hypothesis also explains the appearance of progressively shallower granites from west to east described in Section 2.3, where crustal blocks are thrust up to shallower levels closer to the suture zones.

Continued vergence of the two continents from 1200-950 Ma resulted in increased thrusting, granitoid intrusion and anatexis, which is exemplified in the peak grade of metamorphism and related deposits (such as wollastonite and sillimanite deposits). There is a close spatial relationship between vein-type mineralisation, pegmatite development and the Keimoes Suite granitoids, all of which is located in the Kakamas Terrane. This situation is probably very similar to conditions proposed by Mitchell (1974) for the formation of Sn mineralisation in southwest England. There could therefore be a link between granitoid generation and the situation which led to the development of homogeneous (anatexis) and inhomogeneous (granite-related) pegmatites from deeper (mesozonal to katazonal) seated granites away from the suture zone, and vein deposits, related to shallower (mesozonal to epizonal) granites which were thrust higher, closer to the suture zone (Fig. 5.1c). This hypothesis could explain the apparent lack of direct spatial relations between the Keimoes Granite and the pegmatites, as the granitoids associated with the pegmatites are still unexposed at

depth.

Although the majority of Keimoes Suite granitoids are of the I-type, which is traditionally not associated with Sn-W vein-type mineralisation, Coetzee & Twist (1989) have shown that highly fractionated I-type granites may concentrate Sn in sufficient quantities to produce economically viable deposits. It is therefore likely that the emplacement of the Keimoes Suite is responsible for the generation of the Sn-W mineralisation, with or without a degree of crustal contamination. Base metal sulphide veins in the proximity of the Sn-W veins may have been derived from increased crustal contamination or leached by connate or meteoric waters from the surrounding metasedimentary sequences in a convection cell-type system.

The pegmatites contain both NYF and LCT components as defined by Cerny (1991) (see Section 4.8.2.1) and are generally concentrated in groups which define possible granitic plutons at depth. They therefore, according to Cerny (1991) were derived from depleted lower-crustal levels and were contaminated by undepleted lithologies or were formed from anatexis of a mixed range of depleted and undepleted protoliths. The various pegmatite groups (Section 4.8.3) may reflect increasingly shallower levels of intrusion from south to north in the Kakamas Terrane or deeper crustal levels from north to south.

Major northward-directed compressive forces led to a final period of transcurrent movement and the repositioning of crustal blocks, largely along the suture zone, into their present juxtaposed positions. This period of lateral movement simultaneously led to the formation of "pull-apart" fault-bounded basins and the deposition of the Koras rocks with accompanied mineralising fluids leaching Cu from basaltic host rock. The continued northward movement has been explained by Stowe (1986) to be due to the Southern Cape Conductive Belt which has been postulated by De Beer & Meyer (1983) to represent a northward-directed Cordilleran-type subduction zone.

Chapter Six

6. SUMMARY AND CONCLUSIONS.

6.1 SUMMARY OF EVENTS AND METALLOGENESIS.

A brief summary of events is provided in tabular form in Table 6.1.

6.2 METALLOGENIC DOMAINS.

The various styles of mineralisation are generally, although sometimes not exclusively, confined to the different tectonostratigraphic terranes. The tectonostratigraphic terranes can be subdivided into metallogenic domains which are divided by major shear zones or contrasting proto-lithologies and which contain unique styles of mineralisation (Fig. 6.1). These domains may prove useful in exploration. Detailed descriptions and exploration potential of the various styles of mineralisation have been discussed in previous sections and the following is only a brief outline of the proposed metallogenic domains. The metallogenic domains correspond well with the geological tectonostratigraphic subdivisions which points towards an increasingly correct interpretation of the mysteries created by the Namaqualand Metamorphic Complex.

Table 6.1. Summary of events in the area of investigation.

PERIOD	AGE	STRATIGRAPHIC UNITS	GEOLOGICAL EVENTS	METALLOGENIC EVENTS	
		BUSHMANLAND	GORDONIA KAKAMAS AREACHAP UPINGTON	KHEIS	
QUATERNARY		Kalahari and other		Aeolian and fluvial sands	Formation of evaporitic gypsum deposits Formation of surficial uranium deposits Formation of hydrothermal islandspar deposits
TERTIARY	65Ma			Soil and calcrete development	
JURASSIC	150Ma	Karoo		Dolerite intrusions	
CARBON-IFEROUS - PERMIAN	350Ma	Karoo		Glaciation and shallow marine and/or lacustrine sedimentation.	
NAMIBIAN	500Ma	Nama		Shallow marine and continental fluvial sedimentation. Stabilisation of crust.	

MOKOLIAN	900Ma						<p>Last movement along major transcurrent shear zones (eg. Brakbosch).</p> <p>Rifting related to transcurrent shearing.</p> <p>Influx of mafic and felsic lavas as well as sediments within grabens.</p> <p>Accretion of Provinces comprising of Kheis, Gordonia and Bushmanland.</p> <p>Period of major intrusion of calc-alkaline, mainly I-type, some S-type granitoids.</p>	<p>Final stages of pegmatites development</p>
	1150Ma				Koras		<p>Formation of Cu-rich hydrothermal veins along faults and fault hinges.</p> <p>Deposition of syngenetic Cu in basaltic lavas.</p>	
	1200Ma		Keirnoes	Keirnoes			<p>Development and emplacement of Be-, Fe-, mica-, REE-, Ta/Nb-, Li-, P- rich pegmatites. Major period of granitoid related Sn, W, F, Cu, Pb, Zn, Ag, Au hydrothermal vein deposits. (Possible remobilisation of tourmalinite horizons?)</p> <p>Development of deposits related to metamorphism such as wollastonite and sillimanite. (Possible development of asbestos in serpentinites at this time?)</p>	
	1350Ma				Wilgenhoutsdrif		<p>Formation of marginal basin as a result of rifting related to subduction. Influx of mafic and felsic lavas plus sediments.</p> <p>Intrusion of ultramafic bodies (serpentinites and Jacomynspan).</p>	<p>Syngenetic Cu deposits related to basaltic lavas.</p> <p>Intrusion of the Jacomynspan Ni-Cu-bearing ultra-mafic body.</p>
	1600-1300Ma			Areachap			<p>Development of fore-arc and back-arc related to subduction. Calc-alkaline volcanism.</p>	<p>Formation of Besshi-type volcanogenic massive sulphide deposits.</p>

MOKOLIAN OR OLDER	> 1600 Ma	Brakwater Arribees Droëboom De Kruis Grapples	Korannaland Hartbees River		Vaalkoppies	Bruipan	Sedimentary basins and volcanism on Bushmanland microcontinent. Continental margin sedimentation along Kaarvaal Craton.	Formation of sedimentary exhalative Pb- Zn deposits? Formation of tourmalinite horizons?
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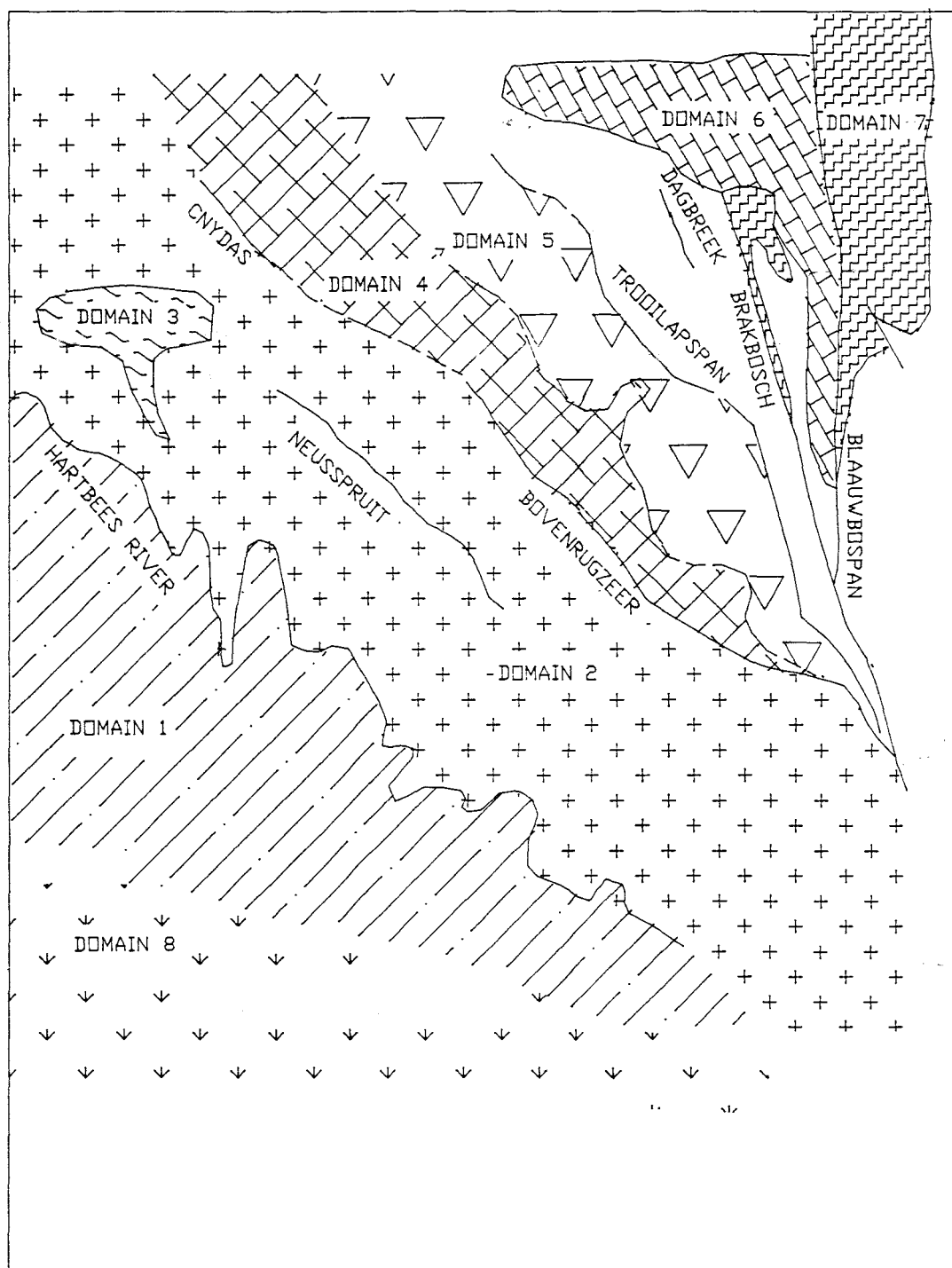


Figure 6.1: The various metallogenic domains within the area under investigation.

- Domain 1 - Bushmanland SEDEX deposits.
- Domain 2 - Kakamas pegmatites.
- Domain 3 - Riemvasmaak W deposits.
- Domain 4 - Kakamas vein deposits.
- Domain 5 - Areachap VMS deposits.
- Domain 6 - Koras Cu deposits.
- Domain 7 - Wilgenhoutsdrif Cu deposits.
- Domain 8 - Karoo islandspar deposits.

6.2.1 The Bushmanland Subprovince.

6.2.1.1 **Domain 1.**

The Bushmanland Subprovince in the area under investigation comprises Domain 1, which is characterised by base metal deposits, believed to be of sedimentary exhalative origin, and are usually associated with shear zones. The presence of the Rozynenbosch Pb-Ag (Zn-Cu) deposit, which is a possible extension of the line of base metal deposits incorporating the Aggeneys, Putsberg and Adjoining Geelvloer deposits, in a totally unrelated metallogenic domain, should continuously remind one of the structural complexity of the area, and that tectonic blocks from a different metallogenic domain may be juxtaposed next to one another.

6.2.2 The Kakamas Terrane.

6.2.2.1 **Domain 2.**

This Domain is bounded by the Hartbees River Thrust in the southwest and the Cnydas-Bovenrugzeer Shear in the northeast. Domain 2 is characterised by pegmatite development and deeper level granitoids and could, as a whole, be seen as a deeper level crustal block relative to Domain 3, to the northeast. On a broad scale, therefore, pegmatites are not necessarily confined to specific lithological units but are rather governed by tectonically controlled crustal segments. The metasedimentary Korannaland Sequence could be seen as a separate metallogenic domain as this sequence contains all the concentrations of metamorphic minerals such as wollastonite and sillimanite. This division is not, however, entirely justified as numerous pegmatites are intruded into these metasedimentary sequences towards the south in the Kenhardt area.

6.2.2.2 Domain 3.

Domain 3 lies within domain 2, its boundary is arbitrarily demarcated by the geographical extent of the W deposits defined by the Riemvasmaak W Province. These deposits are largely characterised by the development of restite fluids during the process of ultrametamorphism.

6.2.2.3 Domain 4.

This Domain is a triangular block bounded by the Cnydas-Bovenrugzeer Shear zone in the southwest and the Areachap Group in the east, and is characterised by the presence of vein type deposits, probably related to shallower level granitic intrusives. This Domain possibly represents a crustal wedge, pinched out towards the north during the final stages of northward movement related to the formation of the Koras basin. Geringer & Botha (1977) pointed out the sinistral movement along the Cnydas Shear as well as the differing structural, and to a lesser degree metamorphic signature of the lithologies on either side of the Cnydas Shear. This would further justify the subdivision of the Kakamas Terrane into Domain 2 and 4.

6.2.3 The Areachap Terrane.

6.2.3.1 Domain 5.

Domain 5 is represented by the Areachap Terrane of arc related calc-alkaline volcanism and sedimentation with accompanied volcanogenic massive sulphide deposits.

6.2.4 The Upington Terrane.

6.2.4.1 Domain 6.

The Koras Group of rocks represents Domain 6, which is characteristic of Cu

mineralisation, mainly related to structural features such as faults and fold hinges. Although no sediment hosted Cu mineralisation is known in this Domain, there is a possibility of this type of deposit being present.

6.2.4.2 Domain 7.

Mineralisation in the Wilgenhoutsdrif Group, which makes up Domain 7, is, as far as is known, restricted to syngenetic disseminated Cu mineralisation within basaltic metalavas and epigenetic mineralised fault breccia, the latter is probably related to Koras-aged deformation.

6.2.5 Post Proterozoic deposits.

6.2.5.1. Domain 8.

This Domain is present in the southern portion on the map and is restricted to the contacts between the Prince Albert Formation of the Ecca Group and the intrusive late-Karoo-aged dolerite sills and dykes.

The delineation of two additional domains representing the uranium and gypsum deposits could tentatively be drawn over the whole region, although this would not amount to a credible portrayal of the potential for these commodities. More detailed studies on the possible source rocks for these commodities would have to be made before potential areas could be delineated. An example of this may be a study on the granites in the area which could provide a potential source for the uranium. At present all recent and palaeo-streams and pans are potential hosts.

6.3 THE EFFECTIVENESS OF GIS.

The metallogenic analysis of an area can be extensively aided by the use of a Geographical Information Systems package. The only limitation is that of the data base itself, to which the information is distributed graphically, and any further addition

to the existing data base could only improve the extent to which the GIS can be applied. The limiting factors to which this project was exposed to are the following:

* The two 1:250 000 scale geological maps that were used proved to contain too little information. An example of this is the Kenhardt Formation which consists of a multitude of different rock types including amphibolites, biotite gneiss, quartz-feldspar-biotites gneiss, calc-silicates, marble and aluminous gneiss. It is therefore not possible to ascertain whether certain pegmatite commodities have a preference for specific host rocks. However on a broader scale, it is possible to delineate metallogenic domains into which certain styles of mineralisation are confined. In the Areachap Terrane the distribution of the Jannelsepan Formation is too broad a base for a detailed metallogenic study of the VMS deposits associated with it. More detail is necessary to show the distribution of the specific rock types within the Jannelsepan Formation so that a comparison can be made between mineralised host rocks and other rock types. This will enable the prediction of additional areas of similar characteristics.

* The mineral deposit data base occasionally did not provide enough detailed information on specific deposits. For example, a more detailed study on the ore mineralogy of the various fault hosted Lutzputs deposits could more accurately depict mineral zoning of this region, which could lead to an interpretation of the direction from which the source fluids originated and if, indeed, the fluids originated from a single granitic pluton or more than one, or if the fluids originated from a single mineralising pulse or from numerous pulses. A detailed investigation, petrological or otherwise, into all the different deposits is beyond the scope of this project, but the addition of such data could only benefit any future metallogenic analysis of the area.

* The extrapolation of aeromagnetic, gravity, radiometric data as well as the geochemical data from the Geological Survey geochemical project and Landsat imagery onto the current data would also enhance the interpretability of the area under investigation and most certainly lead to further targets for exploration within the confines of the metallogenic domains defined.

* A more detailed knowledge of the ore reserves and grades of the individual deposits would also enhance the possibility of assessing the metallogeny of the area and aid in the determination of predictive metallogeny.

* The delineation of areas characterised by various surface processes could also enhance the capability of the GIS. For example, the defining of calcretes according to authigenic or allogenic superimposed on airborne radiometric data would greatly enhance the capability of defining potential areas of calcrete-hosted uranium deposits.

It is therefore concluded that at a scale of 250 000, there are too many limitations to effectively analyse the metallogeny of an area to the extent that further targets for exploration can be defined. However, a metallogenic analysis at this scale can certainly be of assistance in defining new areas of potential interest where further detailed work should be carried out, as well as clarifying what type of information is still needed.

6.4 AFTERTHOUGHT.

The metallogeny of the area under investigation complements the more recent ideas held by researchers. The metallogenic domains correlate well with the various tectonostratigraphic terranes. The styles of mineralisation certainly show some compatibility with the interpreted geology.

More detailed work on individual syn to late-tectonic granites in the area would further assist in unravelling the mysteries of plate tectonics in the area, as well as leading to a better understanding of the granite related ore deposits. This, and the extrapolation of geophysical and geochemical data would almost certainly lead to the discovery of new deposits, although it would seem that most, if not all, the surface deposits have been discovered.

The Gordonia Subprovince, along with the Natal Subprovince, provides only a relatively small "window" through which to observe the Namaqua-Natal belt where it collides with the Kaapvaal Craton. The extrapolation of knowledge gained in the

Namaqua foreland could only improve the chances of finding new ore bodies further north and south under the Nama and Karoo cover respectively, as technology improves. An example of this is the Areachap Terrane, where there is a propensity towards the Kuroko type deposit northwards towards the Areachap and southwards towards the Copperton deposits. The "big one" of the Kuroko type VMS deposit may yet await exploitation under the Nama or Karoo cover.

The geological history before the Namaqua event is obviously more enigmatic due to structural and metamorphic overprinting. The metallogeny provides information in assisting the unravelling of this enigma but further research will certainly lead to increase understanding of the geology in this regard. Most of this, however, lies to the west of the area under investigation and only speculation into the geological history of the Eburnian-aged rocks is provided.

HERE ENDETH THE LESSON

7. ACKNOWLEDGMENTS.

I would like to thank the Council For Geoscience (Geological Survey of South Africa) and Dr. E.C.I. Hammerbeck in particular for giving me the opportunity (financial and time wise) to undertake this MSc. Members of the Mineral Resources Division are also thanked for their friendship and support. Dr. M.C. Du Toit was always available for discussion.

Prof. J.M. Moore is thanked for his excellent presentation of the MSc. course in general, and for his constructive guidance in the preparation of this thesis. Furthermore, Dr. R.W. Harris is sincerely thanked for his time, effort and friendship (way beyond the call of duty) in helping me set up the MIPS GIS. All other members of staff of the Geology Department at Rhodes University, particularly Ms. Sue Brooks, the Exploration Geology secretary, are also thanked for their friendliness and support during my stay in Grahamstown.

The MSc. students; particularly Stefan Mujdrica, Karl Hartmann, Sue Frost, Harilaos Tsikos, Peter Mann, Ian Gendall, Phillo Schoeman and Chimwemwe Chikusa are thanked for their companionship and assistance.

The Zandberg family in Kenhardt; Jan, Jeanette, Carla, Tanya, Janie and Lelanie are also thanked for providing me with a "home" during my field work.

My parents, Theo and Rie and brother Eric, are thanked for providing me with a stable, loving and supportive home (Boelema, pers. comm.) as well as an education, without which I would not have been able to undertake this project.

My girlfriend, Yvonne Hogg, is also thanked for her loving friendship, patience, help and support, as well as understanding, particularly during the long periods while I was away.

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9. APPENDIX A.

Table A-1. Summary of the Kheis Tectonic Province within the quadrangle under investigation.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Brulpan Group	Groblershoop Formation	min. 1750 Ma	Quartz- muscovite schist, quartzite, quartz- amphibole schist, lenses of greenstone.	
	Uitdraai Formation	max. 2252 Ma	Blue- grey massive to layered quartzite, cross bedded in places. Hematite nodules in places.	Hematite nodules possibly of biochemical origin .
	Prynnsberg Formation	max. 2252 Ma	Muscovite quartzite and schist.	
	Kaboom Formation		Quartzite with minor schist	Possible correlation with Vaalkoppies Group (ie. Sultanacoord or Dagbreek Formations). Underlies Brulpan Group (Uitdraai Formation).
Intrusive Rocks				
	Kalkwerf Gneiss		Red- brown coarse- grained granite gneiss.	

1. SACS (1980).
2. Moen (1980).

Table. A-2. Summary of the Upington Terrane.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Koras Group	Kalkpunt Formation(N,C)	extrusion of lavas	Red-brown sandstone, conglomerate in upper portion.	The Koras Group is divided by Moen (1978) into the northern (N),
	Adelsted Formation(C)	at about 1200Ma	Amygdaloidal basalt, epidotised tuff, minor conglomerate and sandstone.	central (C) and southern (S) domains. The succession varies from domain to domain and is
	Leeuwdraai Formation(C)/ Welgevind Formation(N)		Red-brown quartz-feldspar porphyry/ Red-brown quartz-feldspar porphyry with interbedded basaltic lava and breccia.	indicated as such in the Table.
	Rouxville Formation(N,C,S)		Andesitic to basaltic lava, commonly amygdaloidal, pyroclastics. Abundant epidotised breccia and thin layers of tuffaceous sediments.	Basic lavas follow conformably on conglomerates of the Rusplaas Formation.
	Ezelfontein Formation(S)/ Rusplaas Formation(N)		Conglomerate, sandstone/ Conglomerate sandstone.	Rusplaas Formation display sharp contact with Boom River lavas (N).
	Swartkopsleegte Formation(C,S)/ Steenkampsputs Formation(N)		Red-brown quartz-feldspar porphyry/ Dark grey quartz-feldspar porphyry.	
	Bossienek Formation(C)		Conglomerate, sandstone, mudstone.	
	Boom River Formation (N,C,S)		Layered sequence of andesitic to basaltic lava with minor pyroclastics.	Conformably overly the Christiana Formation
	Christiana Formation(N)		Red-brown, micaceous and feldspathic sandstone, lenses of conglomerate at top, quartzite.	Koras Group lies unconformably over Wilgenhoutsdrif and Vaalkoppies Groups (Dagbreek Formation). Absence of conglomerates near base indicate slow initial subsidence of basin.
Intrusive Rocks				
	Blauwbosch Granite		Moderate to highly porphyritic granite.	Small (<5km), no metamorphic overprint. Probably one of the youngest granites in the Korannaland. Highly differentiated possible final stage of Koras Igneous event.
	Rooiputs Granophyre		Medium grained granophyre.	Possibly more than one age of emplacement. Possibly coeval
	Felsic Dykes	1200 Ma	Quartz-feldspar porphyry.	Cross cut Dagbreek Formation and Koras Group up to and including the Rouxville Formation and Betadam Gabbroite. No dykes intruded into the Kalkpunt and Welgevind Formations indicating the dyke pre-date these formations.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
	Betadam Gabbroonorite		Gabbroonorite.	Occur predominantly in the Christiana Formation as well as basement rocks.
Wilgenhoutsdrif Group	Leerkrans Formation	1336 Ma	Metabasite, felsic lavas, greenschist, conglomerate, ferruginous chert.	Cyclic repetition of felsic and basic lavas with mainly clastic sediments.
	Zonderhuis Formation		Phyllitic schists, phyllites, quartzite, ferruginous rocks (eg.chert), carbonates and ultramafics.	Ferruginous chert developed randomly throughout Wilgenhoutsdrif Group. Serpentinite lenses commonly associated with carbonate rocks.
	Grootdrink member		Light purple quartzite with well developed flaser-like lamination. Manganiferous concretions.	Overlies Groblershoop Formation with a gradational contact.
Vaalkoppies Group	Sultanaoord Formation		White massive quartzite with interbedded phyllite.	
	Dagbreek Formation	2400 - 2100 Ma	Quartzite and schist grading into banded gneiss and migmatite, lenses of leucogneiss, amphibolite and serpentinite.	
Intrusive Rocks				
	Kleinbegin Subsuite		Medium to coarse grained weakly foliated granite.	Part of the Kelmoes Suite. Intrudes both the Dagbreek Formation and the Areachap Terrane.
	Swanartz Gneiss		Porphyritic biotite granite.	Intrudes the Dagbreek Formation in the north.

Koras section compiled mainly from Moen (1987).

Wilgenhoutsdrif section compiled from Moen (1988).

1. Kröner (1977).
2. SACS (1980).
3. Botha et al. (1979).

Table A-3. Summary of the Areachap Terrane within the quadrangle under investigation.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Areachap Group	Jannelsepan Formation	1600-1300 Ma	Fine to medium grained amphibolite with lesser interbedded calc-silicate, quartz-feldspar gneiss, mica schist and metapsammite displaying gradational contacts. Massive amphibolite as well as bands defined by mineralogical and textural variations.	Amygdales as well as possible flow structures point towards a volcanic origin for some of the amphibolites. Forms the thickest sequence in the Areachap Group.
	Bethesda Formation		Metapelitic biotite rich gneiss and schist with varying amounts of biotite, sillimanite, cordierite, garnet, quartz and feldspar and minor interbedded amphibolite and calc-silicates.	Generally restricted to the western part of the Areachap Terrane. Relationship with Van Wyks Pan Formation is uncertain but at correlation is possible.
	Sprigg Formation		Highly variable, mica schist, quartz-feldspar ± muscovite schist, quartzite, mega-conglomerate with minor amphibolite and amphibole gneiss.	Predominantly along the eastern boundary of the Areachap terrane. Most of the pebbles in the conglomerate are thought to originate from the Vaalkoppies group and none from the Jannelsepan Formation indicating the probable base of the Areachap Group.
	Ratel Draai Formation		Kinzigitic	Minor component along the western boundary of the Areachap Terrane.
	Van Wyks Pan Formation		Medium grained granoblastic quartz-feldspar rock with lesser layers of amphibolite and minor calc-silicates. Occasional thin layers of magnetitic chert and quartzite also present.	Correlation with the Areachap Group is still speculative.
Intrusive Rocks				
	Keimoes Suite			Mainly syn- to late-tectonic granites. Granites east of the Bovenrugzeer Shear exhibit higher magnetic susceptibility than those to the west in the southern portion where the shear cuts the Areachap Group.
	No named mafic and ultramafic rocks.		Weakly foliated gabbro-norite. Coarse grained largely unfoliated dunite with talc, serpentine and magnetite.	Probably syn- to late-tectonic, but affected by shearing.

Predominantly derived from Slabbert et al. (in press).

1. Barton & Burger (1983).

2. Quoted by Slabbert et al. (in press).

3. Humphreys (1985).

Table A-4. Summary of the Kakamas Terrane.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
	Sout River Formation		Fine- to medium grained banded biotite gneiss, muscovite gneiss and sillimanite gneisses.	
	Goedehoop Formation		Quartzite, sericitic and or feldspathic in places and well as lenses of conglomerate.	
Korannaland Sequence	Valsvlei Formation	1800-2300 Ma for the Korannaland Sequence.	Feldspathic quartzite with minor interlayered calc-silicates.	Valsvlei and Ganzenmond Formations represent lateral facies variations of one another.
	Ganzenmond Formation		Irregularly layered quartz-feldspar-biotite ± sillimanite ± garnet gneiss, quartz-feldspar-biotite gneiss, massive quartzite and quartz-feldspar gneiss.	
	Puntsit Formation		Fine- to medium grained layered diopside-epidote-hornblende-quartz-feldspar rock, massive garnet rich calc-silicate, marble and amphibole-quartz-feldspar gneiss. Also contains appreciable amounts of wollastonite.	
	Toeslaan Formation		Kinzigite, pelitic gneiss, biotite gneiss and leucocratic paragneiss.	
	Sandputs Formation		Thinly layered fine grained quartzite with varying amounts of feldspar and calc-silicate minerals.	Hosts the scarce dumortierite quartzite, recently declared a national monument.
	Omdraai Formation		Leucocratic quartz-feldspar gneiss, amphibolite and quartzite.	
	Piet Rooisberg Formation		Medium grained weakly foliated quartz-feldspar gneiss.	Equivalent to the Rautenbach se Kop Formation . probably of sedimentary origin .
	Renosterkop Gneiss		Quartz-topaz gneiss	
	Koekoepkop Formation		Quartz-feldspar-amphibole gneiss	
	Venterskop Formation		Kinzigite	
	Sandnoute Formation		Aluminous gneiss interbedded with quartzite, fine grained amphibole rich feldspathic quartzite, amphibole gneiss, amphibolite and calc-silicates	Possible correlates to the Jacomynspan Formation or the Venterskop Formation.
Jacomynspan Group			Aluminous metapelitic gneiss with interbedded quartzite, leucocratic quartz-feldspar gneiss, amphibolite, calc-silicate and marble.	Displays a low magnetic susceptibility. Similar sequences further southeast are described as the Vogelstruisbuit Formation and the Copperton Formation . It has also been correlated with the Sandnoute Formation .
Koelmanskop	Collinskop Formation		Kinzigite.	
Metamorphic	Bok-se-puts Formation		Quartz-feldspar gneiss with quartz rich and pelitic rich zones.	

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Sequence	Koukop Migmatite		Migmatitic leucogneiss and biotite gneiss which is garnetiferous in places as well as amphibole gneiss.	
	Witwater Gneiss		Garnetiferous mica-poor gneiss pegmatitic in places.	
	Twakputs Gneiss		Megablastic garnetiferous biotite gneiss.	
Hartbees River	Wolfkop Formation		Dominantly marble (dolomitic) with interbanded quartz-feldspar gneiss, amphibolite and calc-silicates.	Younging of the Vyfbeker Metamorphic Suite is thought to be towards the northeast.
Complex (Vyfbeker Metamorphic Suite)	Hugosput Formation		Quartz-feldspar gneiss with layers of garnet-sillimanite aluminous gneiss, biotite gneiss, amphibolite, quartzite and calc-silicates.	
	Rozynebosch Formation		A dominantly feldspathic unit and a dominantly calc-silicate unit: The feldspathic unit constitutes a quartz-feldspar ± biotite gneiss with interbanded migmatitic biotite gneiss and calc-silicates near the upper contact. The calc-silicate unit comprises of amphibolite, granoblastic calc-silicate with minor quartz-feldspar gneiss and biotite gneiss.	Possibly related to the amphibolites of the Kenhardt migmatites but separated due to the Rozynebosch Formation being fault bounded.
	Dreyer's Put Formation		Medium- to coarse grained granoblastic quartz-feldspar gneiss with banded as well as massive quartzite.	
	Kenhardt Migmatite		Interbanded variety of biotite gneisses, quartz-feldspar ± biotite gneiss with lenses of amphibolite, calc-silicate, marble and aluminous gneiss all with a migmatitic character.	Differentiation between sedimentary or igneous origin for most of the units difficult.
	Piet Rooi's Puts Gneiss		Medium- to fine grained quartz-feldspar ± biotite gneiss with amphibolite lenses.	Possible correlate with the Kenhardt Migmatite.
	Mottels River Formation		Medium grained banded quartz-feldspar ± biotite ± sillimanite ± cordierite gneiss. Abundant pegmatites.	Banding defined by biotite, sillimanite or cordierite which possibly represent clay layers within the original sediment.
	Aasvogelkop Gneiss		Leucocratic thinly banded gneiss.	
	Putsies Gneiss		Fine- to medium grained quartz-feldspar ± biotite gneiss.	Becomes more biotite rich towards the northeast. The biotite rich and biotite poor gneisses display a lit-par-lit character and is probably related to the Hartbees River Shear where pre-tectonic units were emplaced as well as in situ granitisation occurred during the time of shearing.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
	Driehoek Formation		Dominantly amphibolite (hornblende) associated with interlayered calc-silicate, marble, quartz-feldspar \pm sillimanite gneiss and biotite gneiss or schist.	Usually found along the Hartbees River Shear. Amphibolite probably of sedimentary origin.
Intrusive Rocks				
	Keimoes Suite	1200-1054 Ma.	Calc-alkali with dominantly I-type affinity. Differentiation between the granites exist but it is not clear if all the granites of the Keimoes Suite belong to the differentiation series. See text for more detail.	Syn- to post-tectonic granites. Friersdale Charnockite is one of the last granites to crystallise.
	Unnamed mafic and ultramafic rocks		Serpentinites, dunites, gabbros	
	Eendoorn Suite		Biotite rich occasionally garnetiferous granite gneisses.	Pre- to syn-tectonic granites.
	Kalkwerf Gneiss		Coarse grained granite gneiss.	Pre- to syn-tectonic granite.
	Augrabies Gneiss		Granite Gneiss.	Pre- to syn-tectonic granite.
	Kakamas Suid Gneiss		Augen Gneiss.	Pre- to syn-tectonic granite.
	Riemvasmaak Gneiss		Granite gneiss with granular or augen texture.	Pre- to syn-tectonic granite.
	Dyasons Klip Gneiss		Porphyroblastic to megacrystic gneiss.	Pre- to syn-tectonic granite.
	Klip Bakken Gneiss		Coarse to megacrystic quartz-feldspar gneiss.	Pre- to syn-tectonic granite.
	Curries Camp Gneiss		Coarse to megacrystic quartz-feldspar gneiss.	Pre- to syn-tectonic granite.
	Banks Vlei Gneiss		Biotite rich and biotite poor granite gneiss.	Pre- to syn-tectonic granite.
	Lutzputs Gneiss		Granite gneiss with sillimanite and garnet.	Pre- to syn-tectonic granite.

Derived mainly from Slabbert et al. (in press) and Moen (1988). Due to the structural complexity of the area, the table is not necessarily arranged in chronological order.

1. Harris (1985).
2. Slabbert (1987).
3. Potgieter et al. (1980).
4. Cornell et al. (1986).
5. Moen (1988a).
6. Walraven et al. (1982).
7. Botha et al. (1976).
8. Geringer et al. (1988).

Table A-5. Summary of the Bushmanland Subprovince within the quadrangle under investigation.

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Brakwater	De Banken Gneiss		Mainly a medium grained well foliated quartz-feldspar-biotite gneiss with numerous variations thereof.	Possible igneous origin.
Metamorphic Suite	Haakdoorn Formation		Quartz-feldspar gneiss, amphibolite, calc-silicate, aluminous gneiss, biotite gneiss and minor quartzite.	Metasedimentary.
	Sandkoppies Formation		Fine- to medium grained layered calc-silicates with occasional lenses of quartz-feldspar gneiss. Disseminated magnetite common.	Metasedimentary.
	Regt Kyk Gneiss		Granitic gneiss with augen texture	
	Nieuwe Post Wes Gneiss		Garnetiferous biotite rich augen gneiss.	
	Poliesberg Formation		Fine- medium grain thinly banded quartzitic calc-silicates interlayered with quartz-feldspar gneiss and marble. Cross bedding at places.	
	Slypsteenkrans Formation		Quartz-feldspar- and quartz-feldspar-biotite gneiss with minor calc-silicates.	
	Moddergat Gneiss		Variety of biotite gneisses, mostly banded with calc-silicate lenses.	
De Kruis Group	Kokerberg Formation		Medium grained quartz-feldspar gneiss with abundant pegmatites.	Represents an impure quartzite- a paleo heavy mineral deposit .
	Zandberghoop Formation		Fine- to medium grained calc-silicates with lenses of quartz-feldspar-biotite gneiss and amphibolites occasionally migmatitic at places.	The two formations of the De Kruis Group are highly interbanded.
Grappies Group	Soutputs Formation		Quartzitic calc-silicate with lenses and layers of amphibolite, conglomerate and quartzite.	Limited to the deformation of the Geelmoer Shear.
	Kraandraal Formation		Migmatised, epidotised calc-silicate, quartzitic calc-silicate, fine grained calc-silicate and amphibole.	The amphibolite probably represents a metamorphosed tholeiite and is associated with sulphide mineralisation .

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
	Bossiekom Formation		Medium grained quartz-feldspar gneiss interbanded with amphibolite, calc-silicate and white quartzite.	
	Brulkolk Formation		Quartz-feldspar gneiss with pegmatite overlying calc-silicate with lenses of muscovite schist, marble, conglomerate and amphibolite. Contains bedding and cross bedding.	
	Kameelputs Formation		Quartz-feldspar gneiss, banded biotite gneiss, amphibolite, biotite-quartz-feldspar gneiss interbanded with conglomeritic calc-silicates, marble and quartzite.	
	Rietput Formation		Migmatized biotite gneiss, amphibolite, marble, calc-silicates and quartz-feldspar gneiss with varying amounts of hornblende.	
Droëboom Group			Biotite gneiss, medium- to fine grained quartz-feldspar ± muscovite gneiss, calc-silicate interbanded with quartzite, limestone, amphibolite and conglomerate. Migmatitic in the more pelitic rocks.	
Arribees Group			Calc-silicate, marble, quartz-feldspar gneiss and biotite gneiss.	

GROUP	FORMATION	AGE	LITHOLOGY	REMARKS
Intrusive Rocks				
	Basjan Granite		Medium- to coarse grained, moderate to well foliated granite.	Xenoliths of Dwyka tillite found in the granite at places. Other places indicate the infilling of Dwyka tillite into cracks of the granite. Possible paleosol?
	Bakoondsvlei Granite		Coarse grained, moderately foliated porphyritic biotite granite.	Feldspar phenocrysts indicate a possible relation to the De Bakken Granite.
	Lat River Granite		Medium grained, weakly foliated biotite granite.	Intrusive into the Wit Klip Granite.
	Uitkoms Granite		Medium grained, weakly foliated quartz-feldspar granite with varying amounts of biotite and garnet.	Intrusive into the De Bakken Granite
	Wit Klip Granite		Coarse grained porphyritic biotite granite.	Minor copper mineralisation within joints.
	De Bakken Granite	900 Ma	Coarse grained, weakly foliated porphyritic biotite granite with numerous pegmatites.	Intrusive into the De Kruis Group.
Lange Kolk Suite			Three main types: a) Medium grained porphyritic quartz-feldspar-biotite gneiss and a fine grained granite gneiss, possibly intrusive into the former. b) Coarse grained megaporphyritic granite gneiss occasionally intruded by a fine grained biotite granite and garnetiferous dykes. c) Quartz-feldspar gneiss.	Usually display a lit-par-lit contact relation. Includes xenoliths of calc-silicates with large magnetite grains.
	Vaalhoek Granite		Fine- to medium grained granite and coarse grained biotite granite with xenoliths of calc-silicates and amphibolite.	
	Sandbakken Metabasite		Contains a wide variety of rocks with varying grain size, colour, mineralogy and xenoliths.	Possibly a layered complex.
T'Oubep Suite		1200 Ma	Granodiorite, tonalite, granite with numerous xenoliths of amphibolite and calc-silicate.	Mainly intrusive into the Bruikolk Formation. Calc-alkali with an I-type affinity. This suite represents a differentiation series and does not show any genetic relation with the Keimoes Suite.

information mainly derived from Slabbert et al. (in Press). Due to the structural complexity of the area, the table is not necessarily arranged in chronological order.

* unless otherwise specified.

1. Frick & Wheelock (1983).
2. Burger, A.J., quoted in Linström (1977).
3. Agenbacht (1992).
4. Joubert (1986a).

Table A-6. Characteristics of the Keimoes Suite.

NAME	CONTACT RELATIONSHIP	DEPTH LEVEL	TYPE	AGE (Pb/Pb-zircon, unless otherwise stated)
Bakrivier	Sharp, lit-par-lit, concordant	Katazonal	I-Type	1080
Gifberg	Sharp, finer grained towards margins	Mesozonal	I-Type	
Stukkededam	Sharp, concordant	Katazonal	I-Type	
Gemsbokbult	Concordant, sheared foliated near contact	Mesozonal	I-Type	1200
Klip Koppies	Sharp, foliated near contact	Mesozonal	I-Type	
Louisvale	Sharp, foliated near contact	Mesozonal	I-Type	
Colston	Foliated near contact	Mesozonal	I- and S-Types	1156
Cnydas	Sharp, crosscutting	Meso- to Epizonal	I-Type	1155 (Granodiorite) 807 (Langeklip Granite)
Friersdale	Sharp, crosscutting, no meta. aureole	Meso- Epizonal	I-Type	1087 1085 (Rb/Sr)
Kanoneland	Sharp, foliation near contact, no meta. aureole	Meso- Epizonal	I-Type	
Vaalputs	Sharp, crosscutting	Mesozonal	I- and S-Types	
Straussberg	Sharp, meta. aureole	Meso- Epizonal	I-Type	1080 1264 +/- 604 (Rb/Sr)
Kleinbegin	Sharp, crosscutting	Epizonal	I-Type	

After Geringer et al. (1988).

1. Geringer et al. (1977).
2. Linström (1977).
3. Jankowitz (1986).
4. Barton & Burger (1983).

APPENDIX B

The graphs depict the concentration of pegmatites containing a specific commodity within various formations. The numbers on the x-axis represent formations as shown in the following two tables:

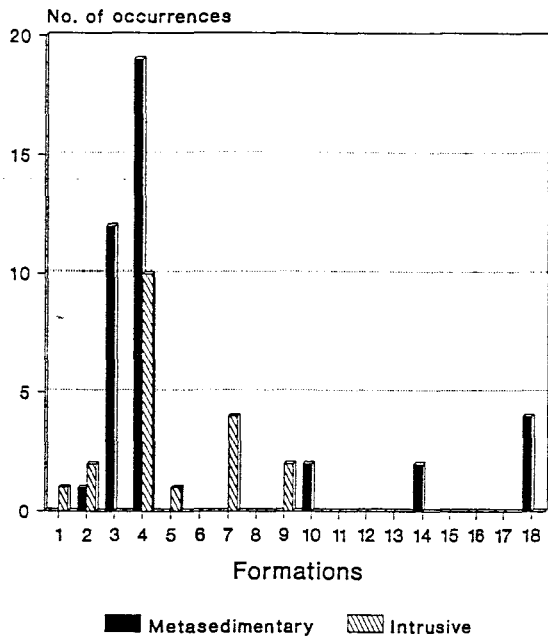
STRATIGRAPHIC POSITION	NUMBER	FORMATION
KAKAMAS TERRANE		
Koelmanskop Metamorphic Suite	1	Bok-se-puts
	2	Kourop
	3	Witwater Gneiss
	4	Twakputs
Korannaland Sequence	5	Valsvlei
	6	Garzenmond
	7	Puntsit
	8	Sandputs
	9	Omdraai
Vyfbeker Metamorphic Suite	10	Kenhardt Migmatite
	11	Mottels River
	12	Putsies
	13	Driehoek
Jacomynspan Group	14	
BUSHMANLAND SUBPROVINCE		
	15	Moddergat Gneiss
	16	Kokerberg
	17	Zandbergshoop
Arribees Group	18	

INTRUSIVE ROCKS		
AGE	NUMBER	NAME
Pre to syn-tectonic	1	Donkiesboud Granite Gneiss
	2	Dabene Granodiorite
	3	Augrabies Gneiss
	4	Riemvasmaak Gneiss
Syn to late-tectonic	5	Unnamed basic rocks
	6	Vaalputs Granite
	7	Rok Optei
	8	Klein Van Wyks Pan Granite
	9	Elsie se Gorra
Post-tectonic	10	Friersdale Charnockite

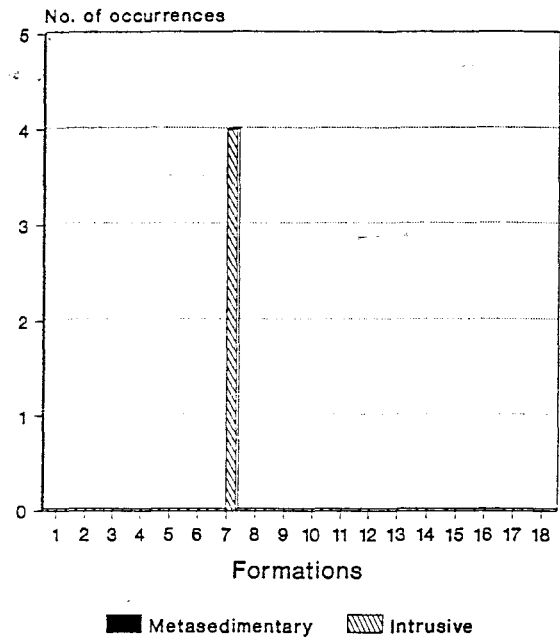
The commodity symbols used are explained in the following table:

SYMBOL	EXPLANATION
Be	Beryl
RE	Rare Earth-bearing minerals
Mica	Mica
Li	Li-bearing minerals
P	Apatite
Ta	Tantalum-Niobium
An	Andalusite
F	Fluorite
Rose Quartz	Rose Quartz
Corundum	Corundum
Magnetite	Magnetite

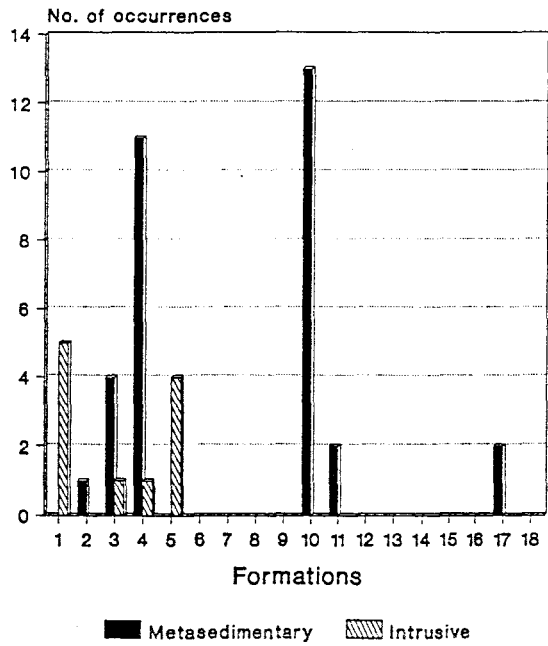
P CONCENTRATION



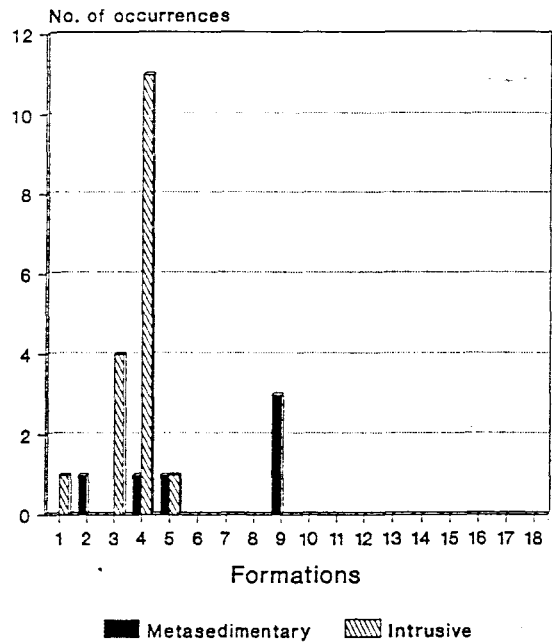
Ta CONCENTRATION



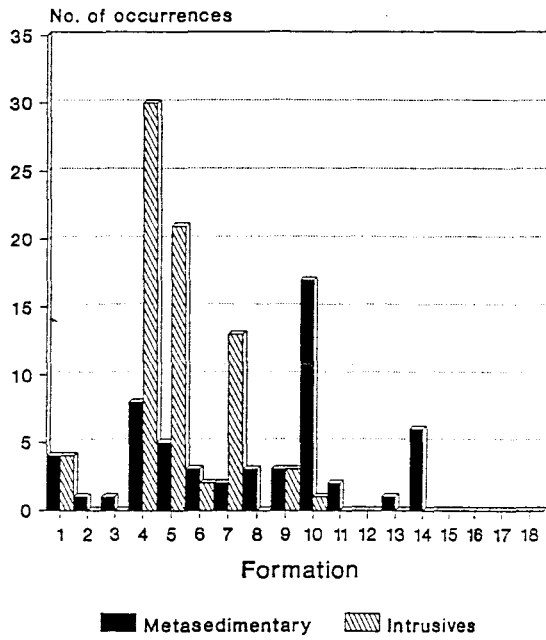
An CONCENTRATION



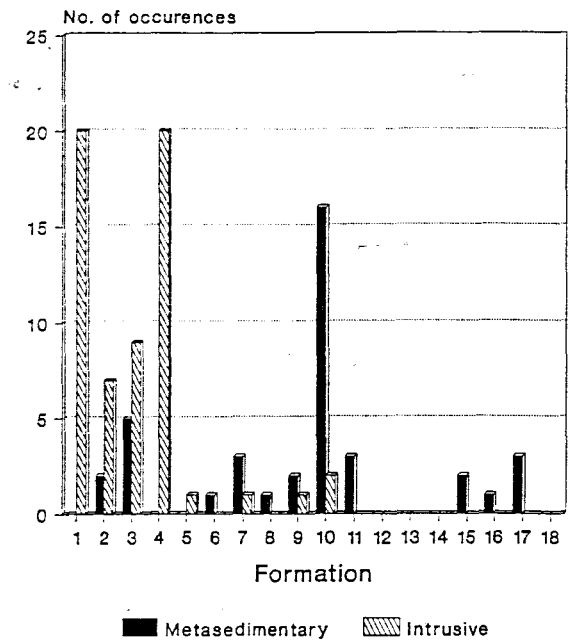
F CONCENTRATION



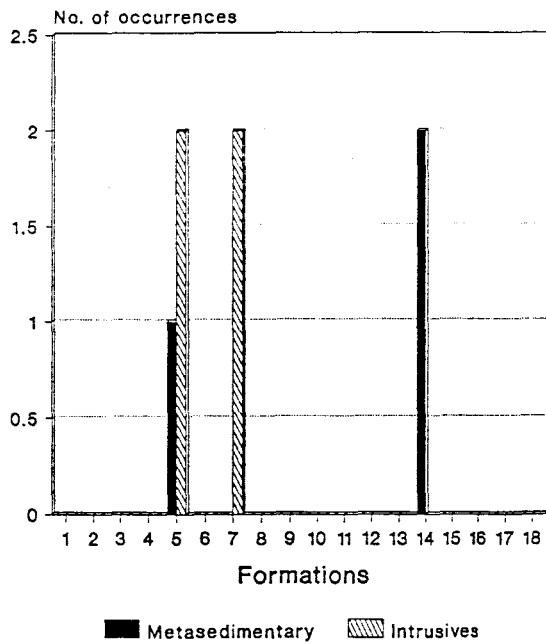
Be CONCENTRATION



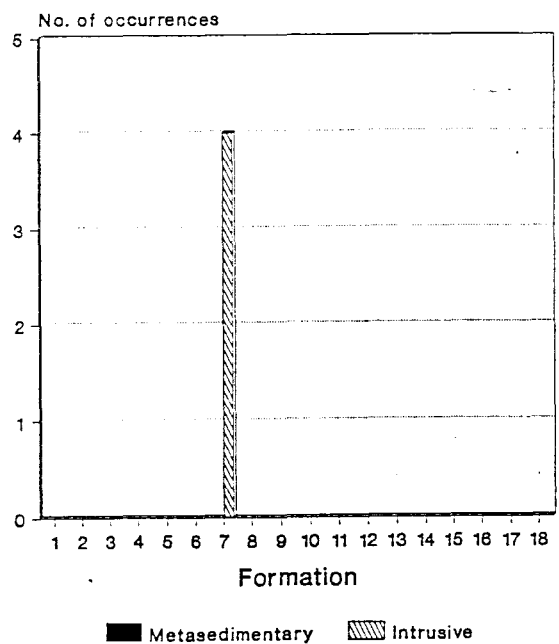
RE CONCENTRATION



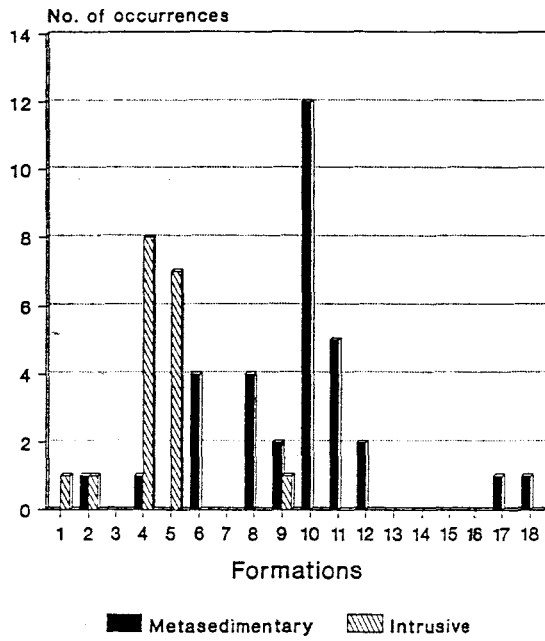
MICA CONCENTRATION



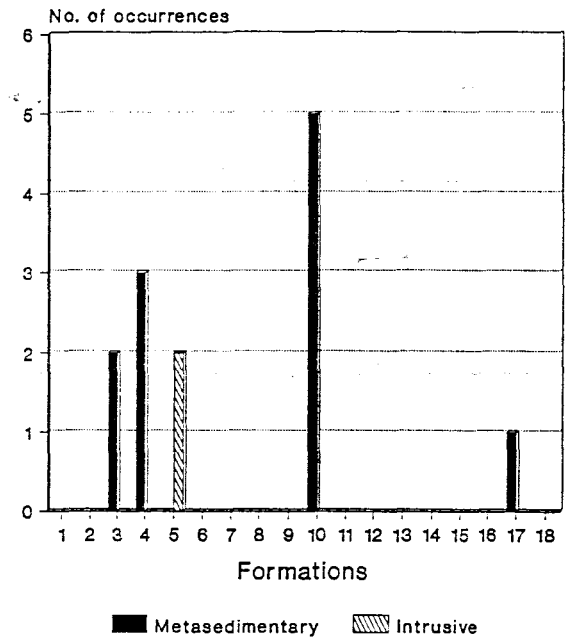
Li CONCENTRATION



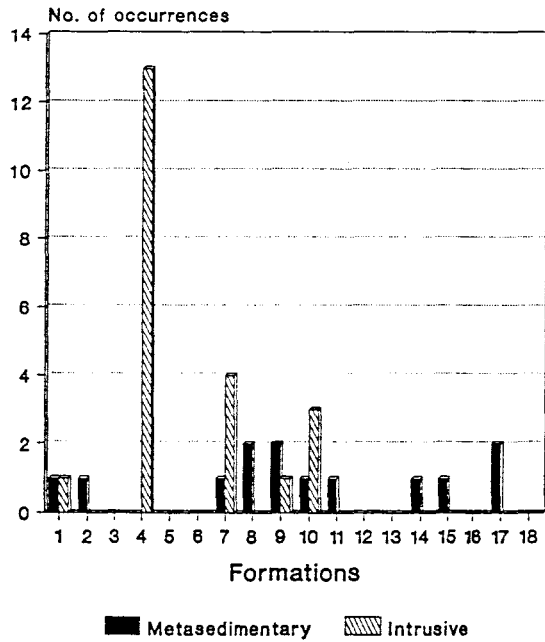
ROSE QUARTZ CONCENTRATION



CORUNDUM CONCENTRATION



MAGNETITE CONCENTRATION



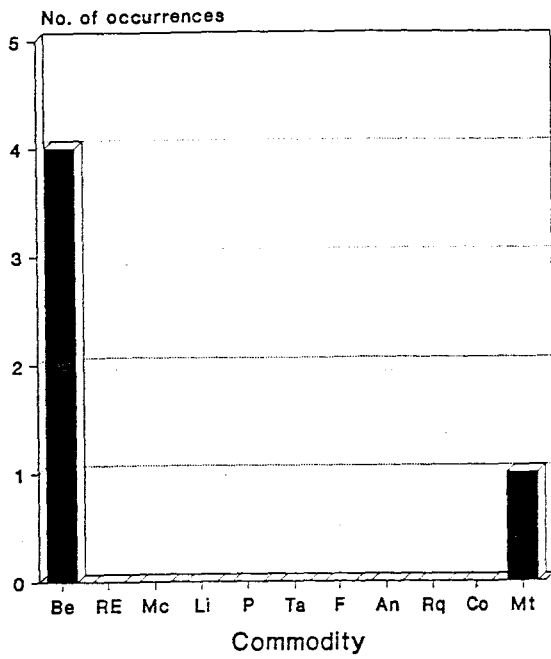
APPENDIX C

The graphs depict the amount of pegmatites within a specified formation that hosts various commodities. The symbols on the x-axis is explained in the following table:

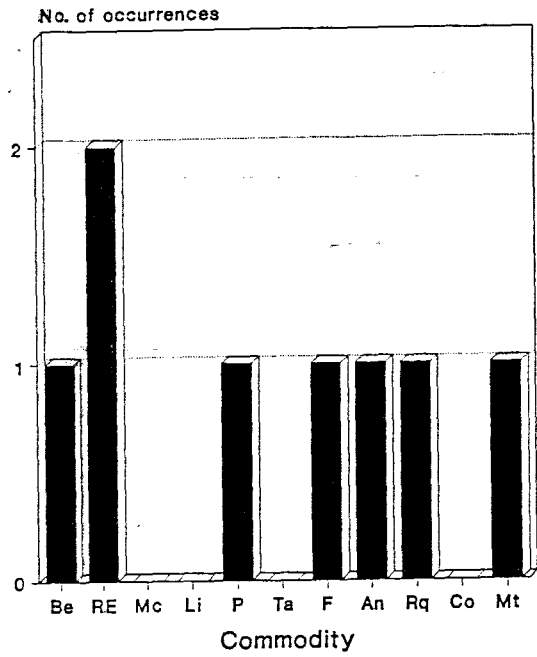
SYMBOL	EXPLANATION
Be	Beryl
RE	Rare Earth-bearing minerals
Mc	Mica
Li	Li-bearing minerals
P	Apatite
Ta	Tantalium-Niobium
F	Fluorite
An	Andalusite
Rq	Rose Quartz
Co	Corundum
Mt	Magnetite

KAKAMAS TERRANE

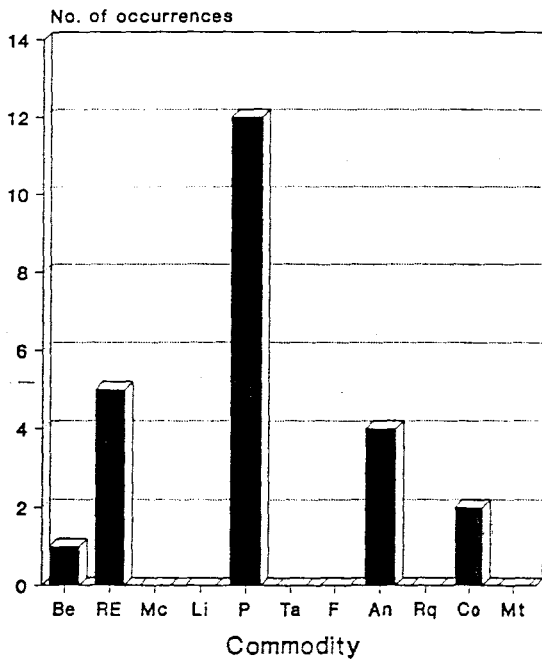
BOK-SE-PUTS FORMATION



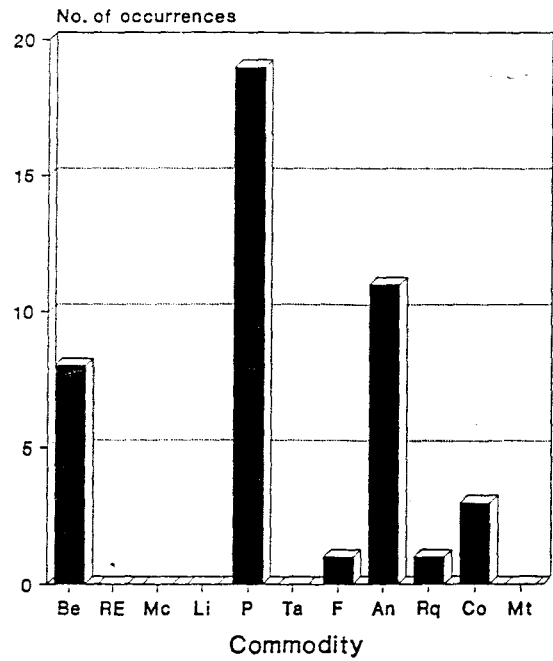
KOUROP FORMATION



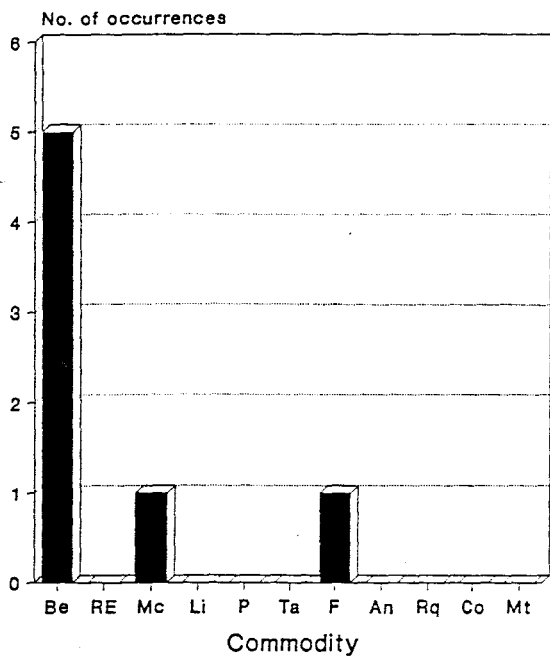
WITWATER GNEISS



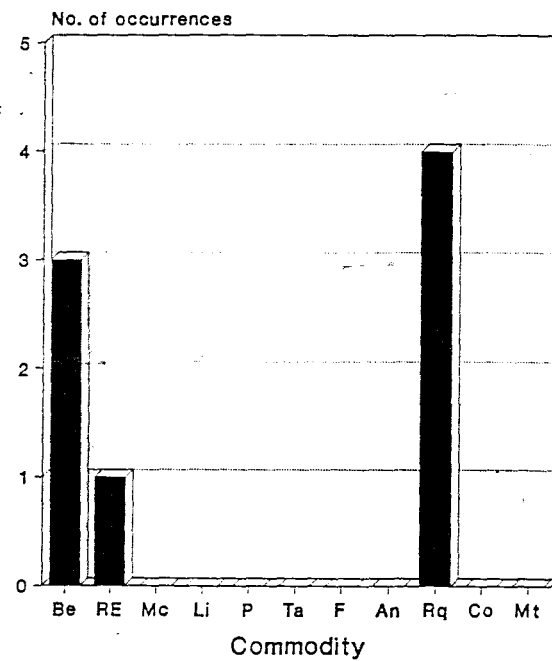
TWAKPUTS FORMATION



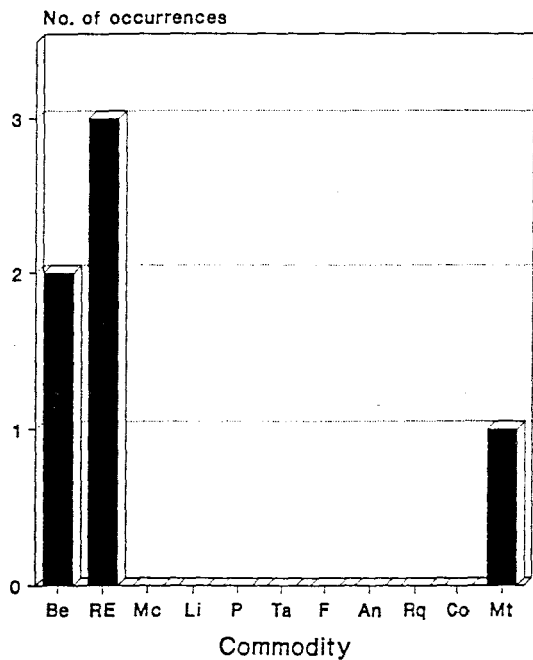
VALSVLEI FORMATION



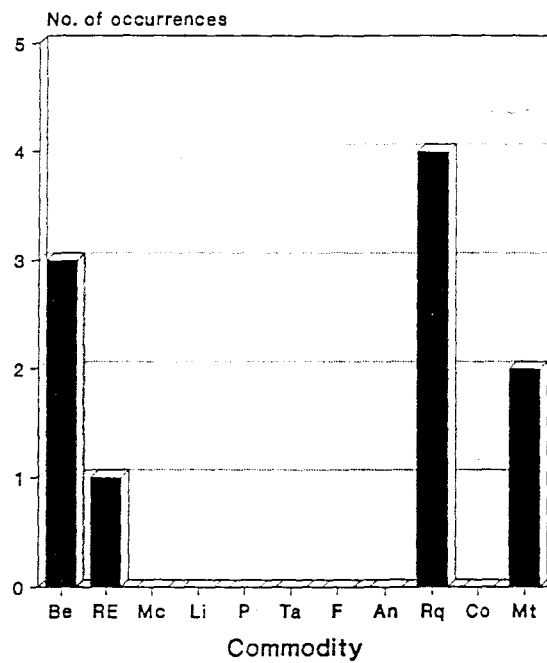
GANZENMOND FORMATION



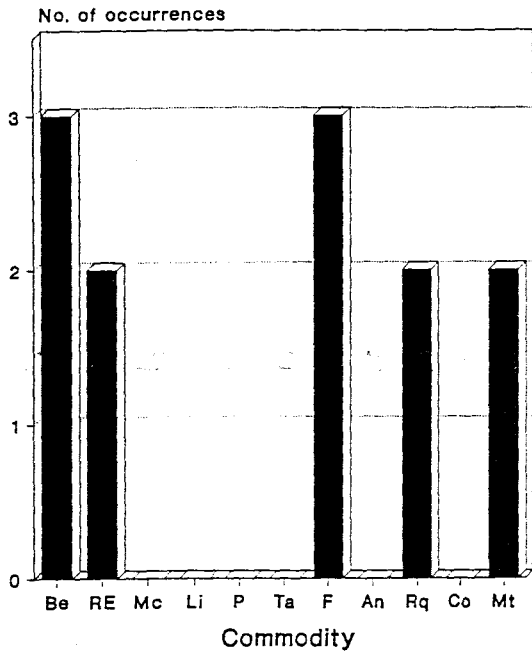
PUNTSIT FORMATION



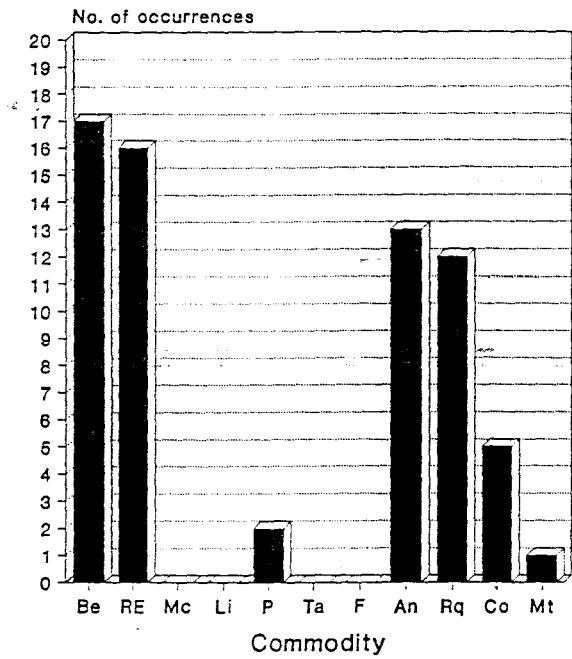
SANDPUTS FORMATION



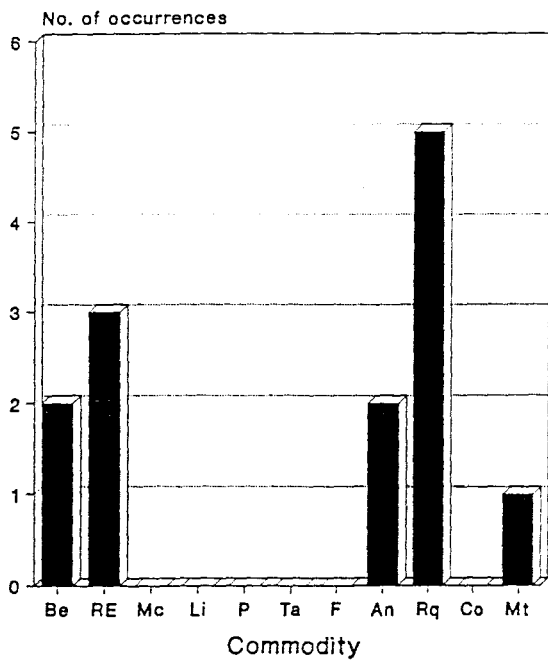
OMDRAAI FORMATION



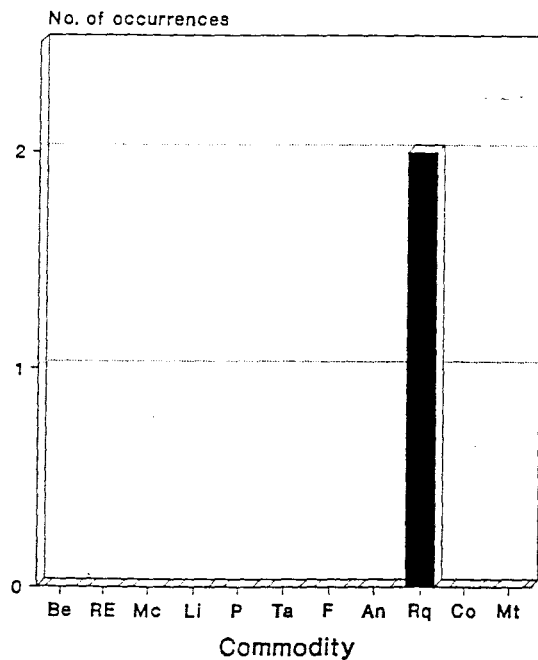
KENHARDT MIGMATITE



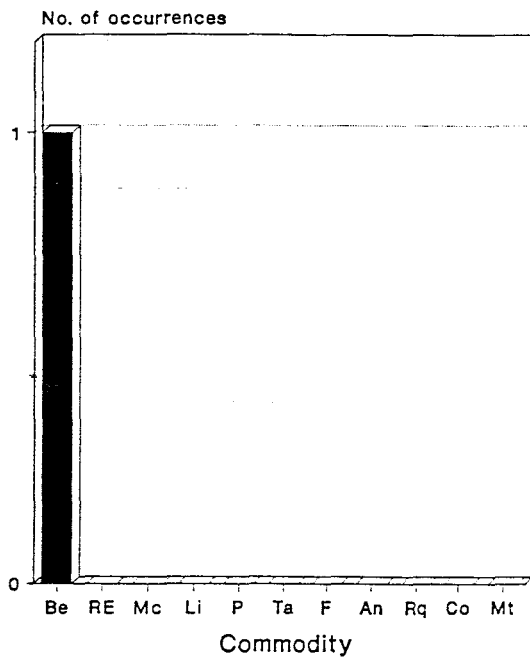
MOTTELS RIVER FORMATION



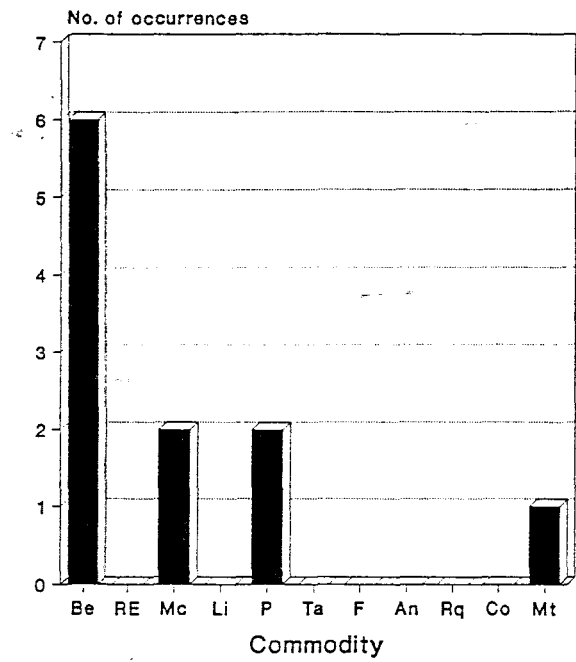
PUTSIES FORMATION



DRIEHOEK FORMATION

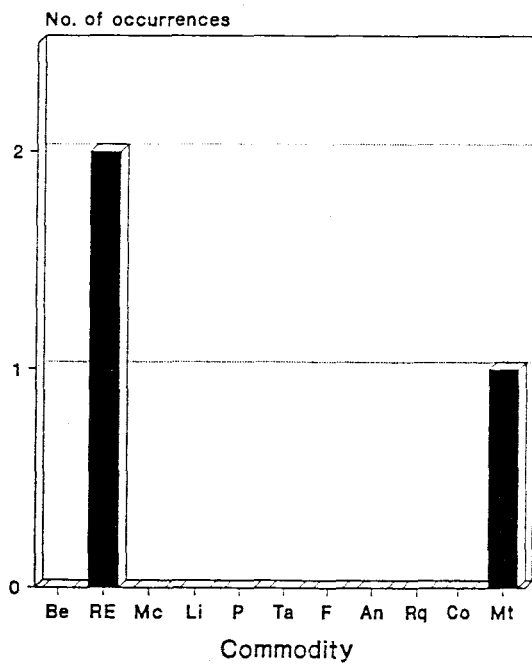


JACOMYNSPAN GROUP

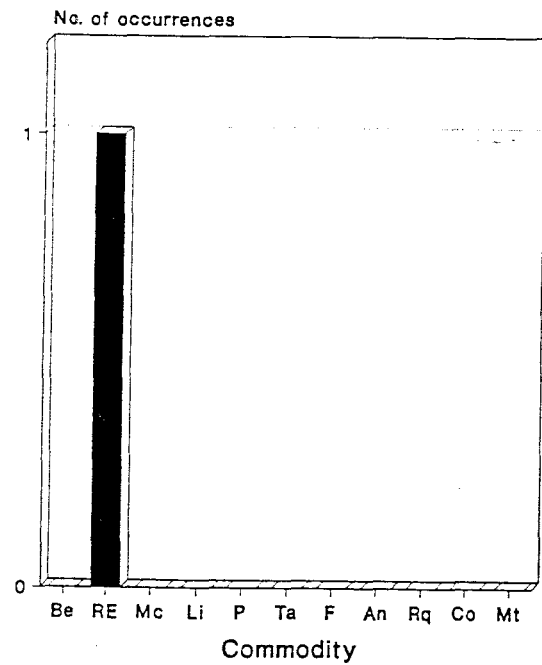


BUSHMANLAND SUBPROVINCE

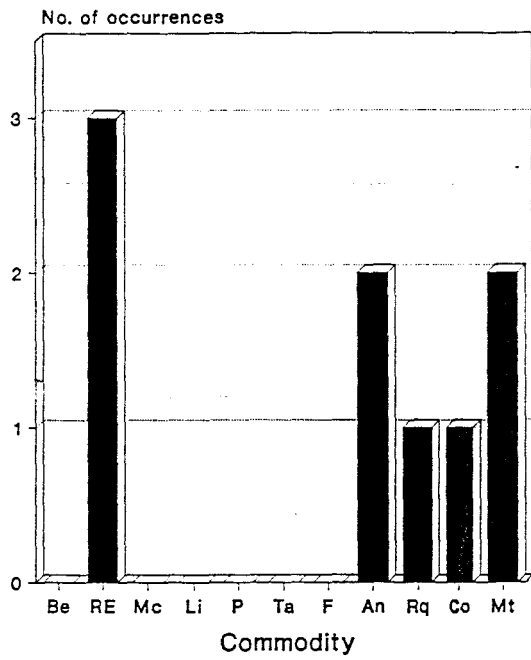
MODDERGAT GNEISS



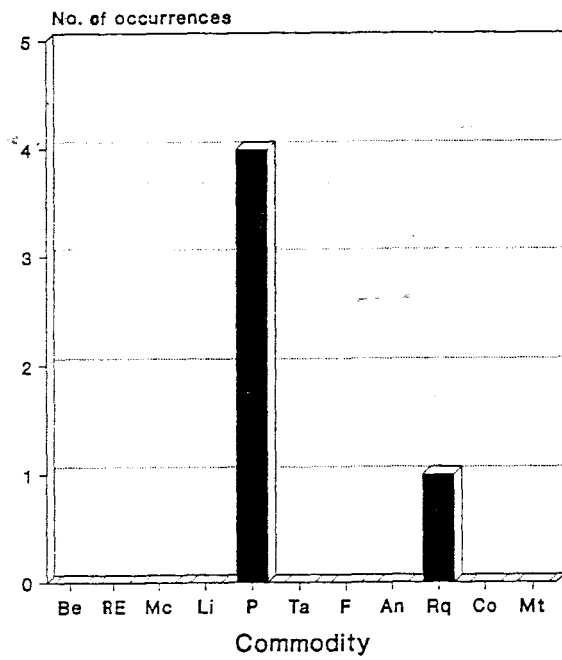
KOKERBERG FORMATION



ZANDBERGSHOOP FORMATION

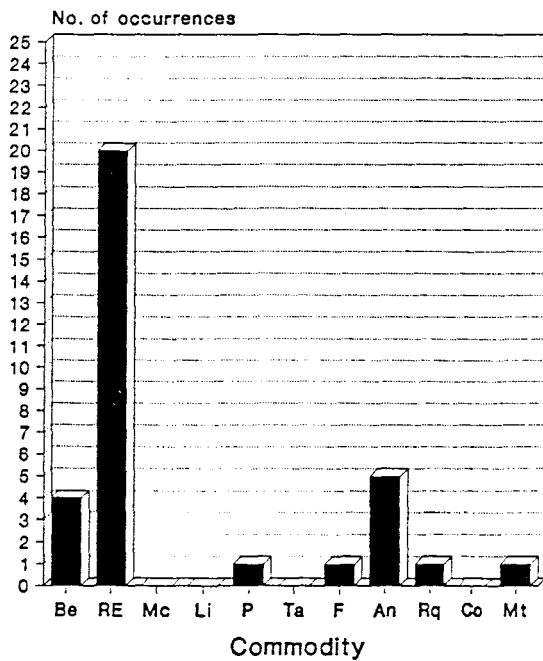


ARRIBEES GROUP

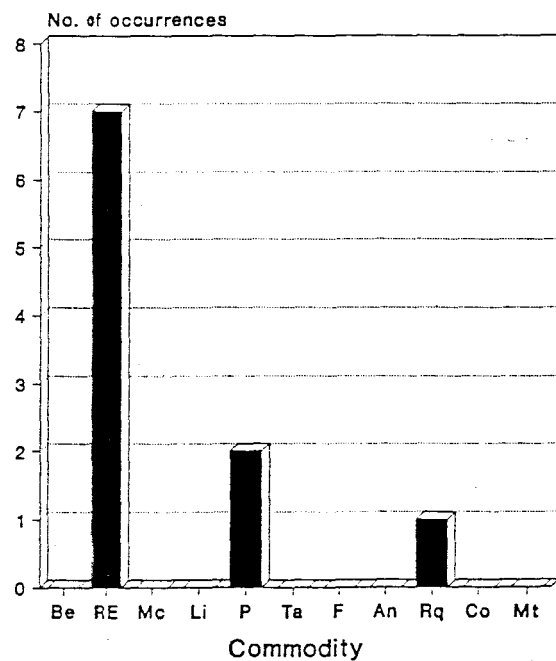


PRE TO SYN-TECTONIC INTRUSIVES

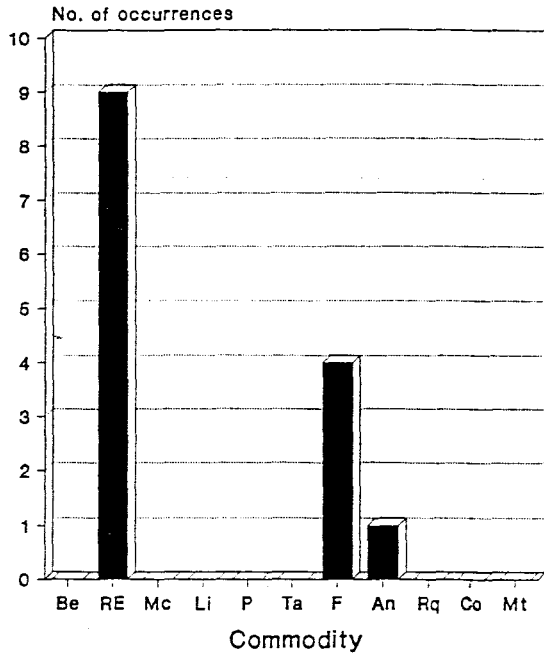
DONKIEBOUD GRANITE GNEISS



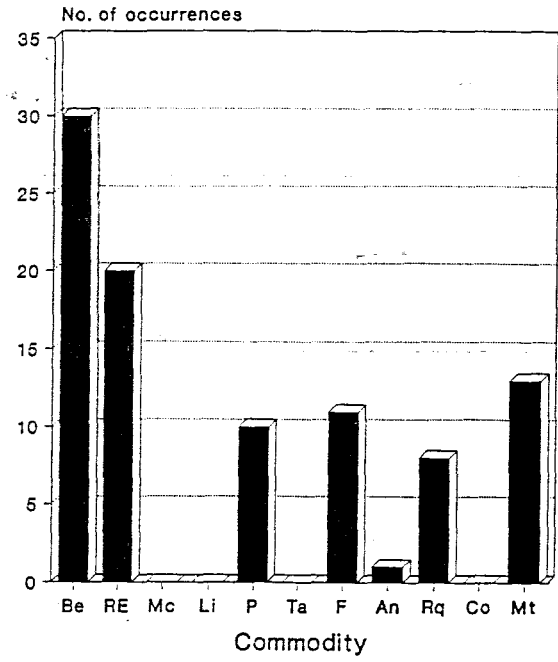
DABERAS GRANODIORITE



AUGRABIES GNEISS

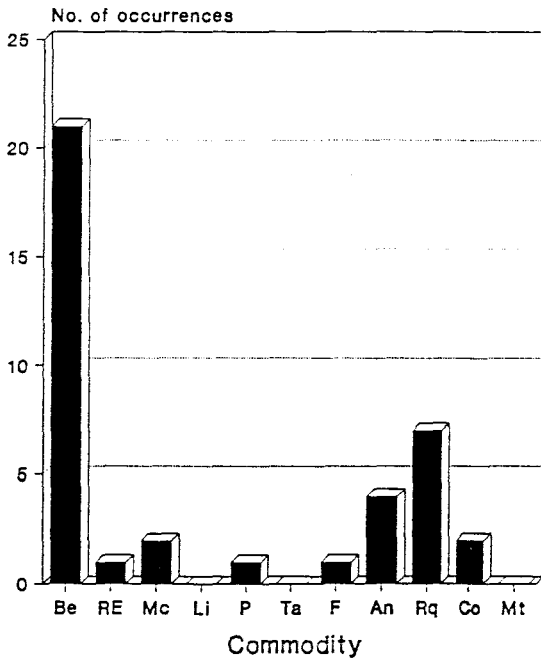


RIEMVASMAAK GNEISS

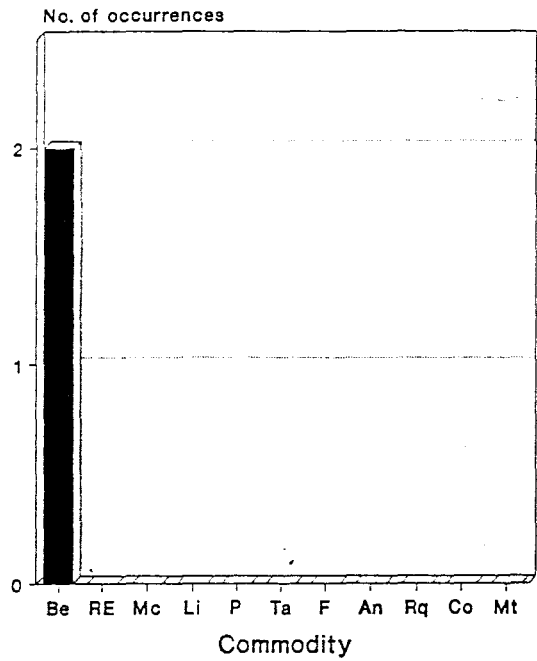


SYN TO LATE-TECTONIC INTRUSIVES

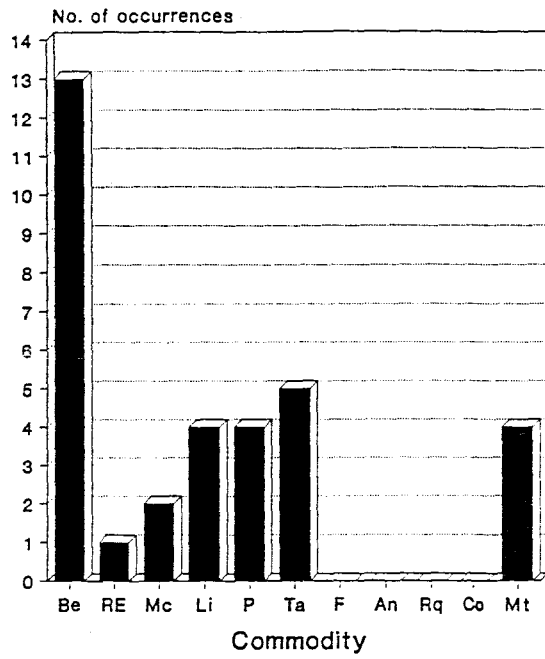
UNNAMED BASIC ROCKS



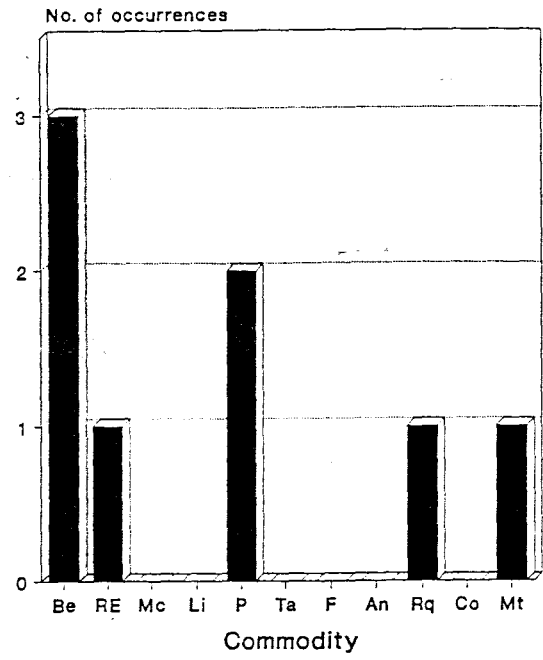
VAALPOTS GRANITE



ROK OPTEL GRANITE

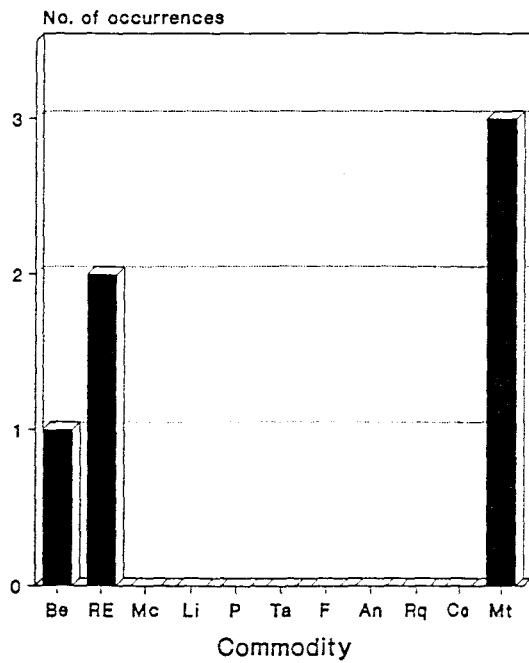


ELSIE SE GORRA GRANITE



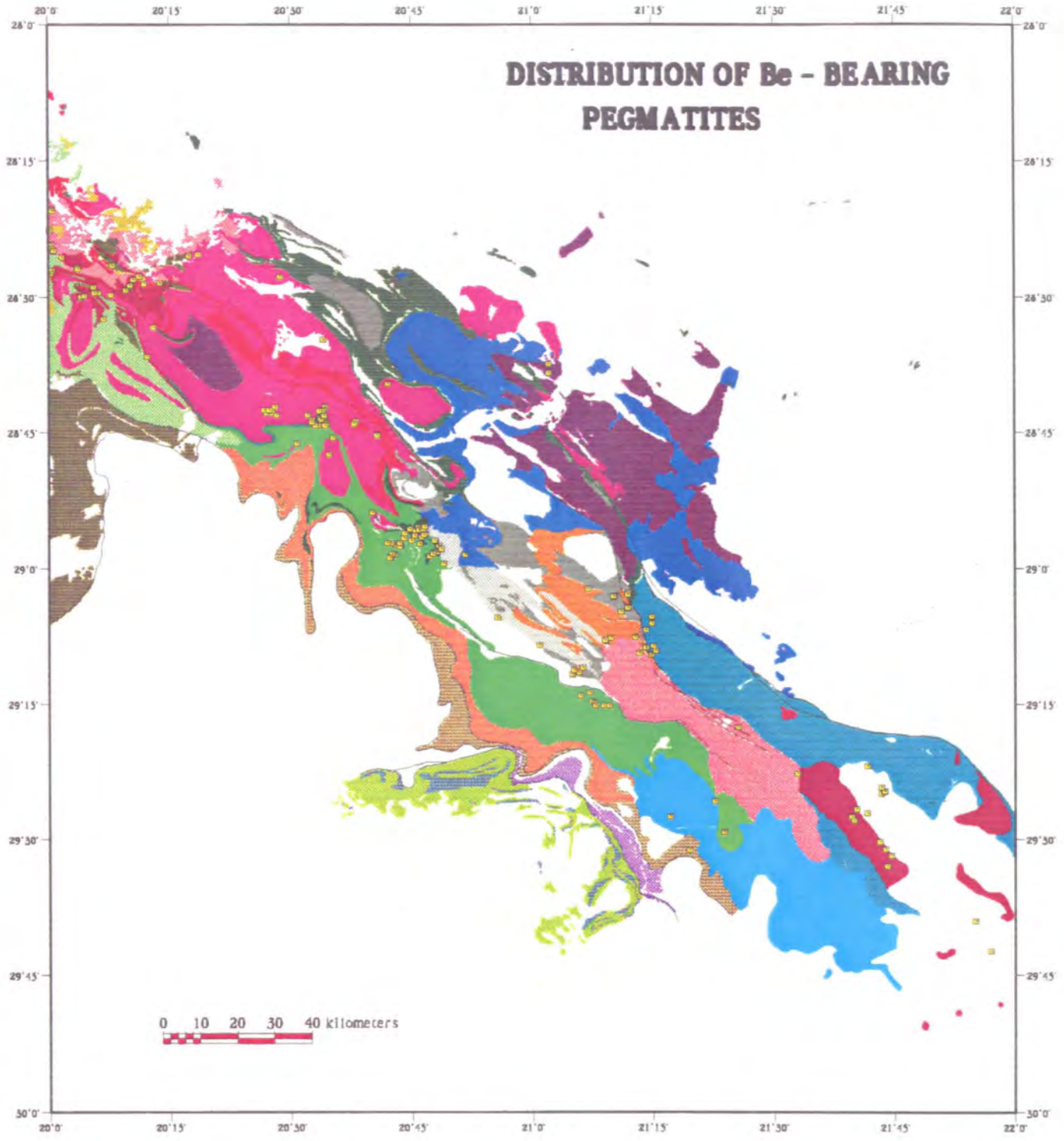
POST-TECTONIC INTRUSIVES

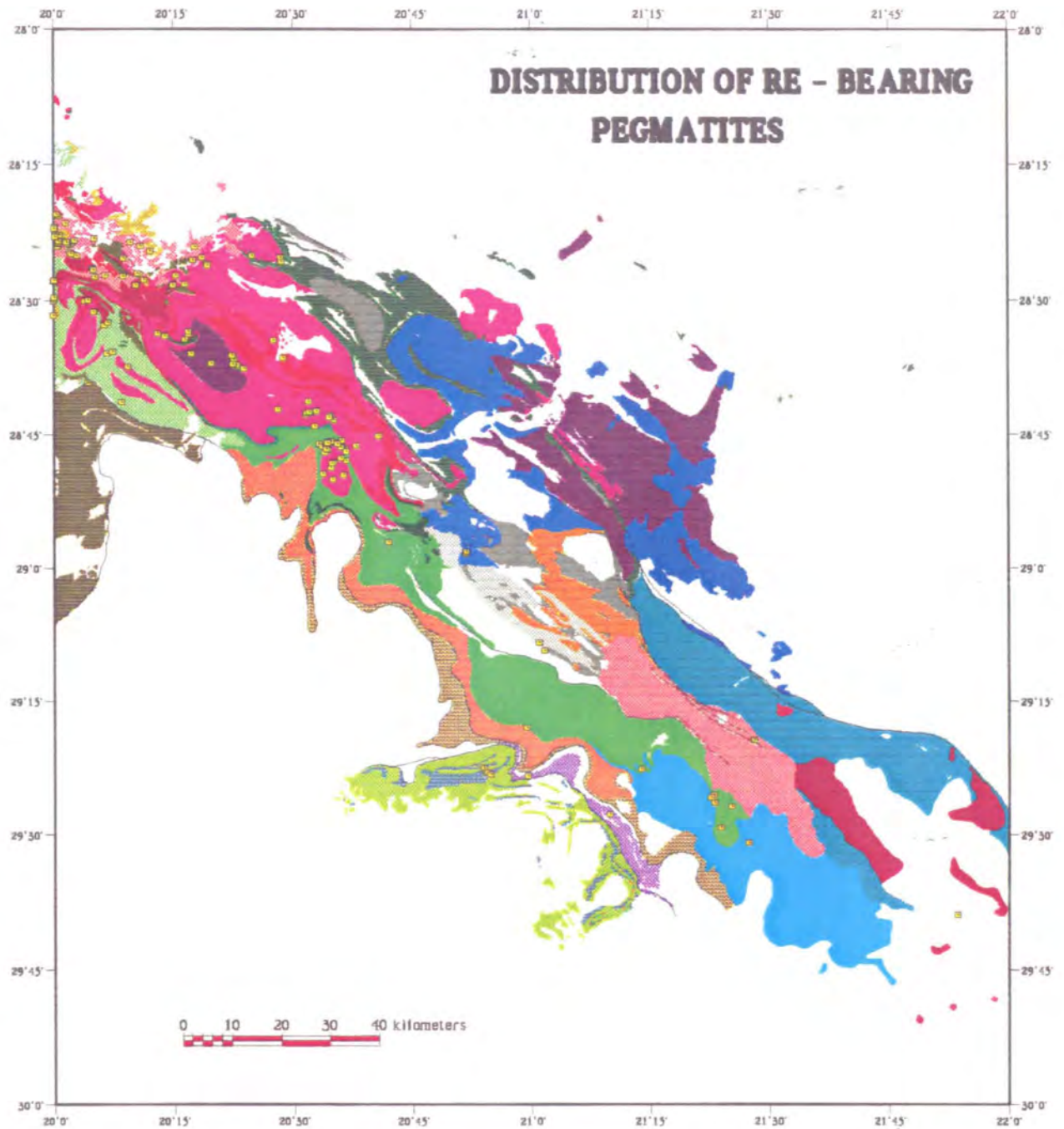
FRIERSDALE CHARNOCKITE

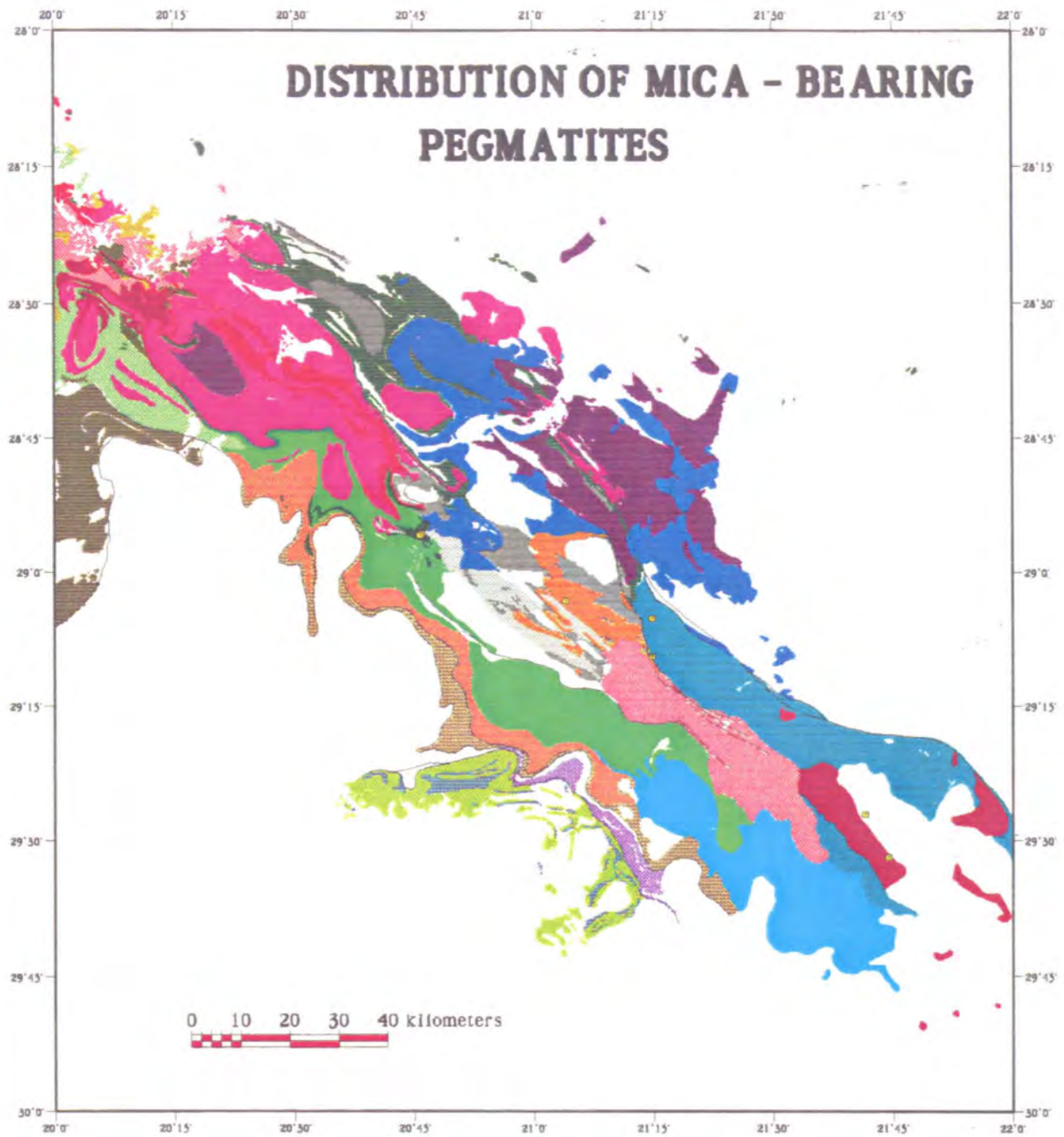


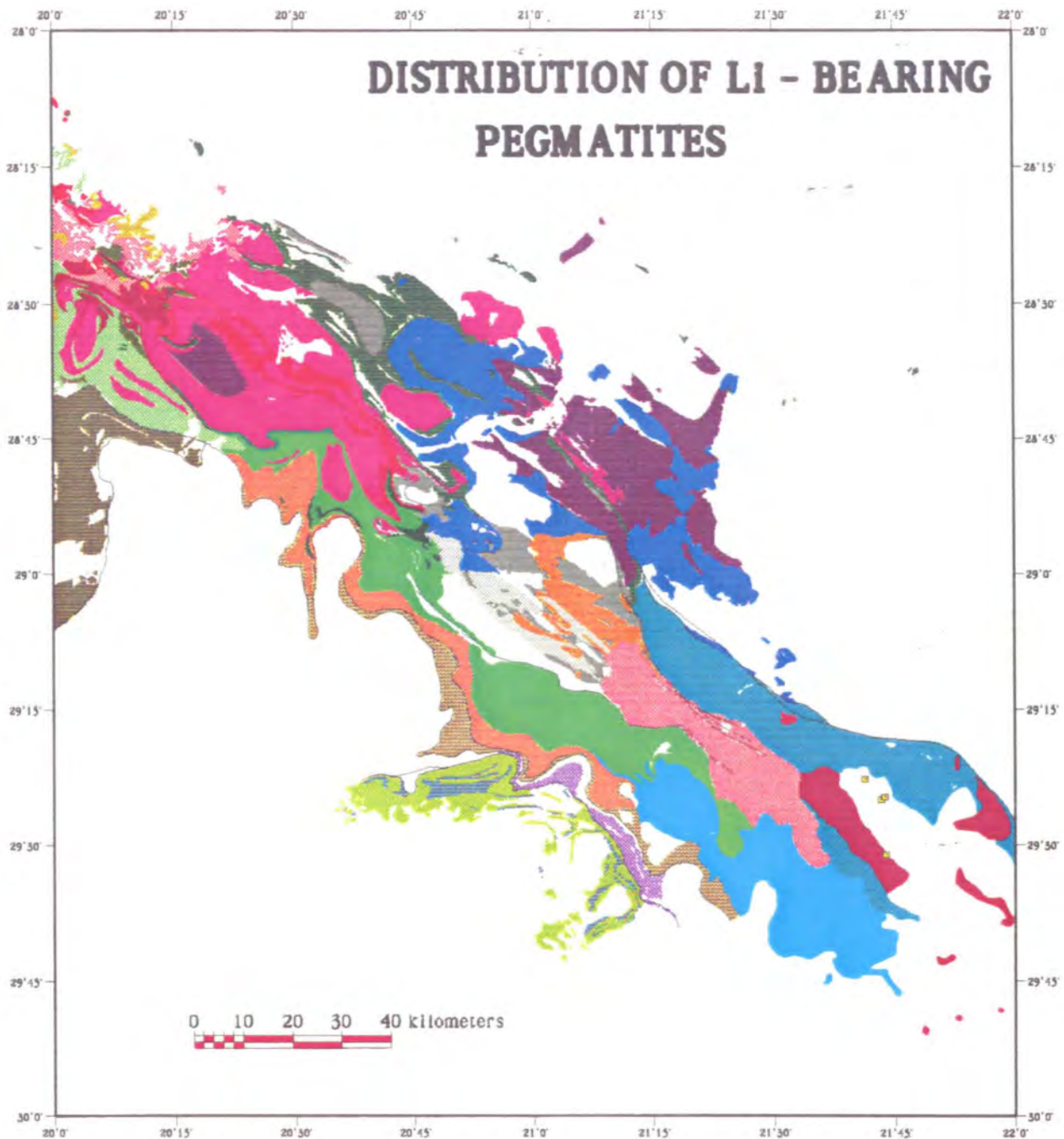
APPENDIX D

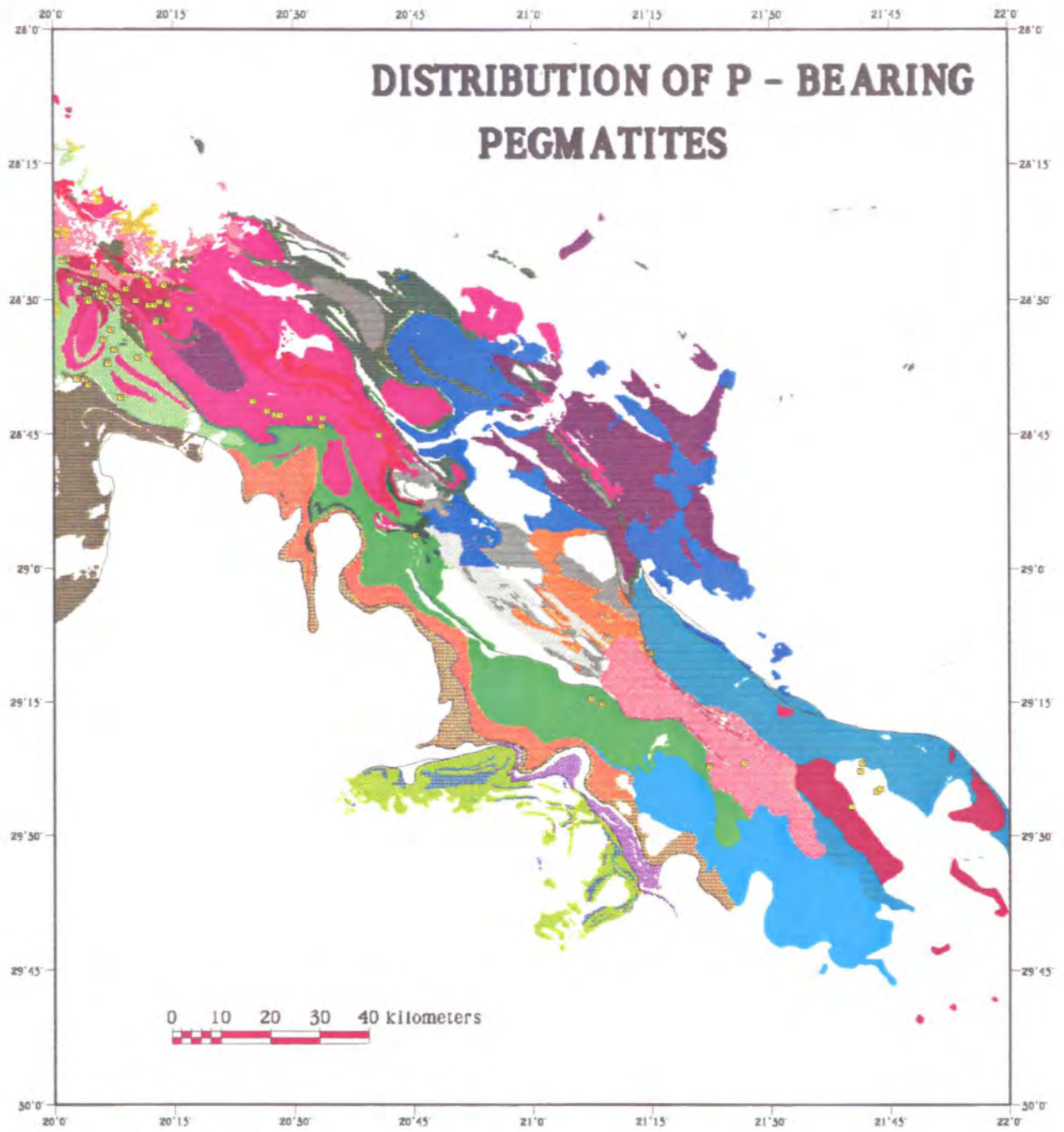
Maps showing the distribution of pegmatites that contain various commodities within the area under investigation. The formations shown are only those which contain pegmatites. For identification of the formations, the reader is referred to the legends in appendix E. For a description of the formations, the reader is referred to appendix A.

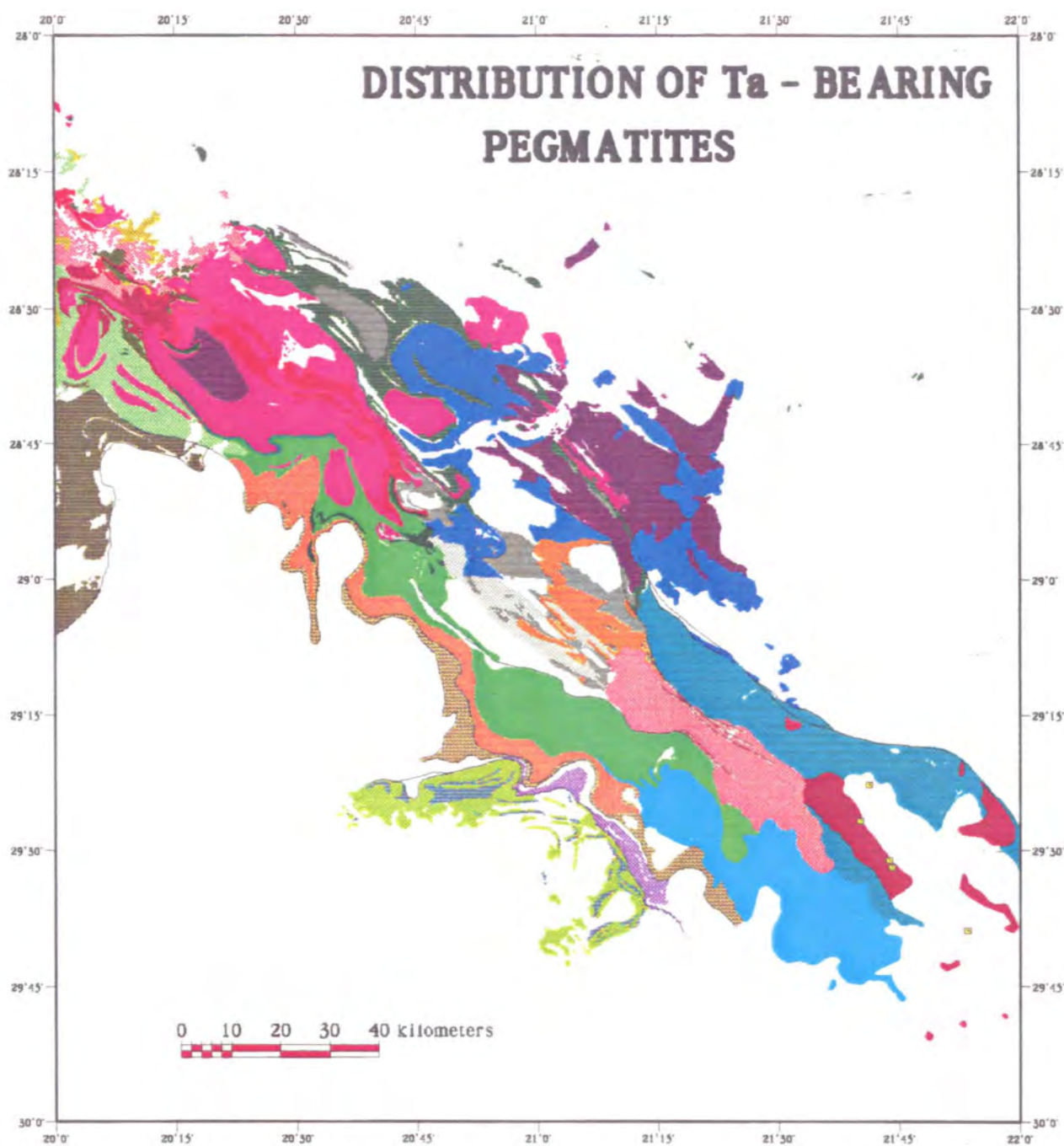


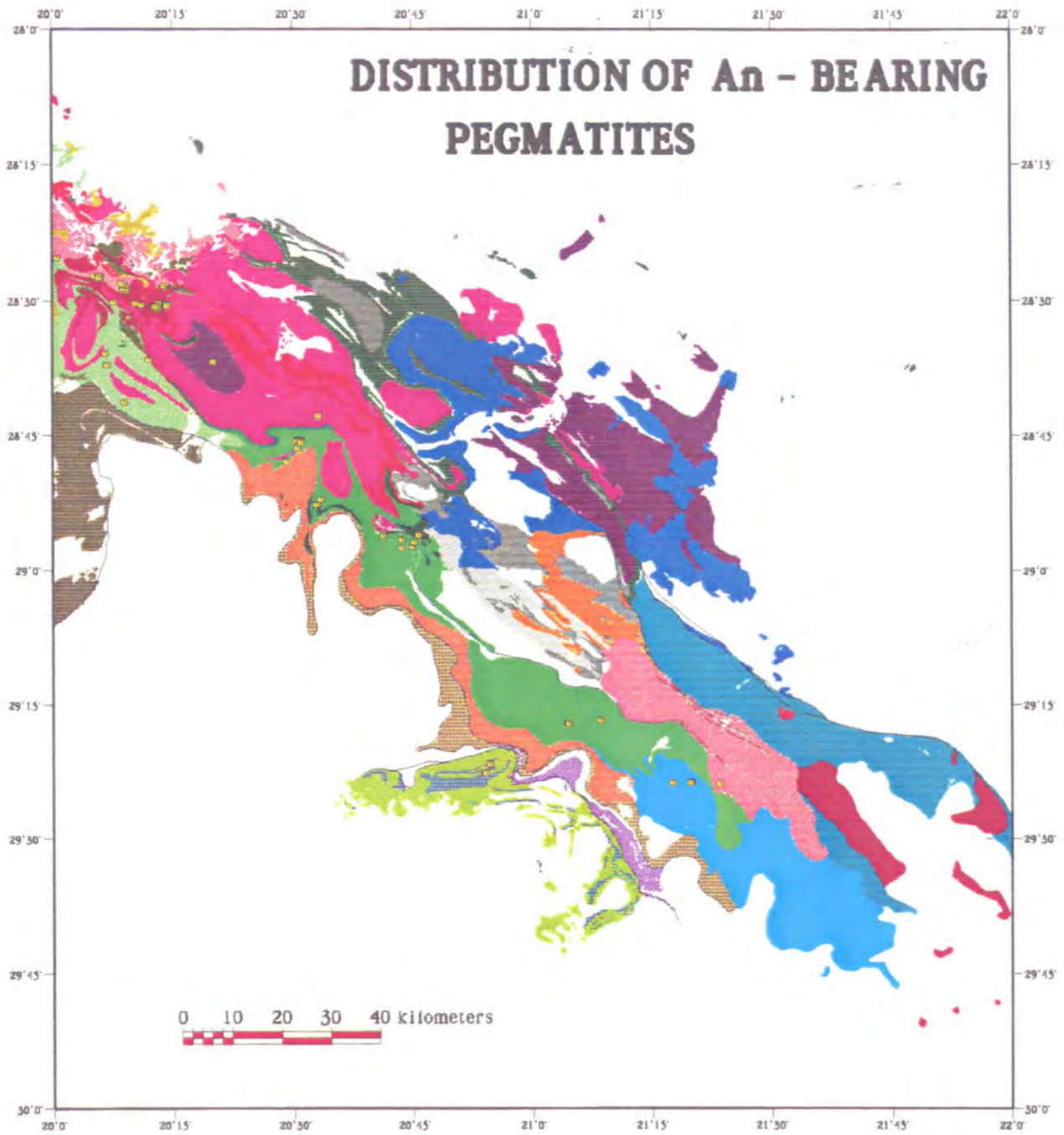


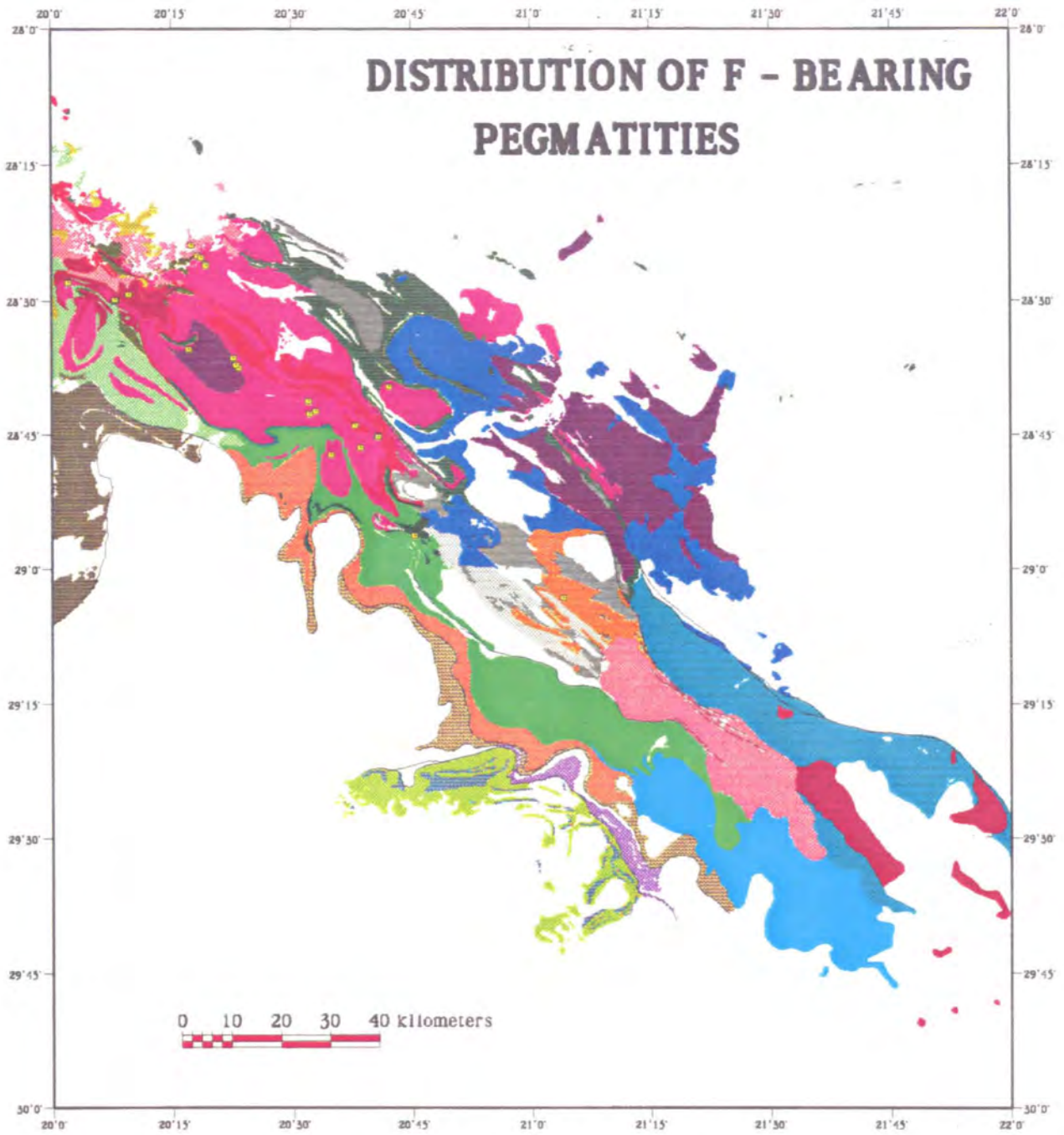


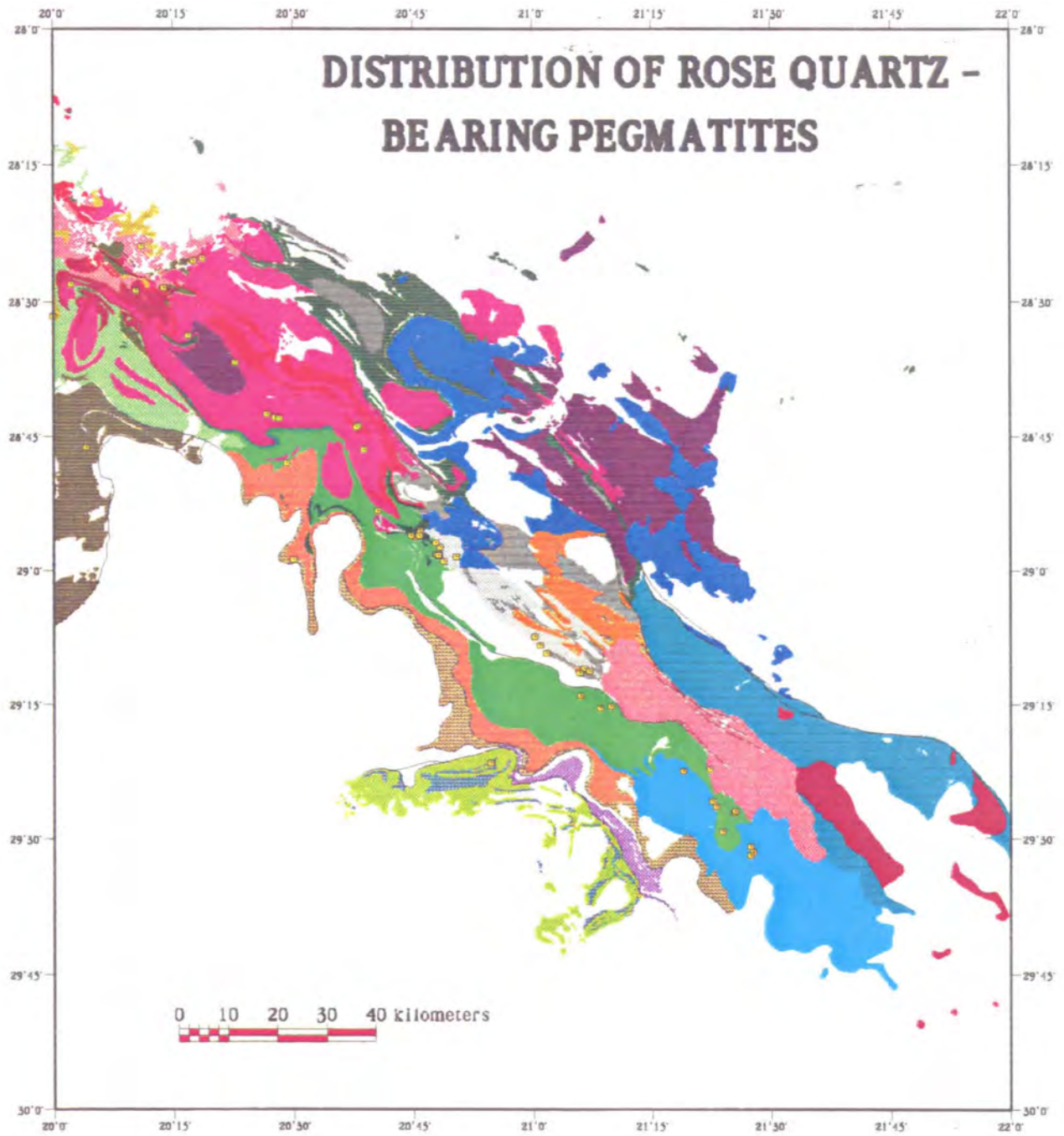


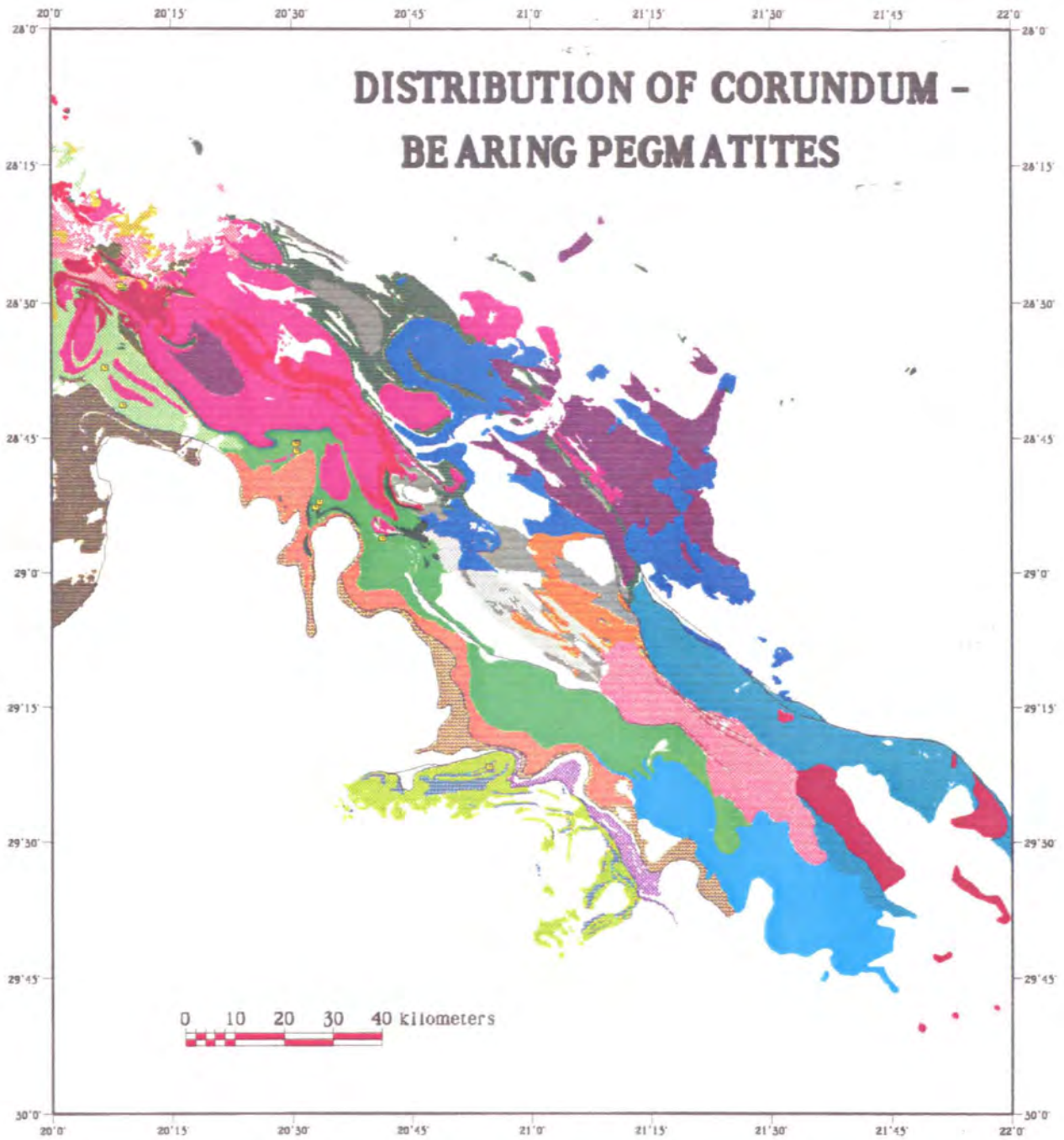


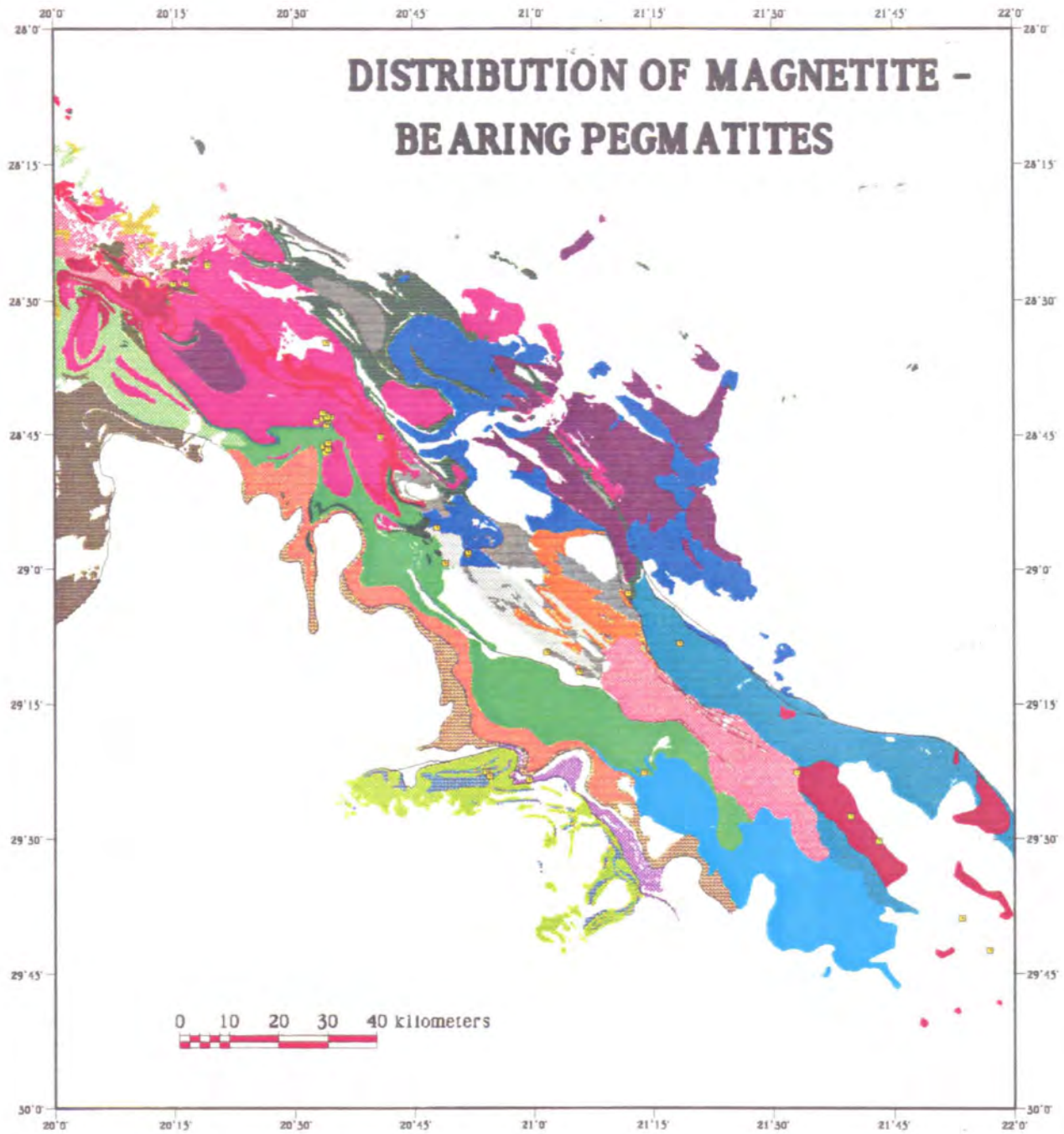










































APPENDIX E

Legends referable to all colour maps in this manuscript. Legends are divided into the various tectonostratigraphic terranes. Intrusive rocks appear on separate legends and are divided according to pre to syn-tectonic, syn to late-tectonic and post-tectonic suites.

LEGEND OF THE BUSHMANLAND SUBPROVINCE STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP
MOKOLIAN AND/OR OLDER	Mdb		De Banken Geniss	BRAKWATER
	Mnu		Nieuwe Post Wes Gneiss	METAMORPHIC
	Mha		Haakdoorn	SUITE
	Msd		Sandkoppies	
	Mp		Pollesberg	
	Msl		Slypstenkrans	
	Mmo		Moddergat Gneiss	
	Mko		Kokerberg	DE KRUIS
	Mz		Sandbergshoop	
	Ms		Soutputs	
	Mkr		Kraandraai	GRAPPIES
	Mbs		Bossiekom	
Mbr		Brulkolk		
Mka = Mk		Kameelputs		
Mr = Mrt		Rietput		
Mlo		Longsiekvlei		
Mdr		Droegrond	DROEBOOM	
Mkp		Klipvlei		
Ma			ARRIBEES	








LEGEND OF THE KAKAMAS TERRANE STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP	SEQUENCE
MOKOLIAN AND/OR OLDER	Mso		Sout River		
	Mgo		Goedehoop		
	Mva		Valsvlei	BIESJEPOORT	KORANNALAND
	Mgz		Genzenmond		
	Mra		Rautenbach se Kop		
	Mpu		Puntsit		
	Mt		Toeslaan		
	Msa		Sandputs		
	Mo		Omdraai		
	Mpi		Piet Rooisberg		
	Mre		Rhenosterkop Gneiss		
	Mkp		Koekoepkop		
	Mve		Venterskop		
	Msn		Sandroute		
	Mja			JACOMYNSPAN	
	Mcl		Collinskop	KOELMANSKOP METAMORPHIC SUITE	HARTBEES RIVER METAMORPHIC SUITE
	Mbo		Bok-se-puts		
	Mko		Kourop Migmatite		
	Mw		Witwater Gneiss		
	Mtw		Twakputs		
	Mw'		Wolfskop	VYFBEKER METAMORPHIC SUITE	
	Mhu		Hugosput		
	Mro		Rozynen Bosch		
	Mke		Kenhardt Migmatite		
	Mpe		Piet Rooi's Put Gneiss		
	Maa		Aasvogelkop Gneiss		
	Mpt		Putsies Gneiss		
Mm		Mottels River			
Mde		Driehoek			






LEGEND OF THE UPINGTON TERRANE STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP
MOKOLIAN	Nka		Kalkpunt	KORAS
	Na		Adesestad	
	N-Ml		Leeuwdraai	
	N-Mw		Welgevind	
	Mru		Rouxville	
	Me		Ezelfontein	
	Mrp		Rusplaas	
	Msr		Swarkopsleegte	
	Mst		Steenkampsputs	
	Mbs		Bossienek	
	Mbm		Boom River	
	Mch		Christiana	
MOKOLIAN AND/OR OLDER	Mle		Leerkrans	WILGENHOUTSDRIF
	Mz		Zonderhuis including Grootdrink member	
MOKOLIAN AND/OR OLDER	Msu		Sultanaoord	VAALKOPPIES
	Mda		Dagbreek	
	Mk		Khaboom	

LEGEND OF THE POST PROTEROZOIC STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP	SEQUENCE
QUATERNARY	Q		unnamed superficial deposits	KALAHARI	
	Qg		Gordonia		
TERTIARY	T		unnamed calcrete		
CARBONIFEROUS TO PERMIAN	Pp		Prince Albert	ECCA	KAROO
	C-Pd			DWYKA	
NAMIBIAN	Nv		Fish River	NAMA	
	Ns		Swartzrand		
	Nk		Kuibis		















LEGEND OF THE AREACHAP TERRANE STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP	SEQUENCE
MOKOLIAN	Mj		Jannelsepan		AREACHAP
	Msp		Sprigg		
	Mbe		Bethesda		
	Mr		Ratel Draai		
	Mvn		Van Wyks Pan		
































LEGEND OF THE KHEIS PROVINCE STRATIGRAPHY

TIME	SYMBOL	COLOUR	FORMATION	GROUP
MOKOLIAN AND/OR OLDER	Mgh		Grobiershoop	BRULPAN
	Mu		Uitdraai	
	Mpr		Pynnsberg	












LEGEND OF PRE- TO SYN- TECTONIC INTRUSIVES

TECTONOSTRATIGRAPHIC UNIT	SYMBOL	COLOUR	NAME	SUITE	
KAKAMAS	Mdo		Donkieboud Granite Gneiss	EENDOORN	
	Mda		Bak River Granite Gneiss		
	Md		Daberas Granodiorite		
	Ma		Augrabies Gneiss		
	Mkm		Kakamas Suid Gneiss		
	Mrm		Rlemvasmaak Gneiss		
	Msw		Swanartz Gneiss		
	Mdy		Dyasons Klip Gneiss		
	Mkl		Klip Bakken Gneiss		
	Mcu		Curries Camp Gneiss		
	Mlu		Lutzputs Gneiss		
	BUSHMANLAND SUBPROVINCE	Mln			LANGE KOLK
		Mbn		Banks Vlei Gneiss	
KHEIS PROVINCE	Mkk		Kalkwerf Gneiss		

LEGEND OF SYN- TO LATE- TECTONIC INTRUSIVES

TECTONOSTRATIGRAPHIC UNIT	SYMBOL	COLOUR	NAME	SUITE		
AND AREACHAP	Mkn		unnamed granite	KEIMOES		
	Mgu		Gouskop Granite			
	Mn		Neilers Drift Granite			
KAKAMAS TERRANE	Mka		Kanoneiland Granite			
	Ms		Straussberg Granite			
	Mco		Colston Granite			
	Mkb		Keboes Granite			
	MI		Louisvale Granite			
	Mv		Vaalputs Granite			
	AND AREACHAP,UPINGTON	Mk			Kleinbegin Granite	
		Mks			Klip Koppies Granite	
MKL			Klipkraal Granite			
Mge			Gemsbokbult Granite			
Msk			Skeerhok Granite			
MI ¹			Liefdood Granite			
Mn ¹			N'Rougas Granite			
Mrk			Rok Optel Granite			
Mb			Brussel Granite			
Mpy			Pypklip Granite			
Md			Dwaalgeest Granite			
Msw			Swartputs Granite			
Mbv			Boven Rugzeer Granite			
Mkn ¹			Klein Van Wyks Pan Granite			
Me			Elsie se Gorra Granite			
Mh			Hartebeest Pan Granite			
Mga			unnamed basic rocks			
BUSHMANLAND SUBPROVINCE		Mla			Lat River Granite	TOUBEP
		Mba			Bakoondsvlei Granite	
		Mdk			De Bakken Granite	
	Mvl		Vaalhoek Granite			
	Mto					

LEGEND OF POST TECTONIC INTRUSIVES

TECTONOSTRATIGRAPHIC UNIT	SYMBOL	COLOUR	NAME	SUITE
	Jd		Dolerite	
UPINGTON TERRANE	Nb		Blaauwbosch Granite	
	Nro		Rooiputs Granophyre	
	Mbt		Betadam Gabbrorite	
	Mf		Friersdale Charnockite	KEIMoes
	Mc		Crydas Granite	
	Mg		Gif Berg Granite	
	Mvr		Varsput Granodiorite	
BUSHMANLAND SUBPROVINCE	Mbj		Basjan Granite	
	Mui		Uitkoms Granite	
	Msb		Sandbakken Metabasite	