PEGMATITE-HOSTED MINERAL DEPOSITS OF CENTRAL AND SOUTHERN AFRICA: REGIONAL GEOLOGICAL SETTINGS AND PRELIMINARY EXPLORATION TARGET CONSIDERATIONS

by

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by

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To our children;
Khangeziwe, Aubrey, and Grant,
but,
most of all,
to my wife Drinnie,
who looked after us all
at a time
when she needed looking after herself

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To Professor John Moore I say, "What a bunch of accents and backgrounds and characteres to cope with?!"

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Abstract

Review of literature shows that rocks of all ages from the Archaean to the early Phanerozoic host pegmatite-based mineralisation in the central and southern region of the African continent. The greatest concentrations occur in the Archaean and late Proterozoic orogenic belts, while early to middle Proterozoic granites do not, in general, host mineralisation. Pan-African mineralisation is present, but is not widespread. Some deposits deposits previously considered to be of Pan-African have been shown to be of Proterozoic age.

In common with occurrences of other regions, the deposits are closely associated with small, late- to post-tectonic granites. Therefore, preliminary assessment of the potential of granites as sources of pegmatite mineralisation should utilise satellite data or aerial photographs. The granites also tend to be alkaline and peraluminous. Thus, in the next stage, chemical analysis for selected major and trace elements should be conducted directly on granites if they occur as small plutons. For large granitic batholiths or terranes, preliminary stream-sediment surveys may be necessary to reduce the size of the target area. Subsequently, pegmatite zonation around suitable granites should be assessed as it allows attention to be focussed on areas likely to host the specific type of mineralisation being explored for.

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1. INTRODUCTION

This report is a review of mineral deposits hosted by felsic or granitic pegmatites in the central and southern region of the African continent (Fig. 1.1). The aim is to identify the main geological provinces which host the mineralised pegmatites, so that some preliminary suggestions concerning exploration-targetting can be made for the region.

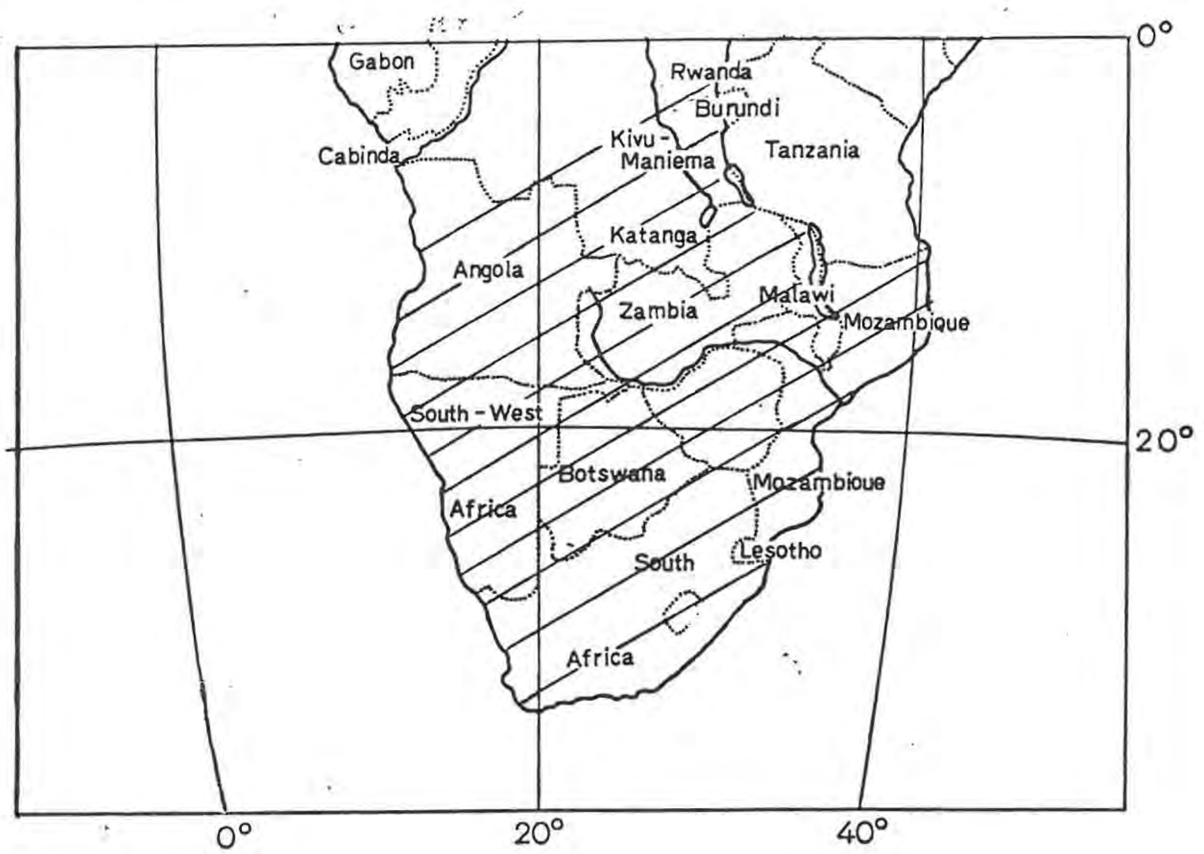

FIGURE 1.1 Location of the area of study (hachured)

The approach adopted takes cognissance of the fact that mineralised pegmatites occur in specific settings, which have occurred at definite stages of the earth's evolution. Firstly, in Section 2, the nature and occurrence of pegmatitic mineral deposits is reviewed briefly and on a global basis. This stage also considers in general the sources, and modes of mobilisation, concentration and localisation of the pegmatites, and the ore materials they host. As the pegmatites of the region are similar to those of other parts of the world, these common aspects are not discussed again in any detail in subsequent sections.

Section 3 briefly summarises the regional geology and structural units. The summary sets the context and order in which the rest of the review was carried out and is reported.

Thirdly, in sections 4 to 8, geological settings known to host deposits in the region are identified, and available information on the general geology and local setting of mineralisation is summarised. Examples are used to illustrate the major features of the mineralisation concerned. Even parts of the region's stratigraphy which are currently not known to host significant deposits are reviewed briefly, as it is possible that mineral deposits, though not discovered yet, may be present.

Information thus obtained is synthesised and used in Section 9, in which areas are ranked depending on the chances of finding new deposits. Common techniques used in preliminary selection of areas for felsic or granitic pegmatite deposits are then recalled in Section 10, and their possible applications in the region discussed in Section 11. The discussion of regional exploration targets in this report stops at the stage when an area is detected to host granitic pegmatite mineral deposits, say by location of a body with the mineralisation of interest, and detailed mapping or preliminary drilling has been assessed to be necessary. Discussion of evaluation of individual fields or pegmatite bodies beyond this is not entered into.

2. OVERVIEW OF PEGMATITES

2.1 DEFINITION

Felsic pegmatites are igneous rock bodies which approximate granites in chemical and mineralogical composition. They occur as segregations or small intrusions in areas of high-grade metamorphism and granitic intrusions. A genetic relationship with the granitic intrusions can often be demonstrated, with the pegmatites being hypogene separations or injections of liquids enriched in water and large, lithophile elements (Cerny, 1982a and b).

2.2 GEOCHEMICAL CHARACTERISTICS

Although similar to granites in general, felsic pegmatites have a lower CaO content, a higher K/Na ratio and Al_2O_3 content. Their trace element characteristics also differ, with the pegmatites showing lower contents of Cr, Co, Ni, Ag, Se, Te, V and precious metals. The pegmatites are also richer in the volatile rock components such as H_2O, CO_2, F, Cl, P and B (Cerny, 1982a and b).

2.3 CLASSIFICATION

Proposed schemes of classification of pegmatites have been based on major element chemical composition (Goldschmidt, 1930), mineralogy (Vlasov, 1952; Smirnov, 1976), bulk chemical composition, trace element chemical composition (Landes, 1933) and internal and external structural and lithological associations (Gevers, 1937). For purposes of exploration, a combination of several of these is desirable. A scheme often adopted in initial stages of exploration takes into account the depth of formation of the pegmatites as indicated by the metamorphic assemblages developed in the host rocks. The scheme works where it can be demonstrated that peak metamorphism occurred in the same thermo-tectonic event which produced the pegmatites. The type of mineralisation and field characteristics of the pegmatites correlate with the metamorphic facies or depth of emplacement (Table 2.1).

2.4 MATERIALS RECOVERED

The elements which may occur in pegmatites in economic concentrations are boron, lithium, beryllium, tantallum, niobium, tin, uranium, thorium and rare-earth elements. The minerals in which the elements are concentrated are shown in Table 2.2. Industrial minerals and gemstones have been included in Table 2.1.

TABLE 2.1 Classification of Pegmatites by Depth of Emplacement

DEPTH	NAME	ECONOMIC MINERALS OR ELEMENTS	FIELD CHARACTERISTICS
Shallow (1.5-3.6km)	Miarolytic pegmatites topaz),optic	quartz, gems(beryl + unzoned fluorite	intrude low-grade hosts,
Moderate (3.5-7km)	RE-pegmatites	Li, Be, Cs, Ta, Sn, Nb	fill fractures in host while under cordierite amphiblt facies
Very deep (7-11km)	Mica-Pegmatites hosts, may be anatectic	mica, some RE and ceramic feldspar	into almandine amphiblt facies
Extreme depth (11km+)	Migmatitic Pegmatites	allanite, monazite,	granulite facies hosts (often no economic minerali-sation)

amphiblt = amphibolite;
(after Cerny, 1982b)

TABLE 2.2 Elements Recovered From Pegmatites and Minerals Hosting Them

ELEMENT	HOST MINERAL(S)	STATE OF ELEMENT OCCURRENCE IN MINERALS
Boron (B)	tourmaline feldspar	$(BO_3)^{3-}$ $(BO_4)^{5-}$
Lithium (Li)	petalite, spodumene eucryptite, lepidolite amblygonite, holmquistite	$(LiO_6)^{11-}$
Beryllium (Be)	beryl, phenacite chrysoberyl, bertrandite	$(BeO_4)^{6-}$
Tantalum(Ta)-Niobium(Nb)	tantalite-columbite, pyrochlore-microlite,	$(Ta,Nb)O_4)^{3-}$
Tin (Sn)	cassiterite	SnO_4
Uranium(U)-Thorium(Th)	uranothorite	$(U,Th)O_2^{2+}$
Rare-earth elements (REE)	monazite, gadolinite allanite, euxinite, samarskite, xenotime, fergusonite, bastnaesite	REE_2O_3

(after Sheeran, 1984)

2.5 INTERNAL RELATIONSHIPS

Pegmatites can be subdivided on the basis of their internal relationships into simple and complex types. Simple pegmatites, which are the more common, have a simple mineral assemblage consisting of coarse-grained quartz, feldspar and mica.

Complex pegmatites frequently contain a wide variety of minor and trace elements. Compositional complexity gives rise to a diverse mineralogy, which is commonly manifested in the development of numerous zones and replacement bodies. In complex pegmatites, three main types of internal bodies are recognised and are referred to as zones, replacement bodies and fracture fillings.

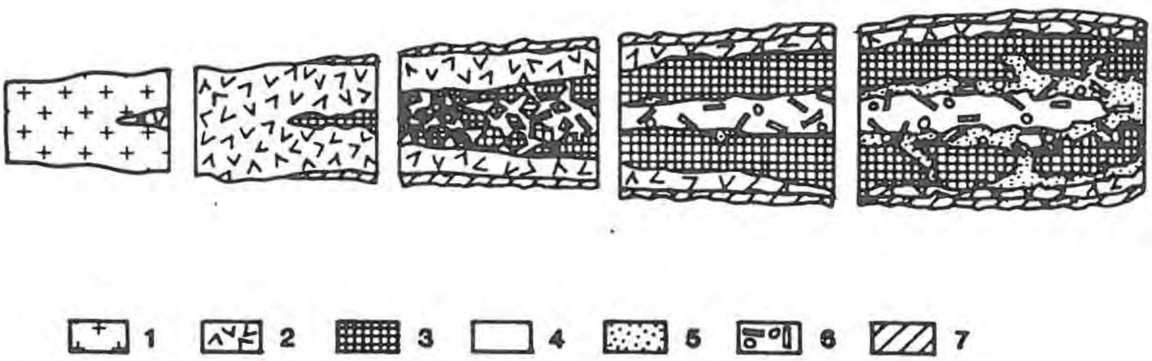

FIGURE 2.1 Idealised zones found within a pegmatite body

1. pegmatite of granitic texture; 2. graphic pegmatite; 3. blocky K-feldspar, spodumene possibly with oligoclase; 4. blocky core-quartz 5. replacement zone (albitisation with muscovite, beryl, tantalite, etc.); 6. crystals of rare element accessory minerals; 7. muscovite + quartz + albite border fringe

Zones are concentric shells, complete or incomplete, which roughly follow the external shape of the pegmatite. Fracture fillings are cross-cutting, usually tabular veins, which infill fractures formed within consolidated pegmatites. Replacement bodies form at the expense of pre-existing bodies.

It is the zoned pegmatites which are of economic significance. Zones may be defined on the basis of composition, texture, and location within the pegmatite, and are arranged concentrically relative to an inner core. Four main zones are considered in complex pegmatites (Fig. 2.1). The border zone is the outermost. It is commonly fine-grained and often a few centimetres or less in thickness. Next is the wall zone which can be up to 3 metres in thickness and is coarse-grained. The core is the innermost zone or nucleus. Any zones between this and the wall zone are referred to as the intermediate zones. Five or more may be present. These are commonly coarse to very coarse-grained, and some consist of crystals many feet in length. They are relatively more variable in thickness, and may be discontinuous or asymmetrically developed. The general mineralogical sequence from the border inwards is outlined below.

1. Plagioclase-quartz-muscovite .... Border zone
2. Plagioclase-quartz .... Wall zone

3a Quartz-perthite-plagioclase + muscovite, biotite	Inter-mediate zones
b Perthite-quartz	
c Perthite-quartz-plagioclase-amblygonite-spodumene	
d Plagioclase-quartz-spodumene	
e Quartz-spodumene	
f Lepidolite-quartz-plagioclase	
g Quartz-microcline	
h Microcline-plagioclase-lithiamica-quartz	

11. Quartz .... Core Zone

(after Cameron, 1959)

Beryl occurs in all the sub-zones. A green variety is commonly found in zones 1 to 4, while a blue, golden, white or pink variety may be found in sub-zones 5 to 10. The changes correspond to variations in alkali content.

Vertical variation is also known to occur (Sodolov, 1964). Pegmatite bodies produced by fractionation of a magma injected in one pulse show microcline assemblages higher up and albitic assemblages lower down. If injected in several pulses, and from a continually fractionating magma, the more primitive assemblages lie at a higher level (Jahns, 1982; Norton, 1983).

2.6 EXTERNAL RELATIONSHIPS

Pegmatites commonly occur in association with granites. Mica-pegmatites are found in regions of high-pressure metamorphism of Barrovian type, and are confined to the high-temperature end of the series in the sillimanite-bearing schists. On the other hand, feldspathic pegmatites are confined to the kyanite-bearing schists (Fig. 2.2). More evolved Be-bearing types occur at slightly lower temperatures, below the staurolite isograd (Cerny, 1982a).

Rare-element pegmatites are generally found in low-pressure metamorphic terranes of the Abukuma-type, where they are restricted to the andalusite-cordierite-muscovite subfacies. An example of this occurs in the Damara Orogenic Belt in Namibia, where they are confined to the low-pressure/high-temperature region (Richards, 1986).

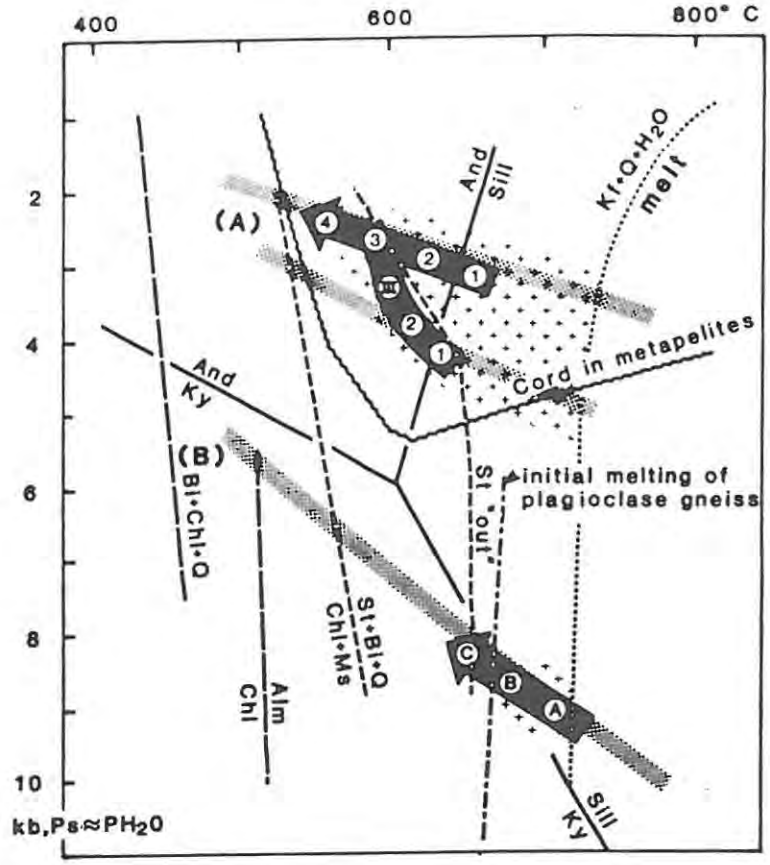

FIGURE 2.2 *Location of pegmatites (black arrows) in metamorphic facies series {stippled bands: (A) - Abukuma low-P facies series; (B) - Barrovian intermediate-P series}. Crosses - pegmatite-generating granites. Intermediate-level rare element pegmatite formations: 1. barren; 2. beryl-bearing; 3. spodumene-bearing; 4. petalite, Ta, Cs-bearing. Deep-level mica bearing pegmatite formation: A - barren, ceramic, B-muscovite-bearing, C - beryl-bearing (after Cerny, 1982c)*

2.7 REGIONAL ZONATION

Zonation around a pluton is concentric with the more evolved pegmatites (i. e. those more enriched in volatiles and rare-elements) lying farther away from the pluton, while granitic ones lie nearer (Fig. 2.3). The latter are very similar to internal pegmatites which lie inside the associated granite. The internal and marginal pegmatites often contain quartz, mica, muscovite, beryl, garnet and tourmaline. External pegmatites may occur up to 15 km away, and are enriched in Li, Ta, Nb, Be and Bi. They may grade into scheelite-bearing quartz veins. The zonation occurs vertically as well as horizontally.

The top of a pegmatite field is often 500-600 metres above the source granite, commonly also the limit of contact metamorphism of the country rocks. The lower boundary generally lies within the parent granite. Significant bodies of pegmatite are rarely developed in the zone of contact migmatisation.

Regional zonation is attributed to thermal gradients after magma has been injected in one pulse (Fersman in Cerny, 1982). It has also been attributed to periodic tapping of a continuously evolving magma from its chamber (Heinrich, 1953), with later magmas being able to travel farther due to the higher content of volatiles (Varlamoff, 1972). The processes are also considered to compliment each other (Cerny, 1982c). Stratification in the chamber is thought to result from a combination of gravity and movements of volatiles.

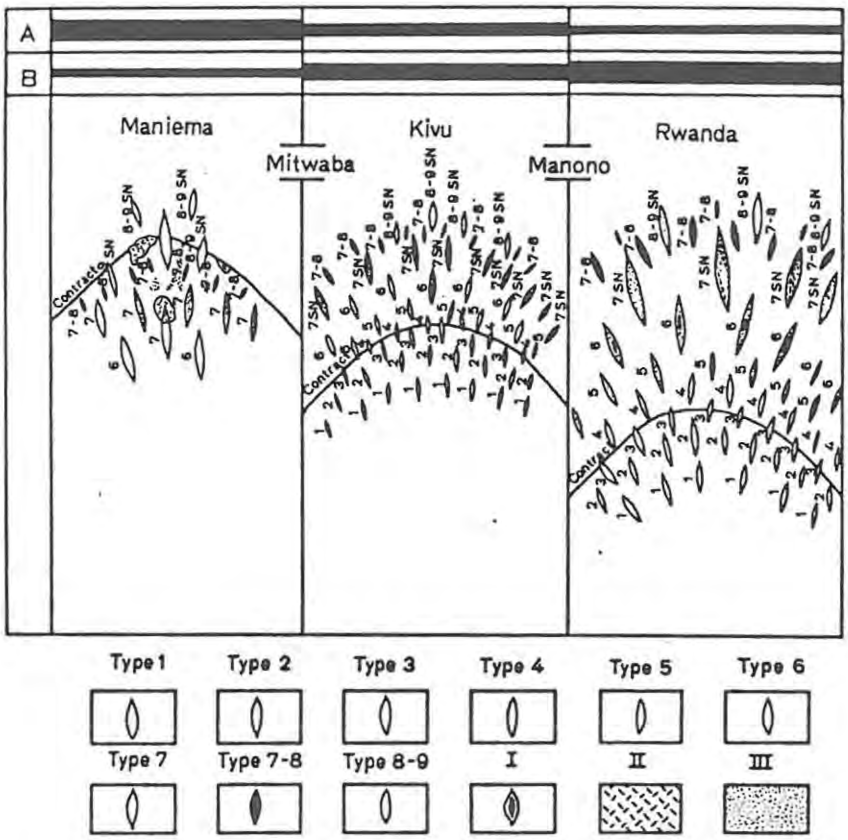

FIGURE 2.3 *Example of regional zonation as modelled in rare-metal pegmatite deposits of Rwanda. Type 1: Biotite pegmatites; Type 2: Biotite and tourmaline pegmatites with strong development of graphic texture; Type 3: Muscovite, tourmaline and pegmatites with graphic feldspars; Type 4: Muscovite and tourmaline pegmatites; Type 5: Muscovite pegmatites; Type 6 Zoned pegmatite, frequently with quartz core, gigantic crystals of amblygonite and spodumene, pockets of beryl, low contents of cassiterite, columbo-tantalite and microlite; Type 7: Albitised pegmatites with spodumene muscovite, lepidolite, small prisms of beryl, cassiterite and columbo-tantalite; Type 7-8: Quartz veins with microcline, frequently albitised and with muscovite and cassiterite; Type 8-9: Quartz veins with scheelite and ferberite, possibly with cassiterite passing in depth to quartz veins with muscovite, cassiterite and minor wolframite–I - pegmatites with some internal zoning and quartz core or pockets; II - parts affected by greissenisation. A - Intensity of greissenisation; B - Intensity of albitisation (after Varlamoff, 1972)*

2.8 RELATIONSHIPS TO GRANITES

High level miarolytic pegmatites are found within and along contacts of epizonal granites. The granites are commonly sub-alkaline to alkaline, porphyritic, and with a poikiloblastic or granophyric groundmass. Microcline predominates over plagioclase. Crystals have highly ordered cores, but become disordered towards the margins. Accessory minerals include fluorite, ilmenite, magnetite, zircon, apatite, cassiterite, monazite, thorite, pyrite, molybdenite and garnet. The granites are undersaturated with respect to water, which only increases to high levels during late-stage differention.

Rare-element pegmatites of intermediate levels are mainly exogranitic and a genetic link is difficult to certify. Where a genetic relationship can be demonstrated, the granites are generally late- to post-tectonic and post-date the peak of regional metamorphism. They are commonly intruded along major fracture zones as elongate bodies, often with an apical capping of pegmatitic granite. Texturally, they may be porphyritic to equigranular and may consist of leucocratic, biotite-, two-mica-, or muscovite-bearing varieties, and a plagioclase composition between An_2 and An_{15}. The K-feldspar is often highly-ordered microcline, which indicates equilibration at depth. The modal composition is granitic. There is minor enrichment of K and depletion of Mg and Ca. Minor elements Ti, Zr and Ba are commonly depleted, whereas Li, Be, Sn, Rb, and Cs are variably enriched (Cerny, 1982b).

Phanerozoic rare-element pegmatites are generally distributed around individual plutons. In Archaean and Proterozoic metamorphic belts, pegmatites of this type may occur along major fracture systems, considerably removed from any parent granite. The pegmatites at Bikita in Zimbabwe are of this type.

Mica-pegmatites from deep levels generally occur as concordant bodies in areas of sillimanite-bearing paragneiss, and their relationship to granite intrusions is uncertain. Where a relationship has been demonstrated, the granites are coarse-grained. They occur under cupolas of migmatitic domes in Barrovian metamorphic terranes, and generally contain two micas, with a plagioclase composition lying between An_{20} and An_{30}. Accessory minerals include garnet, tourmaline, kyanite, zircon, monazite and apatite. The composition varies from trondjemitic to that of true granites. They are enriched in Ca, Ba and Sr, while alkali enrichment is minor.

Pegmatites of maximum depth grade into leucosomes of migmatites from which they are derived (Buddington, 1959).

The genetic relationships between pegmatites and their source rocks may be summarised as follows:

1. partial melting of migmatite terranes of granulite facies metamorphism, yielding REE-, Th-, U-, Nb-, Ti-, Zr-bearing pegmatites,

2. partial melting and restricted calc-alkaline granitoid fractionation in Barrovian facies series producing mica-bearing pegmatites,

3. extensive calc-alkaline granite fractionation in the Abukuma facies series, generating rare-element pegmatites, extended to the miarolytic type of rare-element pegmatites at the subvolcanic levels of orogenic belts,

4. mostly sub-alkaline to alkaline granite fractionation in anorogenic structures, which yields Y-, Nb-, Be- and F-enriched rare-element pegmatites, and, at a shallow levels, the fluorite-, rock-crystal- and gem-bearing miarolytic pegmatites. This last class is not considered in this dissertation.

2.9 FIELD AND REGIONAL ASSOCIATIONS

A system in which pegmatites may be grouped spatially recognises vein zones as the smallest unit in which a group of pegmatites is associated with a single local structure. The vein zones constitute, in turn, pegmatite fields, belts, provinces and megabelts in that order (Table 2.3).

2.10 TECTONO-MAGMATIC SETTINGS

Migmatitic pegmatites are produced during any orogenic event by melting of local protoliths. Alkaline to sub-alkaline granites are believed to have been derived by melting of the lower crust. If fused material is generally granulitic and depleted in granitic components, pegmatites rich in K, Nb, Y, REE, F and Zr are produced. Fusion of granitic material will yield Li, Sn, Be, W and Mo. The granites will be intruded at a high level if they contain substantial amounts of CO_2, and F, and low H_2O.

Rare-element granites are considered to have originated from the lower crust, or a mixture of this and parts of the upper mantle. Evolution and emplacement occurred at depths varying between 8 and 15 km (Section 2.3; Isihara, 1981). Fractionation at source is thought to be insignificant, otherwise the rare-elements would have been lost at an early stage. Some of the materials that are rich in rare-elements and may have formed source rocks are considered to have been derived from alkaline magmatic rocks produced during the Proterozoic, when this form of magmatism was very common. Reactivation of such materials may be responsible for S-type granite-magma generation, as is thought of the Damara Orogen (Miller, 1983; Richards, 1986).

TABLE 2.3 Field Associations of Pegmatites

Vein system	Qualitative characteristics of ranking (systems)	Quantitative characteristics of ranking (number of veins)
Vein zone	Group of pegmatite bodies, localized within a single local structure	Tens — few hundreds
Pegmatite field	System of vein zones within large local or small regional structure	Hundreds up to 1-2 thousand
Pegmatite belt	Linear-elongated system of pegmatite fields, controlled by a regional structure	Thousands
Pegmatite province	System of pegmatite belts, confined to major regional structure (fold region, ancient shield)	Few tens of thousands
Megabelt*	Linear-elongated system of pegmatite belts and provinces, controlled by a global structure	Tens of thousands

2.11 THE SOURCE OF PEGMATITIC MATERIALS

The common association of rare-element pegmatites with granites, and in particular the leucocratic types, is suggestive derivation from them. This view is reinforced where the rare-element pegmatites are seen to grade into pegmatitic granites, as, for instance, in the Black Hills of the USA (Cerny, 1982c). Further evidence takes the form of an antipathetic relationship in the distribution of incompatible elements between granites and associated pegmatites, as illustrated by pegmatites of the Winnipeg River district of Canada (Cerny, 1982a and b).

I-type source granites are produced from partial melting of underthrust oceanic crust in subduction zones. S-type granites originate from higher crustal levels, in continent-continent collision zones. Melting occurs due to raised geothermal gradients resulting from crustal thickening or an early phase of I-type granite emplacement, and is possible within the crust if water is present. For instance, a water saturated amphibolitic to granitic source rock at 5 kb will melt between 600 and 700°C. However, saturation is rare, as pore waters do not occur in sufficient quantities. The breakdown of hydrous phases, and the concomitant release of water, is therefore considered to be the important source (Strong, 1980). Breakdown of hornblende can occur at extreme depth, such as at the base of the crust, in the subcontinental lithosphere and above subduction zones, to produce tonalitic magmas. Due to their high temperature, these would ascend to regions near the surface before the hornblende solidus was reached. Rapid ascent is important as cooling would otherwise immobilise the melt. Breakdown of biotite would produce granodiorites, while that of muscovite would produce granites at higher levels. A complete succession is likely to exist between tonalitic to leucogranitic types (Fig. 2.4).

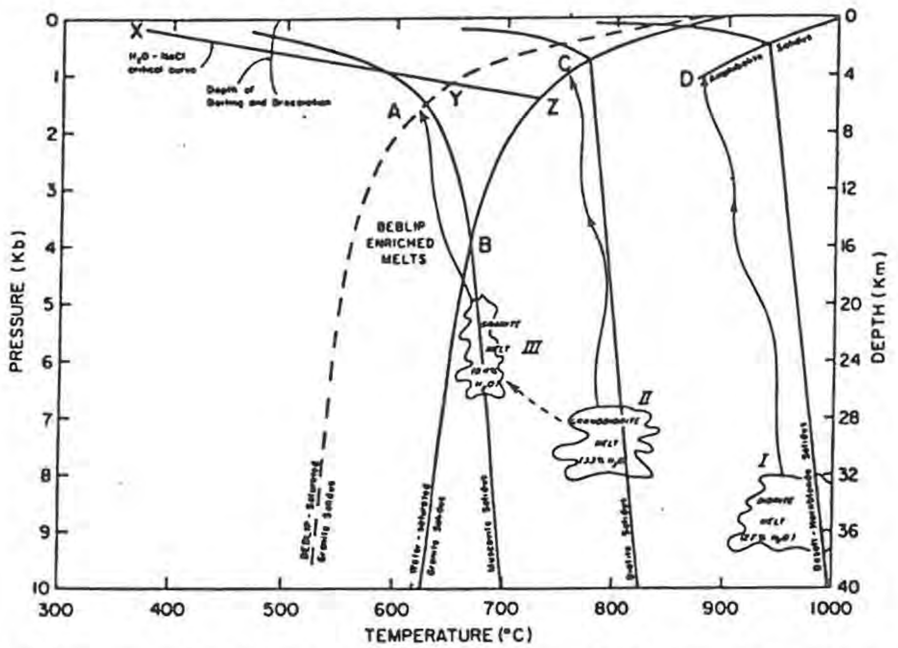

FIGURE 2.4 *Approximate conditions of granitoid generation. Solid arrows indicate paths of ascent. Broken arrows indicate differentiation of one magma from another. (after Strong, 1981).*

Breakdown of biotite will also release rare-elements, but it is associated with large volumes of melt, which would cause their dispersion. Under favourable conditions, the granitoids will fractionate to form granites or leucogranites (Strong, 1981).

2.12 EVOLUTION OF PEGMATITE MAGMA

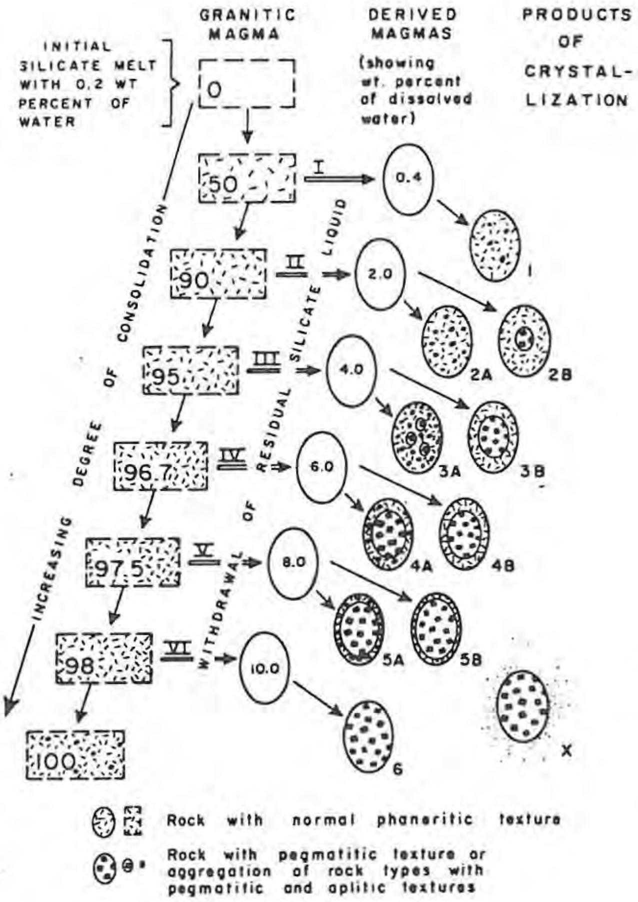

FIGURE 2.5 *Diagram to illustrate diagrammatically the crystallisation of granitic magma and the withdrawal of residual silicate liquid at six different alternative stages to form bodies of granite and pegmatite with contrasting lithologies. Assumed initial water content in magma is 0.2 weight %;. Crystallisation assumed to occur under constant pressure. Figures in boxes on left are % of original silicate used up, and those in adjacent ellipses indicate contents of the stages. Consolidation of these bodies illustrated in right hand column, and each pair of bodies is designated A and B representing contrasting types of segregation. Body X surrounded by zone of metasomatically altered country-rock (after Jahns and Burnham, 1969).*

Concentration of water in an evolving pegmatitic magna occurs due to fractionation of anhydrous phases. Granodioritic or biotite-granite melts will crystallise plagioclase. At constant pressure, water will concentrate in the melt. Subsequent crystallisation of biotite and, later, muscovite, will be inadequate to remove all the water. Extreme concentration occurs if the melt, already enriched in water, is removed from its crystallised fraction (Fig. 2.2). This may occur due to tectonic movements (Nedumov, 1964).

Concentration of rare-elements is likely to be due to this fractionation process, aided by convective concentration at the top of the chamber (Burnham, 1969; Willie, 1979; Manning, 1982; Stewart, 1978; Hess, 1980). Prolonged evolution of the magmatic stage allows extreme concentration of the rare-elements into the aqueous melt.

2.13 CRYSTALLISATION AND CONSOLIDATION

Crystallisation is triggered by separation of an aqueous phase within the melt. This can occur due to the separation of an anhydrous phase resulting from decreasing temperature. Alternatively, it may occur due to a sudden drop in the pressure, caused by filling of a fracture just formed or entered by the magma. Initially, the liquid phase will occur as fine bubbles disseminated throughout the melt. At this stage, K, Li, Sn, Ta, Nb, W, P and F will partition into the bubbles. With continued exsolution, the bubbles will grow and move upwards where they will congregate to form a continuous aqueous phase. Coarse grains of such minerals as mica, spodumene, or petalite will then crystallise. Thus the fluid phase is considered to be an important factor which facilitates melting. Later, it becomes the medium of concentration and transfer, controlling crystallisation of the normal products of granitic pegmatites (Jahns and Burnham, 1969).

Internal zonation is believed to be controlled by the amount of aqueous fluid present, the amount of water loss to the surrounding rocks, the continuity of the aqueous phase, the magnitude of the temperature gradient and the degree of upward concentration of the aqueous phase. Concentric zonation occurs if crystallisation takes place inwards. If the drop in pressure is sudden, boiling results, so that the aqueous phase spreads into the country rock while aplite is formed within the walls. If the change in pressure is gentle, the separation of the aqueous phase is also gradual. K and Li are then carried into the country rock by the separated phase and form K-feldspar and holmquistite in alteration assemblages. This produces border and wall zones enriched in Na with Al, and leads to the crystallisation of plagioclase, muscovite and quartz. Removal of Al allows development of plagioclase-quartz assemblages in the inner wall zone (Norton, 1983).

Once the border and wall zones have developed, the trapped pegmatite melt behaves like a closed system, permitting an increase in the amount of H_2O present as more anhydrous phases crystallise out. Asymmetric zonation in the intermediate zone occurs if liquids migrate upwards. Upward movement of K produces coarse-grained microcline-microperthite assemblages in the upper zones, followed by quartz-K-feldspar assemblages, with spodumene and petalite if the Li content is high enough.

In horizontal or sub-horizontal pegmatites, extreme aqueous segregation may lead to concentration of K-feldspar and spodumene in the hanging-wall. At the same time, an albitic aplite will develop in the floor (Jahns, 1982). The presence of significant quantities of F causes extreme albite enrichment in the residual melt. This may form albitic aplite with pockets of coarse-grained K-feldspar after boiling. The quartz core probably forms by precipitation of SiO_2 from a residual fluid phase whose liquidus and solidus have been depressed by the presence of the minor amounts of K, Na, Al and Li (Norton, 1983). The latter crystallise on the margins of the core afterwards.

2.14 REPLACEMENT REACTIONS

Replacement reactions are likely to occur in semi- to fully consolidated pegmatite by the action of super-critical residual fluids enriched in K, F, P and REE. As the liquid moves through pegmatite, K replaces Na and Ca, while previously dispersed rare-elements may be leached. As temperature drops, leached Na will react with K-feldspar to cause albitisation. This will be accompanied by Sn and Ta-Nb mineralisation (Pollard, 1983). Finally, K-feldspar is altered to muscovite and quartz during greissenisation. This is a process of H^+-metasomatisation which tends to be fracture-controlled, and is commonest in the apical parts of the granite plutons and related pegmatites. It is associated with significant concentrations of Sn, Nb-Ta and W.

2.15 GLOBAL AND CHRONOSTRATIGRAPHIC DISTRIBUTION

Mineralised pegmatites occur on all the continents and have ages ranging from the Early Archaean to the Tertiary (Fig. 2.6 and 2.7). Archaean pegmatites were intruded during two main phases which coincide with the two major episodes of granite-greenstone terrane formation. The first and oldest formed between 3900 and 3300 Ma. Examples are found in the Godthaab Complex of Greenland, the Uivak suites of Labrador, and the Saamian of north-west USSR (Cerny, 1982b). Most of the pegmatites are barren, as they lie within the predominantly tonalitic gneisses which are generally too basic. The other class includes the first important and extensive pegmatites, and has ages ranging from 2800 to 2600 Ma. Again, the pegmatites are found within cores of Precambrian shields. The mineralised pegmatites are restricted to two main settings.

They may occupy shear zones which cut across greenstone belts. Alternatively, they may be present as advanced stage differentiates of late-tectonic granites normally intruded into the margins of the belt. Examples include pegmatites of the Superior and Slave Provinces in Canada (Ayres and Cerny, 1982), the Kimberley, Yilgarn and Pilbara blocks of Australia, and Kola in the Baltic Shield (Norway). In central and southern Africa, these are represented by pegmatites in the Tanzanian, Zimbabwean, Angola-Kasai and Kaapvaal cratons (Anhaeusser and Wilson, 1981). Mineralised Archaean pegmatites are RE-bearing in general, indicating a high-temperature/low-pressure environment of formation. This has been taken as evidence for a thinner crust and a higher geothermal gradient during this early part of the earth's evolution (Windley, 1987).

Mineralised Proterozoic pegmatites are common in the more extensive mobile belts. This class includes Hudsonian (1800-1600 Ma) pegmatites of the Churchill Province in Canada, parts of Sweden, in the Ukraine to Inner Mongolia, and those of the Grenville Belt (1200-800 Ma) in the northern hemisphere. In the southern hemisphere the younger group are represented by occurrences in the Rondonia Belt of South America, Kibaran-Namaqua-Irumide Belt in South Africa, the West Congo Geosyncline of central and southern Africa (Anhaeusser and Wilson, 1981) and the Musgrave Block of Western Australia (Cerny, 1982b). Pegmatites of this era are worked predominantly for their mica and feldspar. These minerals form at great depth, and this is taken as evidence of the presence of a thick crust having developed by this time. However, mineral deposits of tantalite, wolframite, fluorspar and beryl are known in the Namaqualand Belt of South Africa (Hugo, 1986) and related belts farther north on the same continent (Varlamoff, 1972; Pohl and Gunther, 1991). Tin, lithium and tungsten are also produced in the western part of the Zambezi Belt (Ewart, 1960). The co-occurrence of mineral assemblages is a common feature of the pegmatites due mainly to the changing environment in which they are emplaced.

The next major period of formation of mineralised pegmatite occurred between the late Proterozoic to early Palaeozoic in a thermo-tectonic event which reached its peak at about 500 Ma, after a phase of stability which lasted, in general, from 900 to 700 Ma (Hurley, 1974; Sawkins, 1976). This was widespread and occurred in a system of linear belts which cut across the Pangea Supercontinent (Fig. 2.5). The pegmatites are often associated with late- to post-tectonic granites. Examples of pegmatites formed include fields of the pre-drift circum-Atlantic belts such as those of Nigeria, Morocco, Namibia, Mali, Brazil and the Appalachians which formed between 800 and 350 Ma, thereby straddling the Precambrian-Phanerozoic boundary (Cerny, 1982b). Belts of the same phase occur in Madagascar, Kerala (India) and probably Mozambique (Soman et al., 1983).

FIGURE 2.6 *The distribution of mineralisation zones associated with granitoid magmatism and rare-metal magmatism in space and time (after Rundkrust, 1977)*

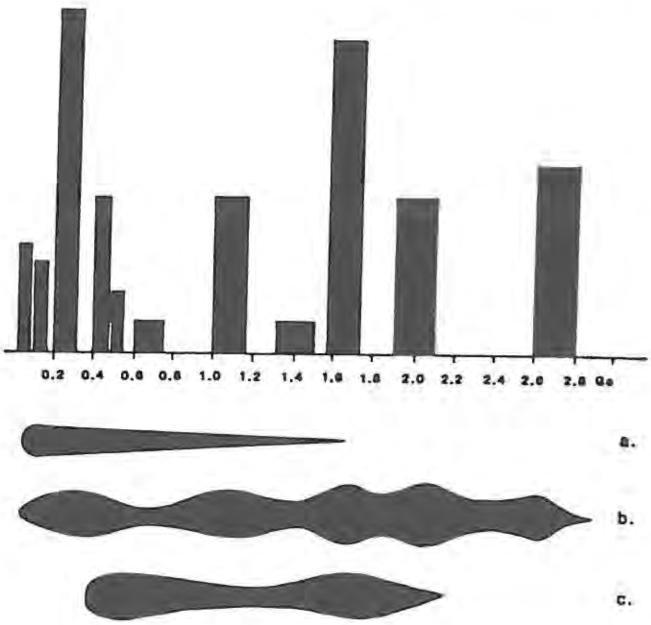

FIGURE 2.7 Frequency of pegmatite generation through geological time. *The histogram indicates the number of pegmatite provinces formed at a given period. The sausagegram indicates the formation of (a) miarolytic, (b) rare-element and (c) mica-bearing pegmatites (after Cerny, 1982b).*

Belts of similar age may be hidden under the Kalahari sand and Karoo cover in Angola. In western Europe, Caledonian belts also belong to this period, but are unmineralised with respect to pegmatitic ore-deposits. The exception is the Rosses Pluton in Donegal, Ireland, which is RE-bearing, and has uraniferous alaskitic pegmatites. Belts of this phase falling within Africa belong to the Pan-African event. For central and southern Africa, the belts include the Katangan, Mozambiquan, and the Damara. Pegmatites of this phase tend to be rich in Sn, beryl, pollucite, lepidolite and petalite. They are worked also for mica and feldspar.

Hercynian pegmatites occur in belts which separated Gondwanaland and Laurasia, and ran across the Urals, Khazakhstan and Turkestan. In addition to those of Caledonian age, new Hercynian pegmatites were occur in the Appalachian belts of eastern North America. Pegmatites of this age are rich in Sn, W, U, Mo, Be and Li (Hall, 1971; Hutchinson, 1984). RE are confined to greissen networks and quartz veins as seen at Ezgerbirge in Germany and Cornwall in Britain (Tischendorf, 1974). Appalachian belts are generally Li-Be types. The Hercynian pegmatites are, as a rule, miarolytic, probably because erosion has not exposed the deeper facies.

Cimmerian (Jurassic-Creteceous) pegmatites are found in the western American cordillera and in Transbaikalia, while the youngest pegmatite fields (85-20 Ma) lie in the Alpine-Himmalayan chain, and the Pamir Hindu-Kush Belt (Ono, 1970 in Cerny, 1982b). The South-East Asian belts are mainly Sn-, W-, and RE-bearing.

3. PREVIEW OF THE REGIONAL GEOLOGY OF CENTRAL AND SOUTHERN AFRICA

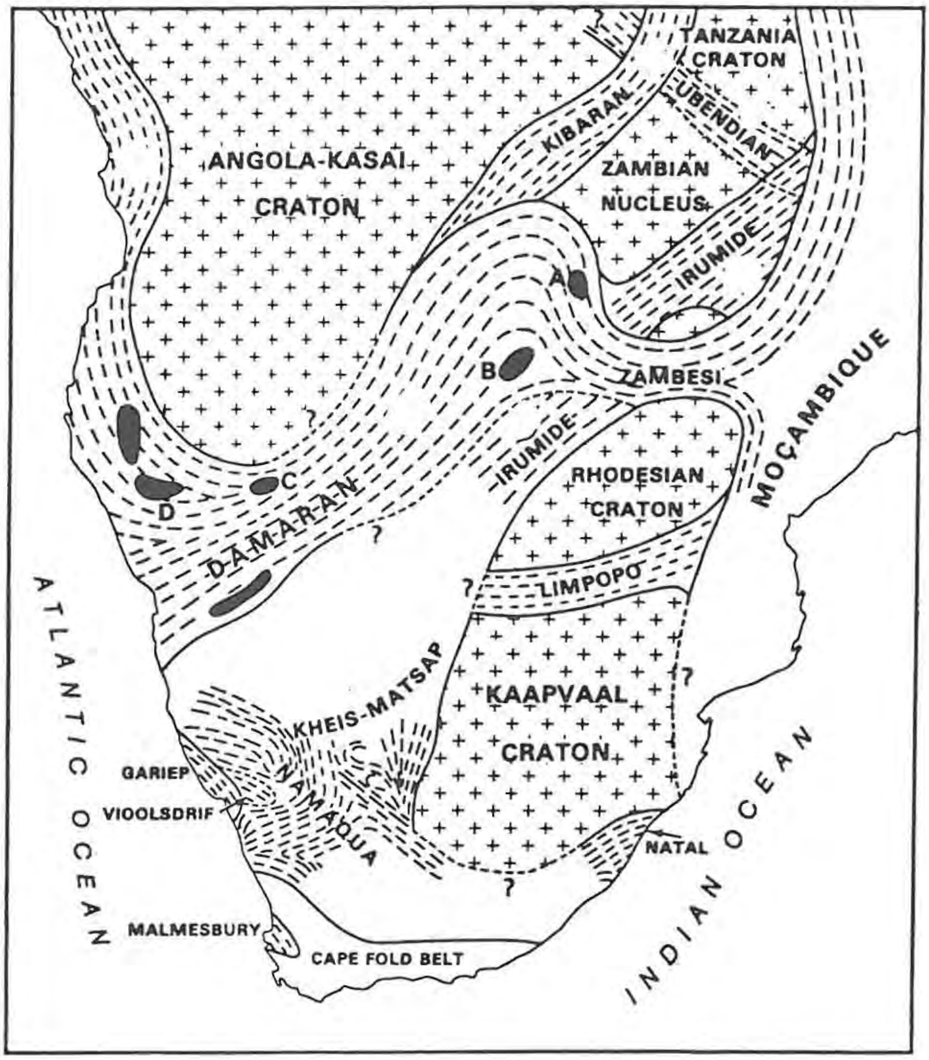

FIGURE 3.1 *Simplified tectonic map of central and southern Africa showing the distribution of cratons and mobile belts. The Limpopo Belt is approx. 2.6 Ga; Ubendian is 1.85 Ga; Kibaran, Irumide, Kheis-Matsap and Namaqua-Natal belts are 1100 Ma; Damara, Malmesbury and Zambezi (in part) are approx. 500 Ma. Solid black areas: B - Hook granite; C - Grootfontein Inlier; D - Kamanjab Inlier. Unlabelled lens south of Damara Belt - extensive inlier south of Windhoek (after Hunter, 1981b)*

On the regional scale, central and southern Africa can be divided into five major Archaean cratons surrounded by belts formed by the Limpopo, Ubendian, Namaqua-Irumide-Katangan-Lurian, and the Pan-African activity (Fig. 3.1).

A brief outline of the lithostratigraphy of the region is given by the various authors of a volume on the Precambrian geology of the southern hemisphere (Hunter, 1981) and a review of chronostratigraphic data on Africa (Cahen et al., 1966; Cahen et al., 1984). Models of regional tectonic settings are given in a review of global and African tectonics (Clifford and Gass, 1970). Briefly, the belts may be considered as the areas in which successive stages of crustal building were taking place, after the shields had formed (Fig. 3.2).

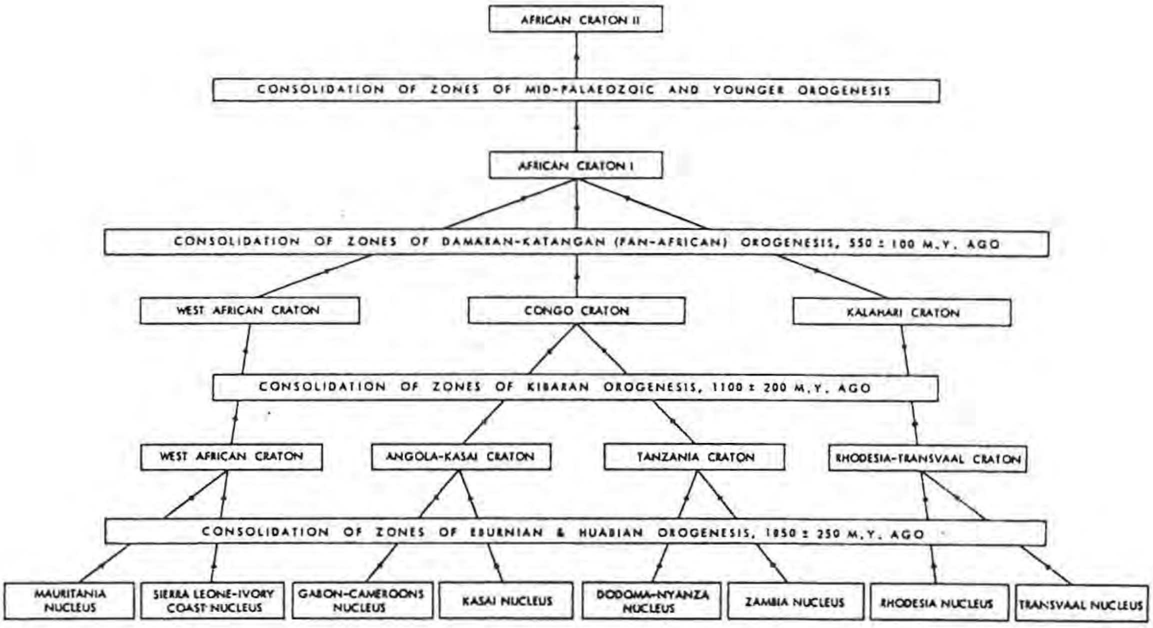

FIGURE 3.2 *Flow chart of the tectonic development of Africa (after Clifford, 1970).*

It is clear from the studies mentioned above that bodies of mineralised pegmatites occur in rocks of all ages that have undergone intense deformation and high-grade metamorphism, where they have been intruded by highly fractionated late- to post-tectonic alkali and peraluminous granites. This review therefore considers the settings of mineralised pegmatites in their geochronological order (Sections 4 to 8).

4. ARCHAEAN CRATONS AND ASSOCIATED MINERALISATION

The cratons, consisting of greenstone belts and granitic-gneiss terranes, underwent deformation and plutonism before the formation of the surrounding belts (Hunter, 1981b). The best studied are the Kaapvaal and Zimbabwe cratons, which are used to illustrate the major geological features of the Archaean cratons with respect to the stratigraphic, lithologic and tectonic setting of the mineralised pegmatites (Fig. 4.1).

4.1 THE KAAPVAAL CRATON

4.1.1 Granite-Gneiss Terrane

The granite-gneiss terrane constitutes 91% of the Kaapvaal Craton, and is considered to represent some of the earliest sialic crust (Hunter, 1981; Fig. 4.1). It is referred to as the Ancient Gneiss Complex, and is best exposed in central Swaziland, where it is composed of interlayered leucotonalite and amphibolite considered to be of igneous origin, and a metamorphite suite, comprised predominantly of paragneisses. The tonalitic component is considered to have been derived from 10-20% partial melting of tholeiitic basalt comparable to the type found in parts of greenstone belts. However, recent geochronologic information (Hunter, 1991) appears to indicate that the granitic-gneiss terrane was produced by major cycles of magmatism at 3.64 Ga, 3.45 Ga, 3.4 Ga and 3.2 Ga. The 3.2 Ga event seems to coincide with formation of a larger part of the Barberton Greenstone Belt. Another component considered to have an igneous origin is composed of homogeneous (unbanded) biotite- and hornblende-gneisses. The gneisses appear to be intrusive into the layered orthogneisses and metamorphite suite. They have metamorphic dates as early as 3395 Ma. The rocks equilibrated under high-temperature/low-pressure metamorphic conditions.

4.1.2 The Swaziland Tin Belt

The granitic gneisses of the Kaapvaal Craton do not contain significant pegmatite-hosted mineralisation (Fig. 4.2). It is believed that they did not evolve sufficiently to yield pegmatites. Mineralised pegmatites found in them can normally be traced back to distinct and younger alkali granites, which usually intrude the margins of the greenstone belts.

These occurrences are seen in the Swaziland Tin Belt (Fig. 4.3). The belt is roughly 38km by 16km and was discovered in 1892 (Davies, 1964). It has yielded low tonnages of Sn mined mainly from alluvial and eluvial sediments, with a total of 11455 tons (410 tons/year) being produced between 1892 and 1920, 150 tons/year between 1920 and 1942, and less than 10 tons/year between 1942 and 1955, after the loose deposits had been worked out. Although hosted in the granitic-gneiss terrane, the pegmatites appear to be associated genetically with a widespread and late-tectonic G4 group of granites. The granites consist of quartz (20-25%), K-feldspar (60-70%) and plagioclase (10-30%). Orthoclase dominates the marginal facies, while microcline dominates the inner facies. Where assimilation of supracrustal rocks has taken place, muscovite with occassional biotite and hornblende may occur. Zircon, apatite, spodumene, epidote, allanite, magnetite and ilmenite occur in accessory amounts (Davies, 1964).

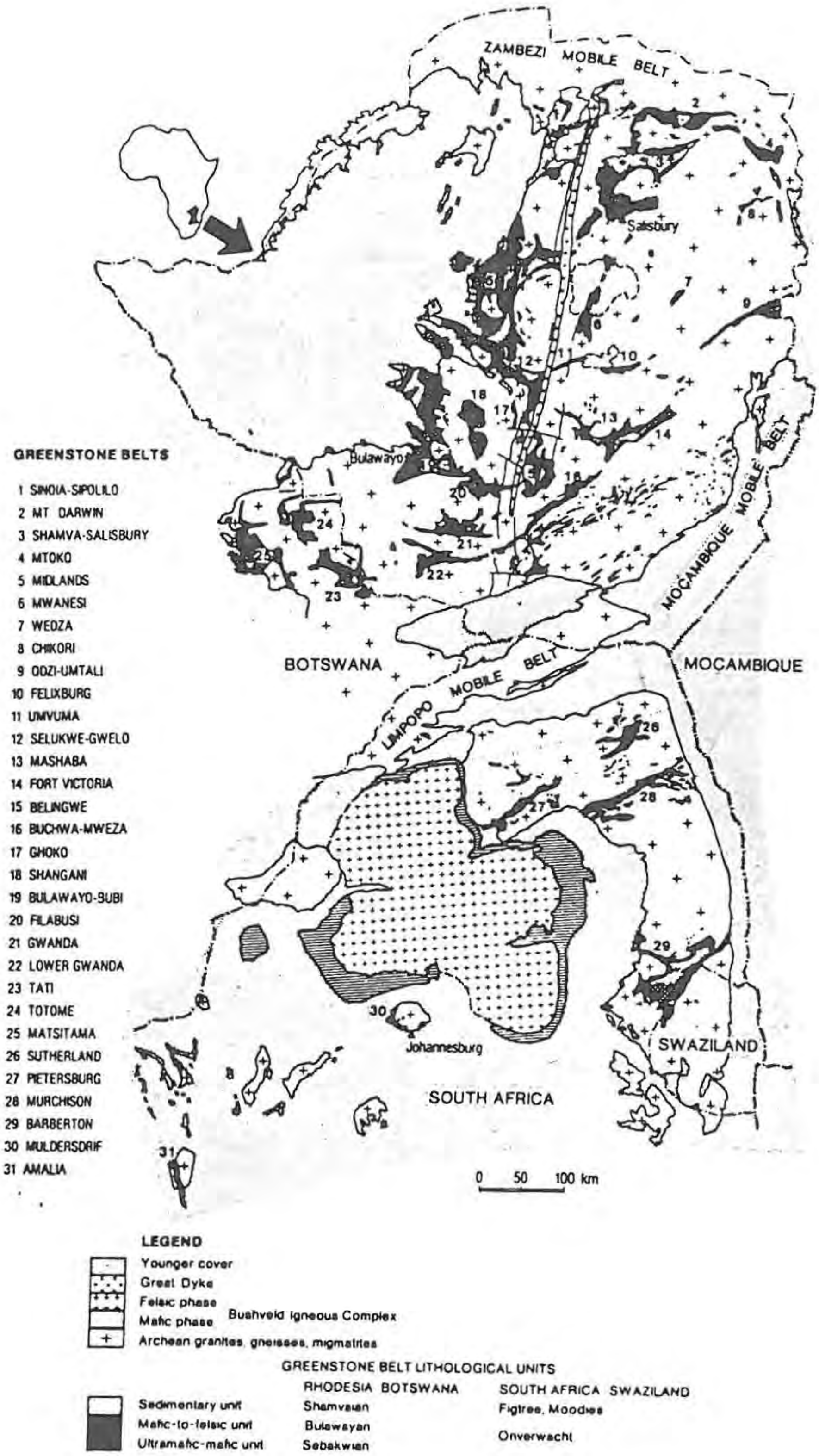

FIGURE 4.1 *Map showing the Kaapvaal and Zimbabwe or Rhodesian cratons of the eastern part of southern Africa, showing the main greenstone belts and other lithologies (after Anhaesser, 1976, 1986)*

These occurrences are seen in the Swaziland Tin Belt (Fig. 4.3). The belt is roughly 38km by 16km and was discovered in 1892 (Davies, 1964). It has yielded low tonnages of Sn mined mainly from alluvial and eluvial sediments, with a total of 11455 tons (410 tons/year) being produced between 1892 and 1920, 150 tons/year between 1920 and 1942, and less than 10 tons/year between 1942 and 1955, after the loose deposits had been worked out. Although hosted in the granitic-gneiss terrane, the pegmatites appear to be associated genetically with a widespread and late-tectonic G4 group of granites. The granites consist of quartz (20-25%), K-feldspar (60-70%) and plagioclase (10-30%). Orthoclase dominates the marginal facies, while microcline dominates the inner facies. Where assimilation of supracrustal rocks has taken place, muscovite with occassional biotite and hornblende may occur. Zircon, apatite, spodumene, epidote, allanite, magnetite and ilmenite occur in accessory amounts (Davies, 1964).

The mineralised pegmatites cut the G4 granite suite and do not show a preferred orientation. Zoning is poor and greissenisation absent. Mineralogically, the pegmatites are complex and consist of quartz, orthoclase, microcline, microperthite and cryptoperthite. Oligoclase may occur in association with magnetite, yttrotantalite, ilmenite, beryl, monazite, garnet, fluorite, biotite, muscovite, pyrite, pyrrhotite and chalcopyrite. Cassiterite is intergrown with the magnetite to the extent that the latter was used as an indicator of cassiterite mineralisation (Davies, 1964). Biotite occurs as small flakes where cassiterite mineralisation is present.

The base of mineralisation was not intersected at the maximum drill depth of 30 metres in the main body of pegmatite. Mineralised pegmatites had up to 18.5kg/m^3. However, one of the well-mineralised bodies yielded 50000 tons of ore with 0.37% Sn metal, so the figure might be confusing alluvial/eluvial deposits with those of the source pegmatite. Some low-grade mineralisation occurs in aplites. Where pegmatites cut basic or ultrabasic rocks, ruby corundum is formed.

Rare-element deposits hosted in pegmatites which occur in granite-gneisses are uncommon. An example is the Osis Lake pegmatitic granite of the Winnipeg River District of southeast Manitoba in Canada (Cerny and Brisbin, 1982). The pegmatites resemble those which intrude sediments of the Bird River Greenstone Belt. The granites from which they evolved are post-tectonic with respect to deformation of the greenstone belt. It would seem therefore that there are no pegmatites which evolved from granites of the granitic-gneiss terrane, and that mineralised pegmatites found are distal intrusions of the type intruding the greenstone margins (Section 4.1.4 to 4.1.8; Cerny and Brisbin, 1982).

4.1.3 Greenstone Terranes

The Kaapvaal Craton has five greenstone belts which occupy only 9% of the surface area (Fig. 4.1). The largest, least altered and sequentially-ordered is the Barberton Greenstone Belt, and is taken as the type belt (Hunter, 1981). It consists of interlayered volcanic, intrusive rocks and sedimentary rocks. The succession is referred to as the Swaziland Sequence or Supergroup (Anhaeusser and Wilson, 1981). At the base is the Onverwacht Group (3.3-3.7), an ultramafic mafic to intermediate volcanic sequence with pillows and komatiitic components. It is succeeded by deep-water sediments of the Fig-tree Group, which are in turn overlain by the shallow-water Moodies Group.

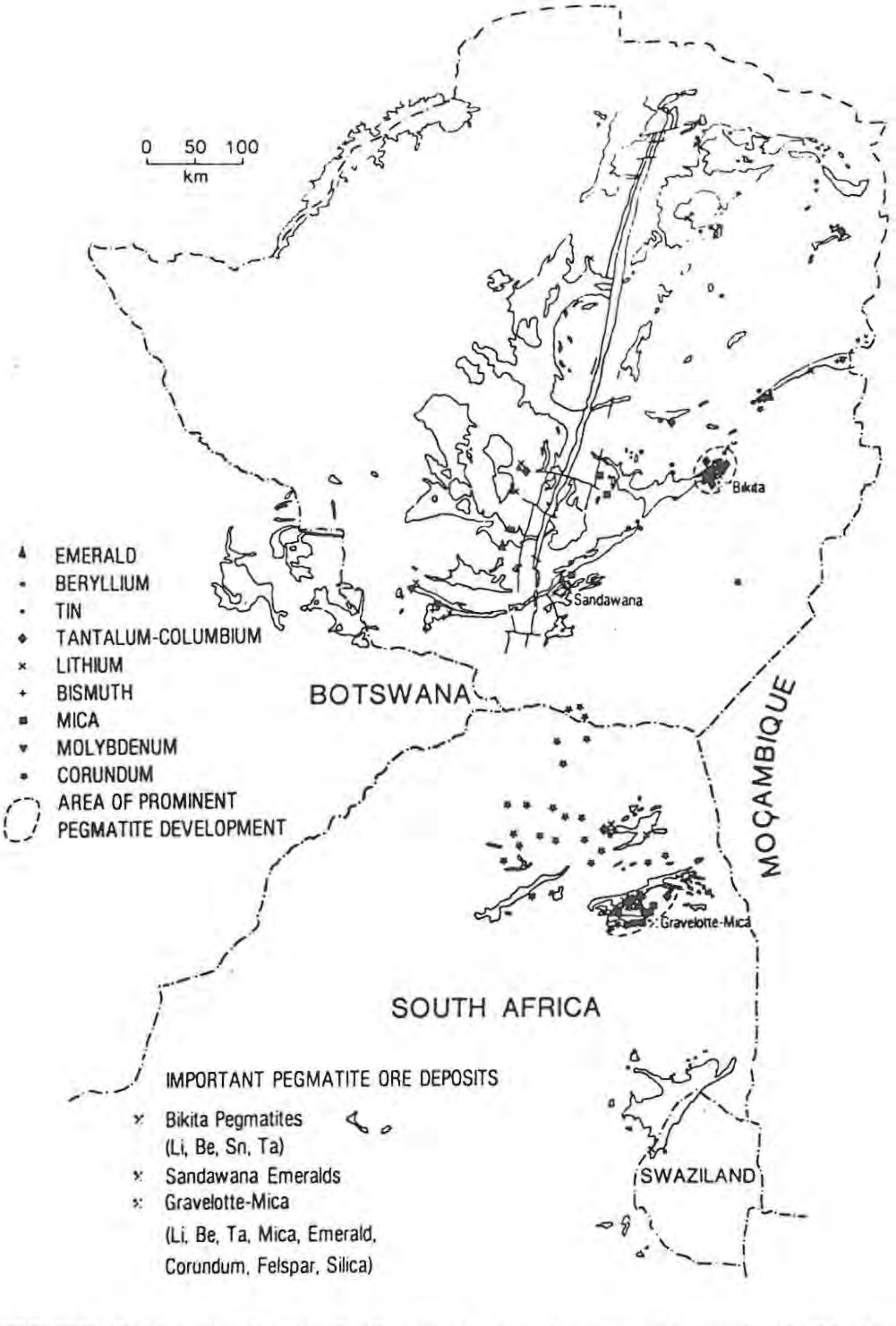

FIGURE 4.2 *Map showing the distribution of mineral occurrences associated with Archaean granites and pegmatites in Kaapvaal and Zimbabwe or Rhodesian craton (after Anhaeusser, 1976)*

Structurally, the greenstone belt forms an open, beltwide synformal structure on which are superimposed tight synclines, some of which are asymmetric, sheared or faulted. Regional metamorphism is retrogressive as it involves alteration to chlorite-epidote assemblages in the basic volcanic units of the sequence.

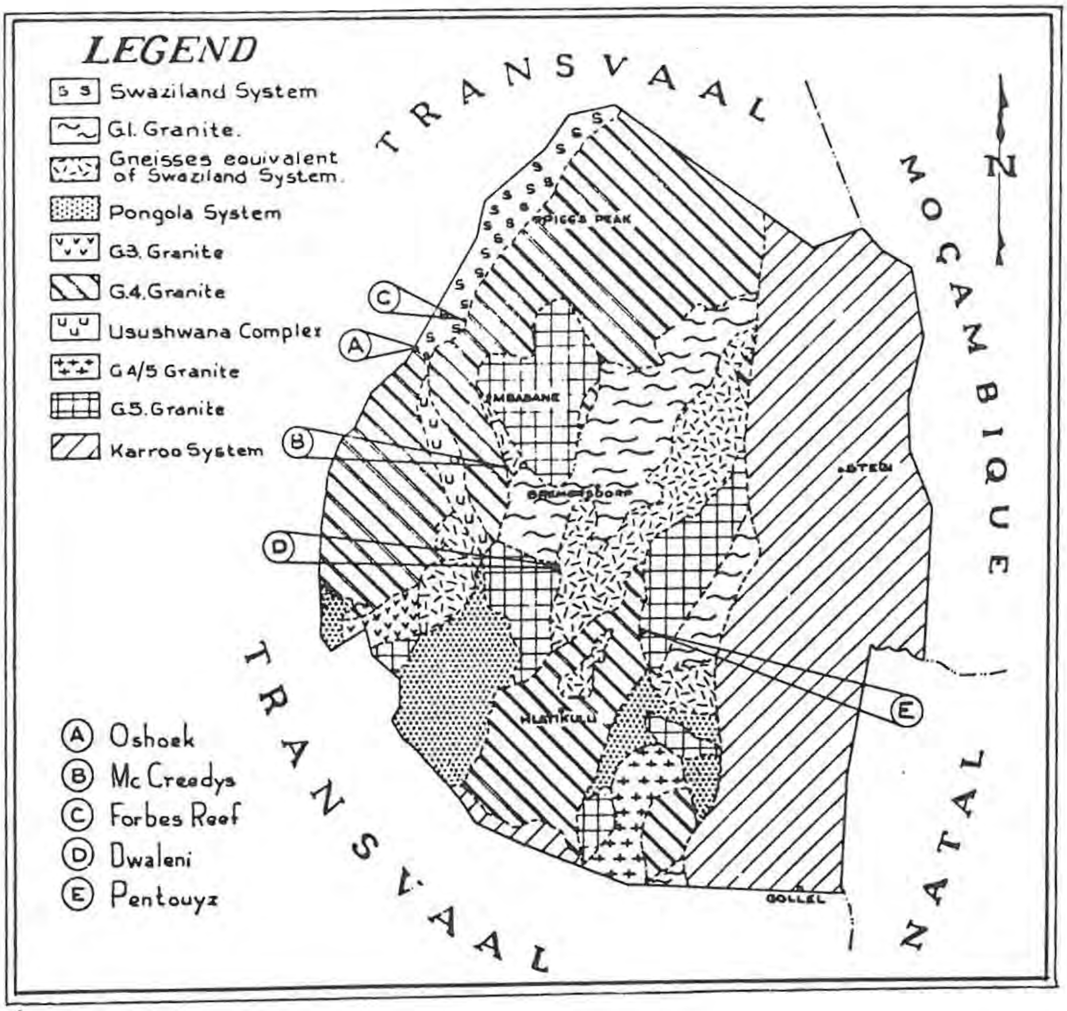

FIGURE 4.3 *Geology of Swaziand and location of tin deposits in the granitic-gneisses terrane (after Davies, 1964)*

4.1.4 The Sutherland and Murchison Deposits

Pegmatites with mineral deposits are related spatially, and probably genetically as well, to the 3.1 Ga marginal granites (Anhaeusser, 1976). Examples of mineralised pegmatites occur in the area between the Sutherland Greenstone Belt to the north and Mica to the south. Three regions are recognised: the southerly and better endowed belt which is subdivided into (i) the Selati Line and (ii) the Olifants River Mica Field, and (ii) a belt lying north of it and with its northern boundary reaching the Sutherland Greenstone Belt (Fig. 4.4)

The southern belt is related to the potassic, 2040 Ma Mashishimala Granite (Robb and Robb, 1986). This is coarse-grained and porphyritic. Its subcrop has been revised substantially recently, thereby extending the area of potential mineralisation. It is surrounded by the tonalites and trondjemitic gneisses with innumerable remnants of greenstones. The concentration of mineral deposits of corundum with beryl and columbite-tantalite occur where plutons related to the Mashishimala Granite invade an area with remnants of greenstone belts.

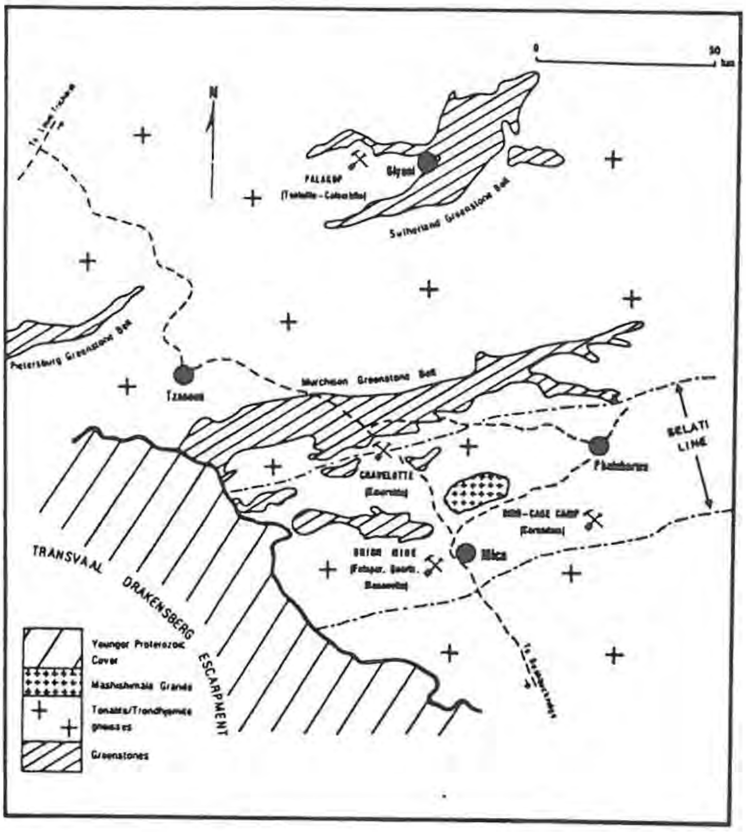

FIGURE 4.4 *Sketch of locations of some pegmatite deposits associated with Archaen greenstones and granites in the Kaapvaal craton in northeastern Transvaal, South Africa (after Robb and Robb, 1986)*

The pegmatite mineralogy is generally uniform. Their shapes, however, are variable, with podiform, lensoid, sill-like and irregular types. Their boundaries with the country rocks are sharp, while they are inhomogeneous internally. Zonation is variable, but typically consists of plagioclase and muscovite in the border zone, a wall zone with microcline and perthite with accessory quartz, muscovite, plagioclase, and a core zone of massive quartz. Three categories of mineralisation have been differentiated in the southern area: (i) the corundum type, (ii) the tantalum-columbite type and (iii) the emerald and beryl type. In the northern belt, quartz-mica-feldspar types of pegmatites are common (Robb and Robb, 1986).

4.1.5 Corundum Deposits

Deposits from the area made South Africa the leading producer of industrial corundum. Two classes of deposits are known and are referred to as: (i) plumasitic if corundum occurs with feldspar, (ii) marunditic if margarite occurs instead. A third class is associated with granitic gneisses and migmatites, occurs sporadically and forms unimportant deposits.

Significant deposits occur on the Selati Line and are marunditic. An example is the Bird-Cage Camp ore body (Fig. 4.5). The pegmatite is semi-circular and intrudes amphibolitic country-rocks. It varies from only a few centimetres to 9m in thickness, and shows poor or no zoning. Marundite occurs in the contact between the pegmatite and amphibolite (Fig. 4.5; Hall, 1964). Pegmatitic apophyses are completely altered to marundite to form reefs of coarse-grained ruby-coloured corundum with margarite plates. Some biotite and tourmaline may also be present. Along the boundaries of the marundite, the amphibolites have been altered to talcose material.

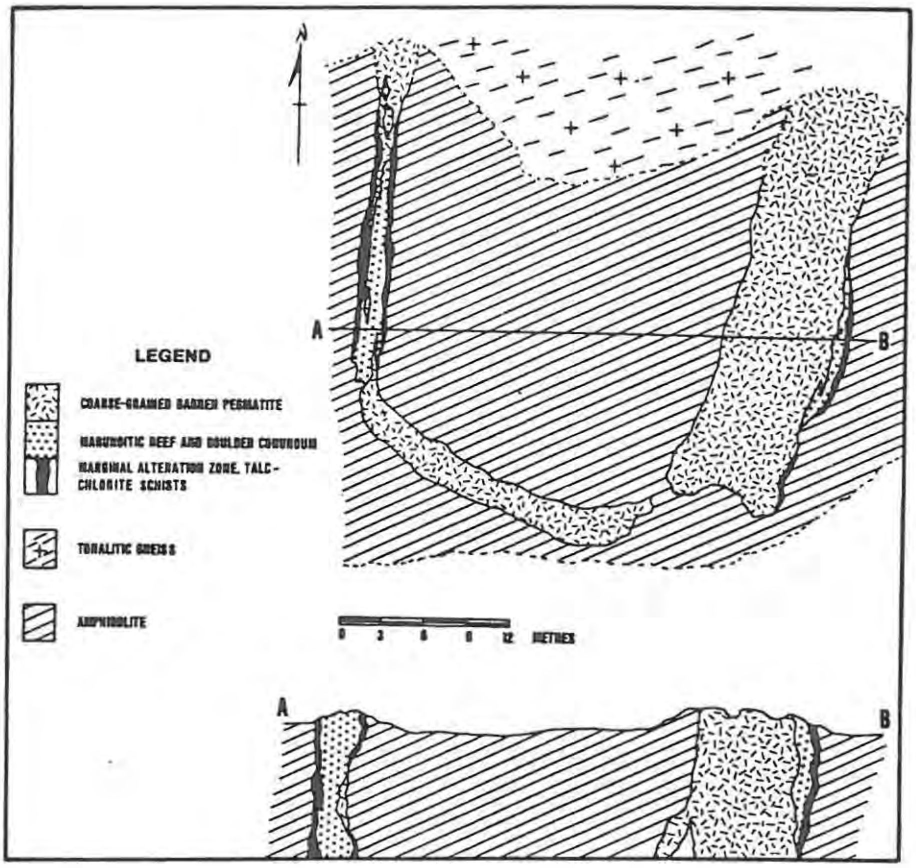

FIGURE 4.5 *Map and cross-section of the Bird-Cage Camp prospect showing the lithological setting and alteration of a marunditic and reef boulder corundum deposit from the location displayed in Fig. 4.4 (after Robb and Robb, 1986)*

Formation of the corundum has been attributed to relative loss of silica caused by intruding pegmatite (Du Toit in Robb and Robb, 1986). This was thought to have been caused also by addition of calcium, which led to formation of wollastonite. The mineral would, in turn, react with K-feldspar to produce anorthite, allowing release of K^+ and silica. The two would enter muscovite, sericite and probably margarite. Al_2O_3 would also be liberated from the K-feldspar and form corundum (Brandt, 1946). Petrographic studies appear to support this model. The common field association also suggests some kind of reaction between the pegmatites and ultrabasic rocks, accompanied by remobilisation of Al_2O_3, K_2O and SiO_2, to produce plumasite or marundite. Whether marundite will form or not may be controlled by the temperature of the invaded rocks at the time of intrusion (Du Toit, 1917 in Robb and Robb, 1986).

4.1.6 Emerald and Columbite-Tantalite Deposits

The Gravelotte Emerald Mine in the Murchison Greenstone Belt area has the single largest emerald open-pit mine in the world (Robb and Robb, 1986). In the main Cobra Pit, the emeralds are set between intruding tonalitic gneisses and country-rock talc-chlorite-actinolite-biotite schists which constitute an offshoot of the Murchison Belt (Fig. 4.6).

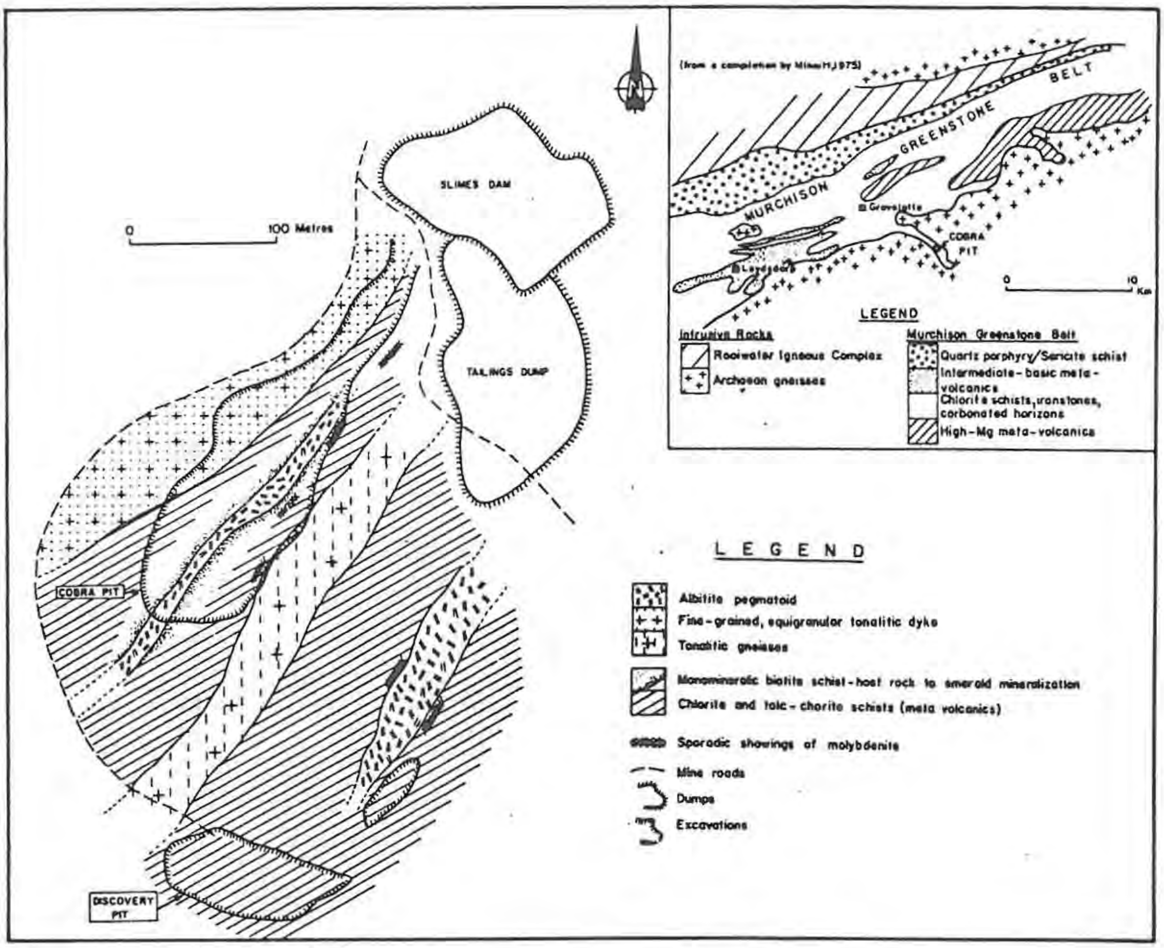

FIGURE 4.6 *Map of the generalised geology around the Cobra Pit Gravelotte Emerald Mine. Inset is a simplified geological map of the Murchison Greenstone Belt showing the regional geological setting and location of the pit (after Robb and Robb, 1986)*

Mineralisation is always linked to albitic pegmatoids, and is characteristically erratic (Van Eeden, 1939). The pegmatoids consist of plagioclase with some quartz, muscovite, beryl and pyrite. The emerald zone is 30 to 40 metres wide and consists of biotite-spodumene schists occurring on margins of the albitite pegmatoid (Fig. 4.6).

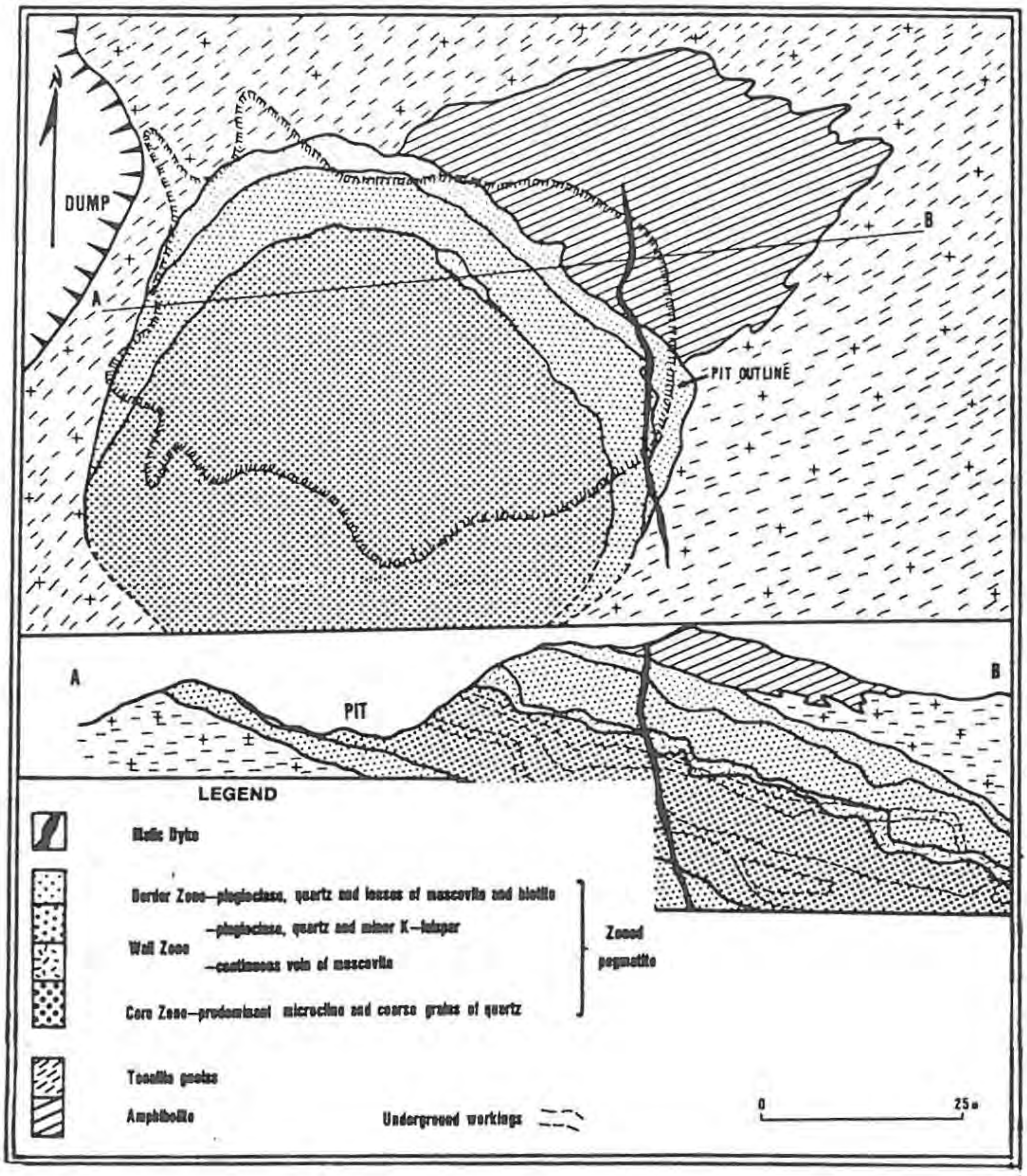

FIGURE 4.7 *Map and cross-section of the Union Mine, Mica District, northeast Transvaal, South Africa (after Robb and Robb, 1986)*

The main process of emerald formation is believed to be Be-metasomatism. This is accompanied by K-metasomatism which causes chlorite and talc to alter to biotite (Robb and Robb, 1986). Emerald coloration is highly variable and seems to be deeper where the content of Cr is higher. At similar deposits in Colombia, it has been argued that the Cr is transported as a chloride-complex (Beus, 1979). Close examination of some of the stones also indicates that movement of such solutions was controlled by the fracture system inside the beryl. This may account for the wide variation in colour.

4.1.7 Quartz-Feldspar-Mica Deposits

The largest quartz-feldspar-mica mine is Union, 40 km west of Phalaborwa (Robb and Robb, 1986). The pegmatite outcrop is roughly circular, dips 30°E, and intrudes gneisses containing greenstone remnants. The pegmatite is asymmetrically zoned with respect to the circular plan. The mineralogy is simple. Unaltered, strain-free and unspotted muscovite is recovered from the border zone which consists also of plagioclase, quartz and biotite, and from the wall zone where some K-feldspar is also present. K-feldspar and quartz is mined out of the core zone.

There are a few similarities between the Murchison deposits and deposits from the Hartz Range in Central Australia. These include (i) the presence of better-quality mica in the hanging wall, (ii) association of better-quality micas with plagioclase as opposed to K-feldspar and (iii) the occurrence of high-quality mica in the core zone (Joklik, 1955). In addition to quartz, feldspar and mica, a mine at Palakop (Fig. 4.4) produces small tonnages of columbite-tantalite as a byproduct.

4.1.8 Tungsten Deposits

The Shangoni tungsten deposit may be a higher-level or distal manifestation of the modes of mineralisation just outlined, since it consists mostly of late-stage cross-cutting pegmatite, quartz veins and stringers. The tungsten and accompanying emerald is also associated with zones of albitised and biotitised tremolite or tremolite-actinolite schists (Furnell, 1986).

4.2 THE ZIMBABWE CRATON

4.2.1 General Geology

This craton is bound by the Zambezi Belt to the north and north-east, the Limpopo Belt to the south, and the Mozambique Belt to the east (Fig. 4.1). As opposed to the Kaapvaal Craton, the Zimbabwe Craton is composed predominantly of greenstone belts and subordinate granites and granitic gneisses (Wilson, 1981). The greenstone belts fall into three distinct stratigraphic groups. In order of younging these are (i) the Sebakwian, (ii) the Bulawayan and (ii) the Shamvaian. The belts are associated with granitic rocks, ultramafic intrusions and a swarm of mafic dykes which culminated in the intrusion of the Zimbabwe Great Dyke at approximately 2.5 Ga (Wilson and Wilson, 1981).

The Sebakwian is preserved in the south-east of the craton in the Selukwe, Fort Victoria and Shabani areas. It comprises metasediments, mafic and ultramafic volcanic lithologies and associated plutonic rocks, and is dated at 3.5 Ga (Robertson, 1973). The igneous rocks have Sr ratios compatible with a basic lithosphere of upper mantle origin.

The 2900 Ma Bulawayan Group comprises diorite, adamellite and tonalite. The belt associated with it has two successive greenstones which are separated by an unconformity. The lower greenstones lie to the south-east and consist of quartzites, chloritic grits and conglomerates, high-Mg lavas, locally pillowed and interbedded with phyllites. To the west, where they are more extensive, the lower greenstones are divisible into four formations: a lower mafic and acidic volcanic unit, a succeeding mafic unit with spinifex texture, a conglomeratic and agglomeratic unit and, finally, sedimentary rocks and banded-iron-formations.

The Belingwe Greenschist Belt is taken as an example of the upper greenstone belts. It consists of a 7km volcanic pile of massive and pillowed lavas divided into a lower mafic volcanic unit and a succeeding tholeiitic basalt unit. These are overlain by shallow-water clastic rocks, banded-iron-formations and limestones. The Shamvaian Group was deposited in a number of isolated basins and consists of immature arkose and subgreywacke, indicating derivation from a volcanic and granitic source. It is also polymictic and divisible into four units which are separated by unconformities.

The Sesombi and Chilimanzi granites post-date the phases of greenstone belt formation. The Sesombi suite granites comprise tonalites and granodiorites of syn- to post-Shamvaian age. They show isotope characteristics compatible with mantle-derived materials. The Chilimanzi suite granites constitute the so-called Younger Granite Suite, and their intrusion, the last major granitic event of the Zimbabwean Basement Complex, was dated at 2.6 Ga. The granites are adamellitic to granodioritic (Wilson, 1981).

The structural evolution in the Zimbabwe Craton was controlled by relative south-westward movement of the Zimbabwe Craton against the Kaapvaal Craton during Limpopo orogenesis. The many circular intrusions may have experienced the "billiard-ball effect" within the craton, causing local shear zones and intense flattening. This is known to be as much as 50% for greenstone belts in south-western Zimbabwe and eastern Botswana. There is a belt-scale concentric metamorphic zonation with lower greenschist facies in the centre, rising into upper greenschist and amphibolite facies as the rim and surrounding belts are approached. On this is superimposed the intracratonic, pluton-associated amphibolite facies metamorphism (Wilson, 1981).

4.2.2 The Bikita "Tin" Field

Widespread pegmatite-based mineralisation appears to be associated genetically with the late-tectonic Chilimanzi Suite of granites (Broderick, 1981). An account of the Bikita "Tin" Field, which was the largest Li producer and the 6th largest Be producer in the world during the 1960s and 1970s, is presented. The review will illustrate additional aspects of Archaean rare-element mineralisation.

Lithium-rich pegmatites are emplaced into partly metamorphosed greenstones, and metasediments of the Bulawayan Group (Section 4.2.1) (Martin, 1964). Their age is still uncertain but they intrude the greenstone belts (Fig. 4.8). The pegmatites are lensoid or tabular, show a regularity of strike and a length-to-depth ratio of 5 : 2. Where exposed, the ends are sharp. They do not show regional zonation, and are therefore assumed to have been produced by one or similar batches of magma. Local fractionation within the confines of the field may account for differences between the pegmatites.

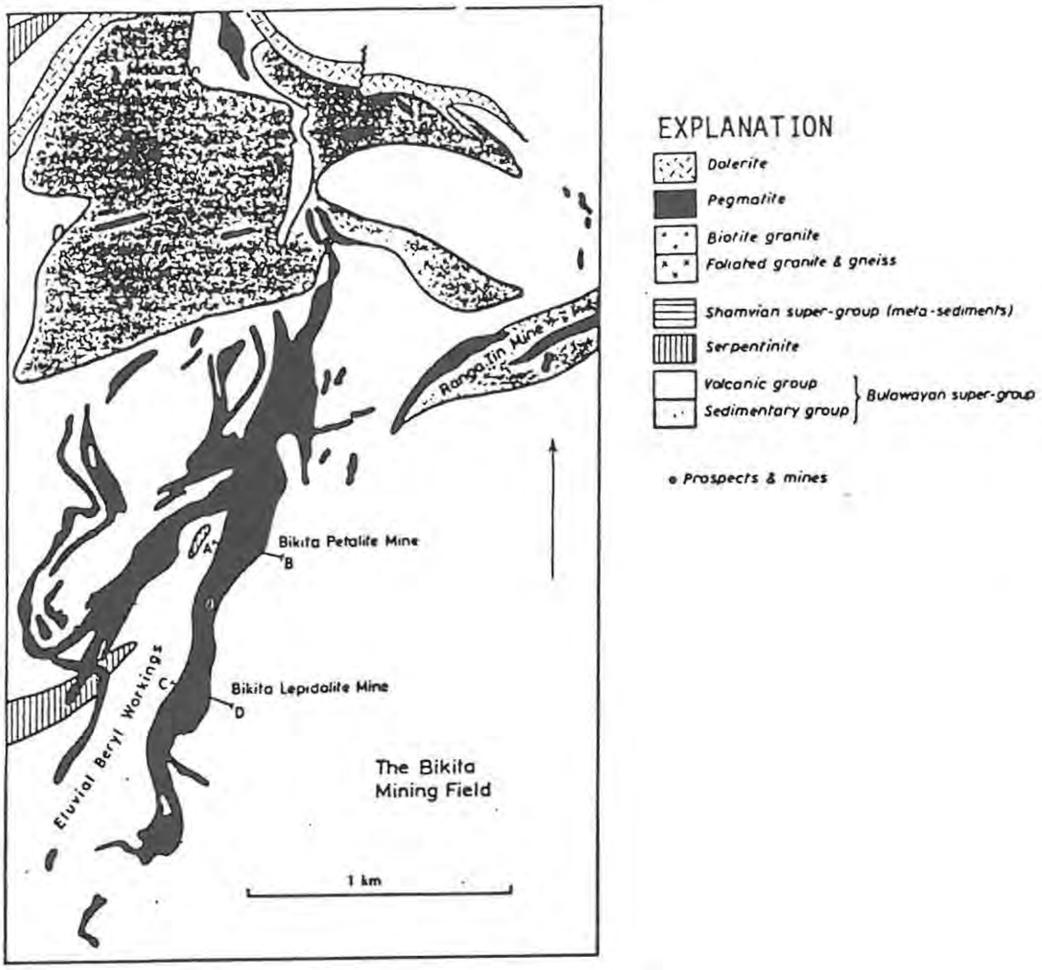

FIGURE 4.8 *The Bikita Mining Field (after Dixon, 1979)*

The Main Pegmatite is 1500 metres in length, and on average 50 metres in width. Dips vary between 15 and 44°. The Bikita Lepidolite and the Al Hayat quarries had 65% of the Li reserves of the field in 1964. The Main Pegmatite has large bodies of Li-minerals with the main ore mineral, lepidolite, being associated with petalite, spodumene, amblygonite and eucryptite in addition to beryl. The Bikita Section is part of the Main Pegmatite.

4.2.3 The Bikita Section

The footwall border zone consists of blocky greenstone referred to as "spotted dog zone" (Fig. 4.9 and 4.10). It is a persistent marker horizon 0.80 to 3m in thickness. It has a 2 to 10cm lining in which zinnwaldite is associated with albite and quartz. Being 15 centimetres in thickness, the hanging-wall border zone is thicker than that of the foot-wall.

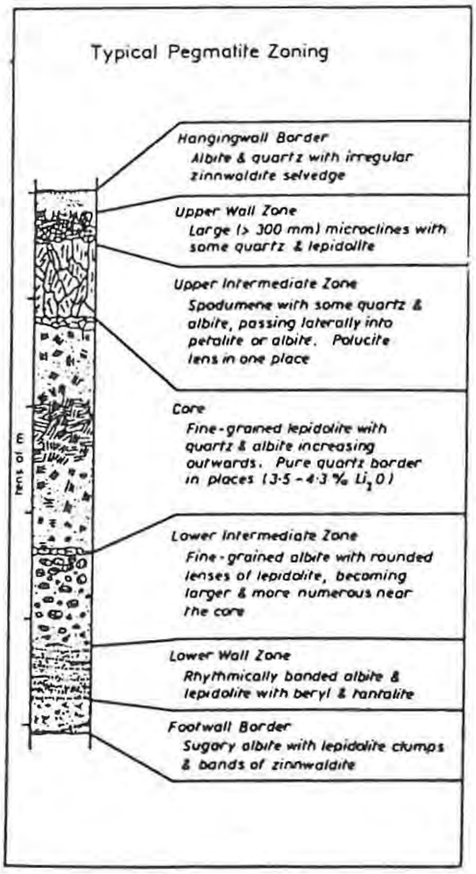

FIGURE 4.9 *Typical zoning in the pegmatites of the Bikita Field (after Dixon, 1979)*

The wall zones comprise albite, oligoclase, quartz, mica with sporadic beryl. This is the main carrier of beryl at Bikita. Rhythmic and very concentric layering occurs due to antithetic variation between albite and lepidolite. The upper wall zone is 0.8 to 6m in thickness and contains microcline and quartz. It is the main zone from which microcline, with crystals up to 30 centimetres is produced.

The intermediate zones consist of albite, petalite, spodumene and pollucite. The lower intermediate zone is rich in Li, and has fine intergrowths of albite, lepidolite and minor quartz. Beryl, which typically occurs in the wall zone disappears in the transition to the intermediate zone, but aggregates 15 to 50cm across occur near the core to form what has locally been termed the "cobble zone". The upper intermediate zone shows the mineralogical assemblage in which spodumene > quartz > albite > > petalite. Spodumene and quartz are intergrown while albite is sugary, probably suggesting that the latter is of replacement type. The zone hosts a pollucite body roughly 186m x 7.5m x 314m. This consists of roughly equal amounts of lepidolite, petalite, feldspar and pollucite.

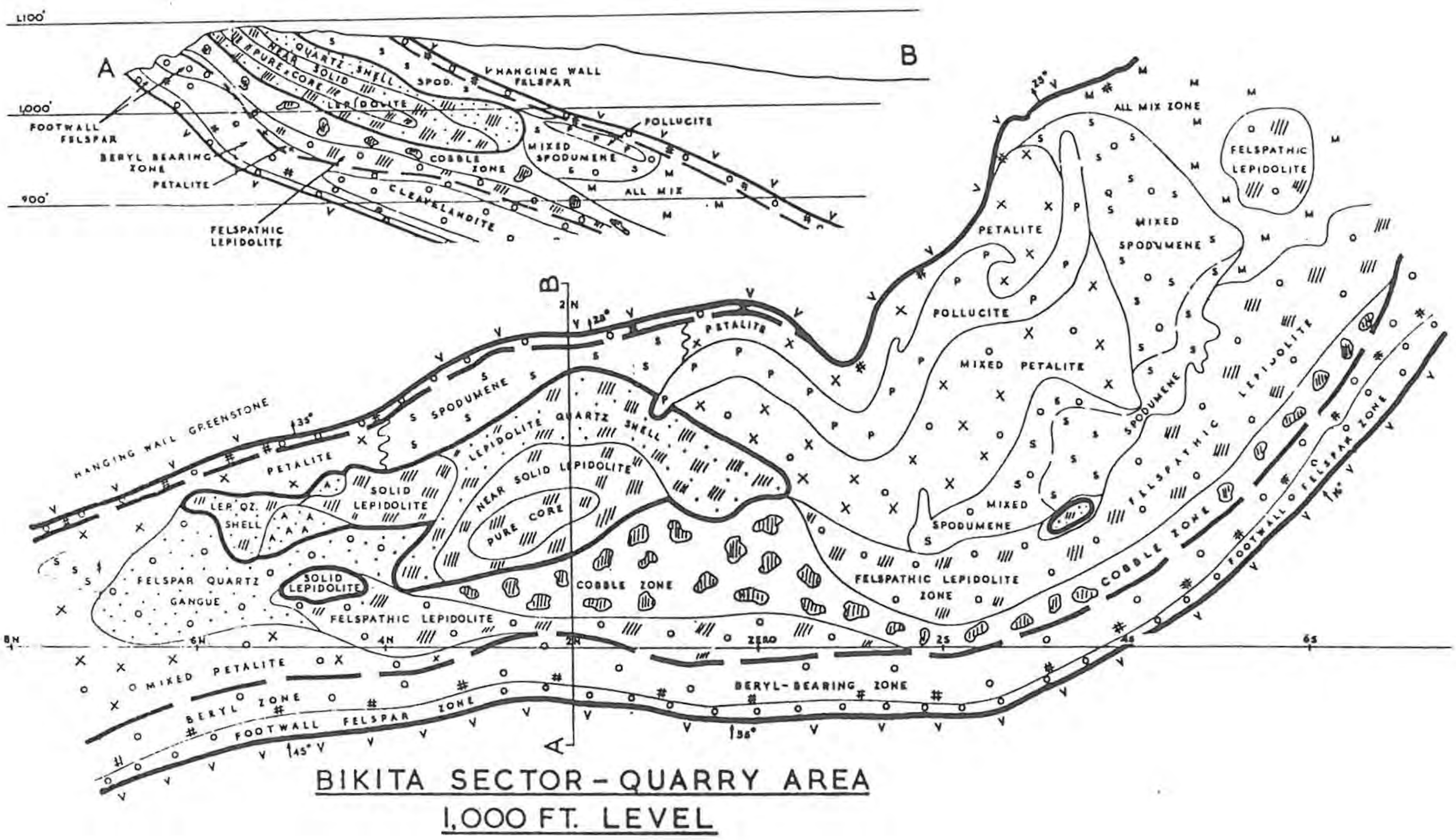

FIGURE 4.10 Map and cross-section (inset) of one of the levels of the Bikita Sector, part of the of the Bikita Pegmatite Field, Zimbabwe (after Cooper, 1964)

The core zone is the principal economic unit of the Main Pegmatite. 450,000 tons of lepidolite with 3.5 to 4.3% Li_2O had been produced by 1964 (Martin, 1964). It is continuous for 330 metres in one part. Asymmetry is shown by its variable distance from the hanging wall, which ranges between 14 and 90m. Telescoping is also displayed by its occassional gradation into intermediate zone assemblages. Mineralogically, it consists of massive lepidolite. Chemical comparison of the core with the intermediate zone indicates higher lithia contents with correspondingly reduced amounts of quartz and albite.

Other minerals recovered from the pegmatites at Bikita include Al- and manganotantalite, microlite, cassiterite and beryl. Also present are insignificant quantities of euxinite, tetrahedrite, fergusonite, malachite, azurite and chalcocite.

4.2.4 Genesis of the Bikita Pegmatites

The genesis, emplacement and mineralisation is not related to a specific and obvious alkaline parent granite. The pegmatites are, however, likely to be linked to the Younger Granite Suite referred above (Section 4.2.1). Unlike the Kaapvaal Craton bodies mentioned (Sections 4.1.2 & 4.1.4), the Bikita pegmatites have not had any significant interaction with the country rocks. This suggests intrusion into cold rocks by pegmatitic magma already enriched in the rare-elements. The country rocks were not so cold, however, as to cause marginal chilling. Evolution and consolidation was along the lines already proposed (Section 2.1.1 & 2.1.3).

An alternative model involves intrusion of a normal (K-rich) zoned pegmatite followed by intrusion of a Na-rich melt, to account for the albitic character (Grubb, 1973). This model is not endorsed as evidence for a later crosscutting albitic magma is not available.

4.3 THE ANGOLA-KASAI CRATON

65% of Angola is covered by Creteceous and Kalahari rocks and sand, so that geological knowledge of the Precambrian of the area is very poor. An early Archaean to Proterozoic Angola-Kasai Craton can be inferred from gravity data (Pretorius, 1981). Outcrop occurs in three provinces are recognised within the craton: (i) the northeast (ii) the northwest and (iii) the central and southwest of Angola (Anhaeusser and Wilson, 1981).

In the north-east province, six geological complexes have been distinguished, mainly on a geochronological basis (Cahen and Snelling, 1966). All the complexes are reviewed here, although some or parts of some are younger than the Archaean, because the limited data does not permit subdivision into the different sections of this report to be made. The complexes are:

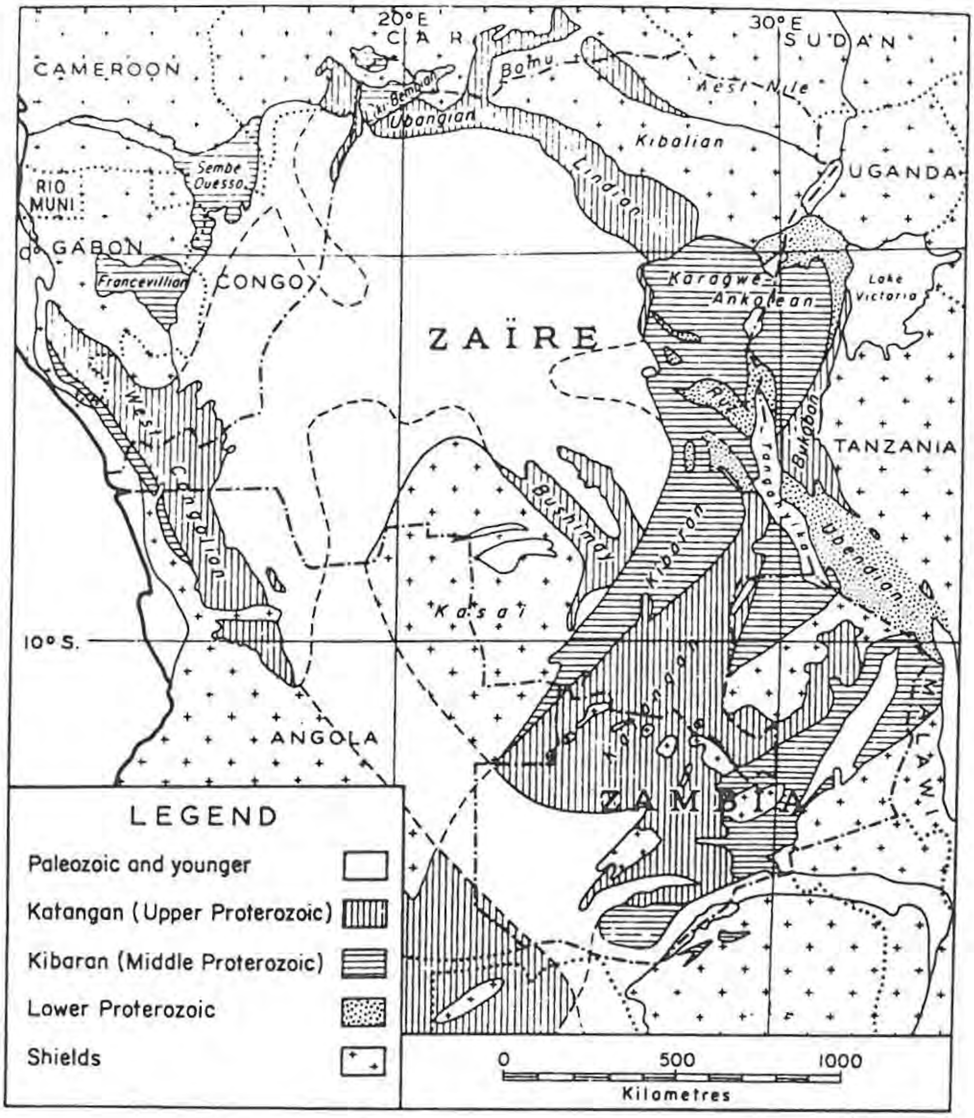

FIGURE 4.11 *General geology of Zaire and Zambia and adjacent areas*

1. the 3300 Ma Upper Luanyi Gneiss,
2. the 2.7 Ga Dibaya Granite-Migmatite Complex,
3. the 2120-1910 Ma Kasai-Lomani Noritic Gabbro-Charnockite Complex,
4. the Luiza Metasedimentary Complex,
5. the Lulua Sedimentary and Volcanic Complex and, finally,
6. the mainly granitic complex south of 8^{o}S.

The Luiza metasediments have been involved in migmatisation, charnockitisation, deformation and uplift accompanied by pegmatite invasion.

In northwest Angola, the granitic basement which appears below the West Congo Geosyncline, is considered to be 2500 Ma or older by southward extrapolation from the Zanidian orogenic zone of the Congo Republic and Gabon (Anhaeusser, 1981).

Central and south-west Angola is underlain by the Metamorphic Series ranging in age from 1690 to 1400 Ma (Cahen and Snelling, 1966). This consists of schists, metaquartzites and amphibolites, and is overlain by crystalline limestones and gneissic to migmatitic rocks. The area also has a wide variety of granites and porphyritic granitoid bodies ranging in ages from 2080 to 925 Ma. The Kunene Anorthosite Complex is the largest known mass of its type and is composed of massive leucocratic anorthosite, made up mainly of plagioclase (An_{50-66}) with minor amounts of orthopyroxene, clinopyroxene and olivine (Anhaeusser, 1981).

The Early Precambrian rocks host pegmatites with mica, beryl, quartz and uranium deposits. Information on their modes of occurrence is not available.

4.4 ARCHAEAN NUCLEII OF CENTRAL AFRICA

These form a ring of shields with the Angola-Kasai Craton to the west, the Gabon Shield to the northwest, the basement for the West Nile Complex to the north and north-east, the Tanzanian Craton to the east, the Bangweulu Block to the south-east, and the Zimbabwe Craton to the south (Fig. 3.1 & 4.11). The Tanzanian Craton is taken as lying just outside the northern limit of the area of this study. The Bangweulu Block in northeast Zambia is, on the basis of recent studies (Andersen and Unrug, 1984), regarded as an Ubendian (early to mid-Proterozoic), rather than as an Archaean block. No significant mineralisation of early Precambrian age has been reported. However, they are reviewed to emphasise that they display similarities to the Kaapvaal, Zimbabwe and Angola-Kasai cratons where mineralisation has been found, and that further exploration in them may lead to discoveries of pegmatite-hosted mineral deposits.

4.4.1 The Gabon Shield

The Gabon Shield or Massif du Chaille lies in the northwest of Zaire (Fig. 4.11) and has similar lithologies to those of the Angola-Kasai Shield although. It also displays radiometric ages from 700 Ma into the Archaean (Vachette in Clifford 1970).

4.4.2 The Western Nile Complex

This stretches from north-east Zaire, north and east into Uganda and the Sudan (Fig. 4.11). Archaean basement consists of the Ancient Gneisses which may be as old as 3500 Ma (Cahen and Lepersohne, 1967). It consists of Kibalian formations which have a minimum age of 2050 Ma, and appears to be Archaean in habit, since it comprises greenstone belts as elongated patches in granite. It has, however, been classed as Lower Proterozoic and correlated with the Buganda-Toro of Uganda. The Bomu Formation in the Central African Republic appears to be its continuation.

4.4.3 The Tanzanian Craton

The central part of this craton consists of granite, granodiorite, acidic gneisses, migmatites with deformed and metamorphosed Dodoman greenstone relics. Its metamorphism probably occurred at 2.6 Ga (Clifford, 1970; Wendt et al, 1972). In the north, Nyanzian greenstone belts are composed of basic and acid volcanic rocks, quartzites, pelites with banded-iron formations and have a pre-2.8 Ga Rb-Sr isochron age given by the post-Nyanzian Migori Granites (Dodson et al., 1975). The Nyanzian greenstones are overlain by arenaceous and argillaceous sediments, and volcanic rocks of the Kavirondian Group, which is older than 2.5 Ga (Clifford, 1970). Thus the Tanzanian Craton appears to be of a similar age to the Kaapvaal and parts of the Zimbabwe cratons, but older than the adjacent Bangweulu Block.

4.5 SUMMARY OF EVOLUTION AND MINERALISATION

There are two main schools of thought as to how the Archaean Cratons may have formed. One school states that evolution of the Archaean cratons may have involved some form of plate-tectonics, probably with higher rates of plate accretion, destruction, partial melting and crustal growth than at present, since the earth's internal energy can be considered to have been greater at the time (Windley, 1987). Support for this model is seen in deep sections of exhumed parts of orogenic belts such as in the Andes and Himalayas. Here, rocks with characteristics similar to those of Archaean terranes are present. The greenstone belts compare well with back-arc-basin sequences, while the granite-gneiss terranes correspond with a combination of the associated continental margin and magmatic arcs of modern subduction zones (Windley, 1987). The late- to post-tectonic granite suites bordering the greenstone belts from which the pegmatites evolve also have analogues throughout geological time in all orogenic zones.

An alternative model views the greenstone belts as having evolved on pre-existing sialic crust in intracontinental settings. The granitic-gneiss terrane is taken as the precursor sialic crust. Radiometric dating gives older ages for the sialic materials than for the associated greenstone belts which supports this model. This seems to be the case for the Kaapvaal Craton and its Barberton Greenstone Belt (Hunter, 1991; Sections 4.1.1 to 4.1.4). The model is also supported by the Zimbabwe Craton settings (Section 4.2.1), which show cycles of greenstone and granitoid development, indicating that processes of granite-greenstone terrane development did not take place during a single major global event, but occurred in cycles. They also occurred diachronously in the different parts of the world. Thus, a typical age relationship between the greenstones and granite-gneisses applicable to all terranes is not necessary in the first place.

Finally, it is likely that both models are valid depending on which Archaean nucleus is being examined (Anhaeusser, 1991). Whichever model is adopted, the key to finding pegmatite-hosted mineralisation in the Archaean stable regions seems to lie in locating of the late- to post-tectonic granites with respect to greenstone belt deformation, and searching for associated bodies of rare element pegmatites.

5.THE LIMPOPO BELT AND ASSOCIATED MINERAL DEPOSITS

5.1 INTRODUCTION

This is the oldest belt under study. It is an east-northeast-trending high-grade terrane which underwent several tectonic and metamorphic cycles between 3200 and 2000 Ma. It separates the Zimbabwe and Kaapvaal cratons by 260 km and is 680 km long (Fig. 4.1). Its north and south boundaries are gradational into the cratons and are defined by the orthopyroxene isograd (Robertson and Du Toit, 1981; Shackleton, 1987). To the west, it passes under younger cover where it probably dies out. It is truncated to the east by the Mozambique Belt. It is divided into three main zones on the basis of lithology, structure and metamorphism (Fig. 5.1) namely the Northern Zone, the Central Zone and the Southern Zone. More recently, seven domains distinguished on the basis of stratigraphy, structure and metamorphic style have been recognised (Watkeys, 1983; Barton, 1983; Anhaeusser, 1991). Only the traditional three zones are considered here, since the more recent subdivision appears to be a refinement only (Fig. 5.1).

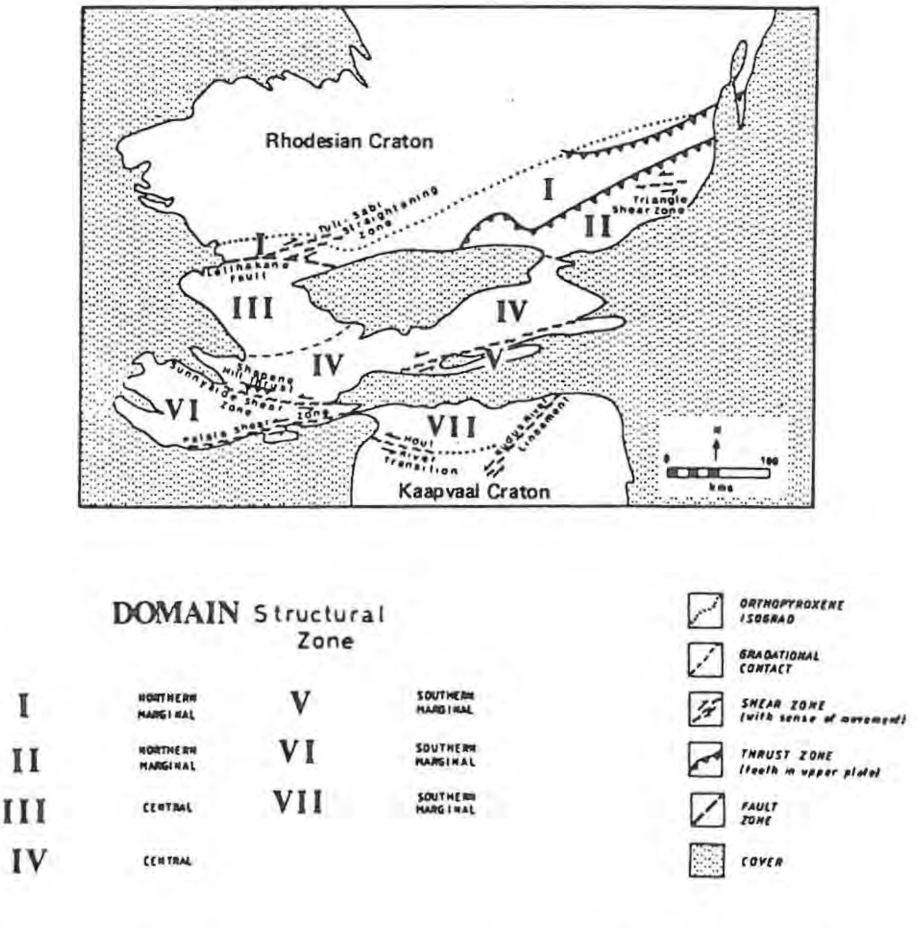

FIGURE 5.1 *Main domains of the Limpopo Mobile Belt (after Anhaeusser, 1991)*

5.1 THE NORTHERN MARGINAL ZONE

The North Marginal Zone lies in Zimbabwe and Botswana, and consists of reworked granite-greenstone terranes occurring as amphibolite and granulite facies assemblages including charnockites, enderbites, and scattered, deformed and foliated metasediments and orthogneisses. The granulite facies components equilibrated at 5kb and 750-800°C so that the previous history is obliterated.

5.2 THE CENTRAL ZONE

The Central Zone shows internal structure oblique to the belt. In the middle, basement occurs as inliers of generally tonalitic and mafic dykes. It has been metamorphosed repeatedly and bears relics which underwent granulite facies metamorphism. The succeeding supracrustal Limpopo Group is referred to as the Beitbridge Group in Zimbabwe and as Assemblage IIA to Assemblage G2 Paragneiss in Botswana.

The Central Zone records three structural events in the Botswana area. The first produced a north-northeast-trending fabric, and was accompanied by high-grade metamorphism which occurred at 7.6kb and 770°C (Wakefield, 1974) at 2660$\pm$60 Ma (Hickman and Wakefield, 1975). The second event enhanced earlier folds and occurred under amphibolite facies, which was accompanied by localised partial-melting. The last phase produced dome-and-basin structures under lower amphibolite facies. Shearing occurred towards the end, and brittle deformation is recorded at 2000$\pm$100 Ma.

In Zimbabwe, the Macuville Group shows pre-Beitbridge deformation. Extensive anatexis of the basement produced such plutons as the Bulai Granite at about 2700 Ma.

In the west, around Mahalapye in Botswana (Fig. 5.1), major structures swing westwards. This appears to be related to intrusion of post-tectonic granites dated 2240$\pm$40 to 2010$\pm$80 Ma (Van Breemen, 1968). The granites intruded the Limpopo Mobile Belt Central Zone basement and cover, and resemble the Porphyritic Granite and Matok Granite of the northern and southern marginal zones, respectively, although they did not rise as high.

5.3 THE SOUTHERN MARGINAL ZONE

The poorly exposed Southern Marginal Zone is composed mainly of disjointed and contorted, keel-like remnants of pelitic, quartzo-ferrugineous, mafic and ultramafic rocks equated with rocks from the Sutherland greenstone schists. Deformation occurred in four phases similar to the north and central zones. Anatexis was more advanced, and where extensive, produced granites. Emplacement of pyroxene-rich syenite such as the Schiel Pluton at 2100 Ma marked the end of activity for the Southern Marginal Zone (Robertson and Du Toit, 1981).

5.4 EVOLUTION OF THE LIMPOPO MOBILE BELT

Various plate-tectonic models have been proposed to explain the evolution of the Limpopo Belt. They include dextral motion of up to 200 km (Coward, 1976), development and poly-phase deformation of an intracratonic basin (Barton and Key, 1981) and subduction of ocean crust (Fripp, 1982). Thus the evolution of the Limpopo Belt is still not fully understood. What seems to be clear, however, is that the there was a period of crustal thickening, in which supracrustal rocks underwent very high grades of metamorphism, followed by uplift leading to exposure of the high-grade rocks (Shackleton, 1987).

5.5 POTENTIAL FOR PEGMATITE MINERALISATION

Granulite facies rocks form at temperatures and pressures in which pegmatitic materials could not have been formed and preserved. Intrusion of the granulites by felsic pegmatites would be possible if water, needed to trigger partial melting (Section 2.11 and 2.12), were introduced under the granulite terrane subsequent to its uplift into regions where pegmatite preservation could occur. However, a mechanism to introduce water in the Limpopo Belt such as subduction has not been identified.

Thus the Sand River Gneiss at the base and in the centre of the Limpopo Mobile Belt contains no mineralisation. Some pegmatite bodies and quartz veins are known to host beryl, corundum, sillimanite, magnesite, talc and andalusite in the Northern Marginal Zone. Tungsten-, tantallum-, beryllium- and lithium-bearing pegmatites and pegmatitic veins are also spread symmetrically about the belt. Emeralds occur at Klein Letaba in South Africa while high-quality emeralds have been mined in the Mewza Schist Belt in Zimbabwe (Phaup in Odell, 1975). The deposits are considered to be equivalents of those described for the greenstone-granite terranes, but which lie within the area affected the "Limpopo deformation".

6. EARLY TO MIDDLE PROTEROZOIC MOBILE BELTS AND THEIR RELICS

6.1 GENERAL GEOLOGICAL OUTLINE

The lower Proterozoic belts are poorly represented and fragmentary. Their age is dubious in some cases. This is mainly due to lack of information, either because belts have not been studied adequately, or due to extensive surficial cover. Most of the belts and blocks have also undergone remobilisation and therefore occur as isolated terranes with no clear regional trends. Geological assessment and correlation other than geochronologic is therefore difficult. An example of this is the Bangweulu Block which was grouped with Archaean cratons until recent studies showed it to be of Ubendian age (Andersen and Unrug, 1984).

6.2 THE WESTERN SEGMENT OF THE ZAMBEZI BELT

In the western outcrops of the Zambezi Belt, an inlier of metamorphic rocks with ages as early as 2120 Ma (Dodson et al., 1969) is surrounded and overlain by unmetamorphosed and undeformed Sijarira, Karoo and Kalahari cover (Fig. 6.1). Referred to as the Dete-Kamativi Inlier, it consists of four north-east trending belts of metamorphosed and highly deformed supracrustal rocks known as the Kamativi, Tshontanda, Inyantue and Malaputese. The belts overlie, and are surrounded by granite-gneiss (Fig. 6.1).

Three major fold phases are evident. An initial F_1 accompanied high-grade metamorphism of the supracrustal rocks, producing NE-trending isoclinal folds and local large-scale recumbent structures in the south-eastern inlier. Granite intrusion probably preceded F_2, during which more open northwestward structures were formed. Tight F_3 folds were mostly co-axial to F_1 and deformed both the granitic rocks and the supracrustal belts (Lockett, 1981).

TABLE 6.1 Geochronologic dates from the Zambezi Belt

METHOD	MATERIAL	AGE
Rb-Sr	granite (w.r.)	2150 $\pm$ 100 Ma
Rb-Sr	lepidolite (from Lutope)	2100 $\pm$ 20 Ma
Pb-Pb	galena (Elbas Mine)	1310 $\pm$ 39 Ma
		1250 $\pm$ 38 Ma
K-Ar	biotite (granite)	990 $\pm$ 15 Ma
Rb-Sr	Sn-pegmatite	988 $\pm$ 20 Ma
Rb-Sr	biotite (granite)	940 $\pm$ 30 Ma

w.r. = whole rock
(summarised from Broderick, 1981)

Geochronologic data (Table 6.1) indicate that the Zambezi Belt was involved in events which occurred prior to 2000 Ma but was reactivated by subsequent activity, and in particular the Namaqua-Natal orogenesis. The Sn mineralisation seems to be related to this event (Lockett, 1981).

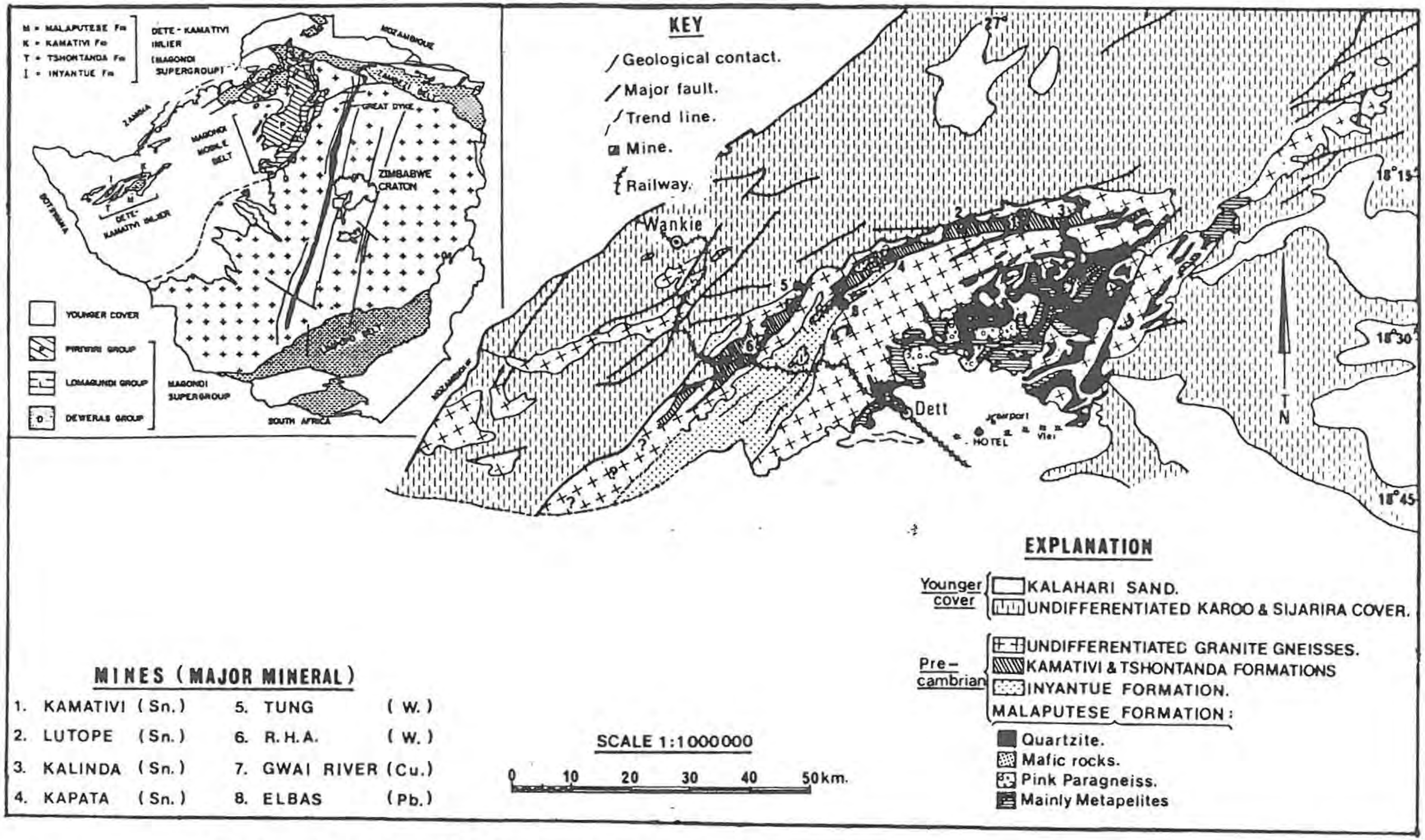

FIGURE 6.1 Geological map of the western part of the Zambezi Metamorphic Belt, an example of early to middle Proterozoic belts (after Broderick, 1981)

No pegmatite-hosted mineral deposits of early to middle Proterozoic is reported from the area. Those mined are associated with late-stage granites of Irumide (Kibaran) age and are considered in a later section (7.4.2). However, one of the granites from which the lepidolite age was obtained (Table 6.1) may have potential as a source for pegmatite.

6.3 THE NORTHERN SEGMENT OF THE ZAMBEZI BELT

Early to middle Proterozoic rocks are also found in the northeasterly segment of the Zambezi Belt (Fig. 6.2). , where they occur as relics in an area metamorphosed and deformed intensely by the Irumide and Mozambiquan. Resting on the Archaean Basement Complex (2600-3500 Ma) are paragneisses to the west of Mount Darwin, and schists to the east (Fig. 6.2; Anderson, 1972; Broderick, 1981). These have a whole-rock isochron age of 2300-2500 Ma. They were metamorphosed before deposition of the Magondi Supergroup (Master, 1991).

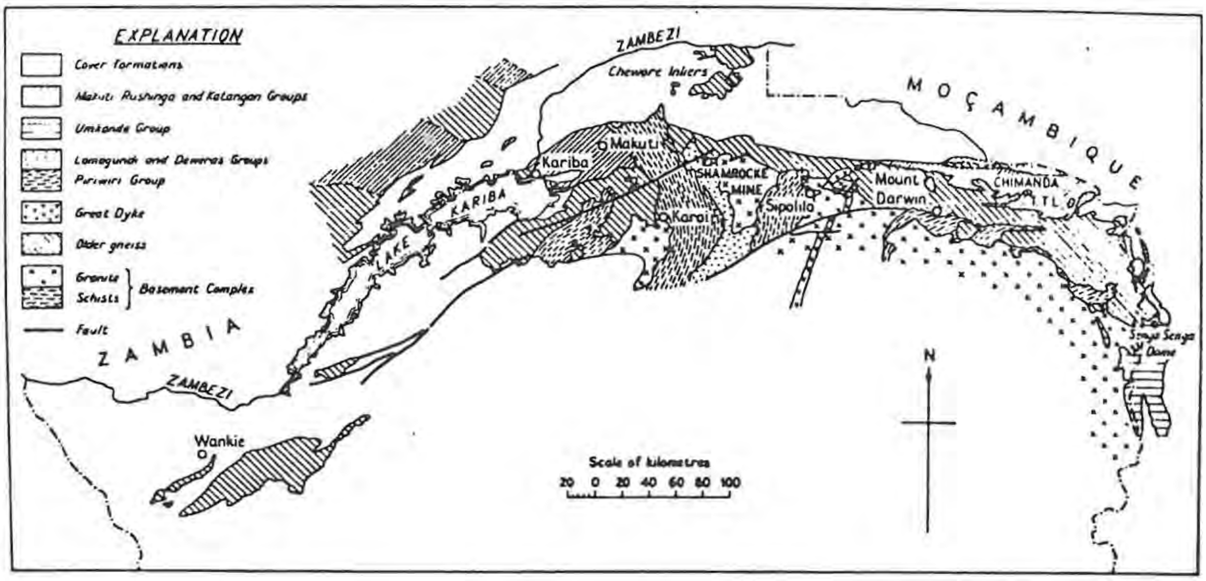

FIGURE 6.2 *Simplified geological map of the Zambezi Metamorphic Belt (after Broderick, 1981; See Fig. 6.1, inset, for location.)*

6.4 WESTERN CENTRAL AFRICAN BELTS

In Angola, Gabon and the Western Nile Complex (Fig. 4.11), evidence of involvement in tectonothermal activity approximately during the period is given by a concentration of radiometric ages between 1700 and 1300 Ma, which are really neither early Proterozoic nor Irumide but somewhere in between. The Western Nile Complex, which stretches from north-east Zaire, has already been noted to consist in part of rocks of early to middle Proterozoic age (Section 4.4.2; Cahen and Lepersohne, 1967). The Bomu Formation in the Congo may be of a similar age. Pegmatite mineralisation is not reported.

6.5 THE RUZIZIAN BELT

This is 1800 Ma old, and stretches from eastern Zaire to Burundi (Fig. 4.11), passing through an area where it has been reworked or covered by the Kibaran. It then bends southeastwards into southern Tanzania where it is known as the Ubendian. The latter segment passes through north-eastern Zambia before entering north Malawi, where it is known locally as the Misuku Belt (Thatcher, 1974 Fig. 4.11). The general stratigraphy of the Ruzizian Belt is outlined below:

TABLE 6.2 Sequence of Ruzizian Sediments

Upper Ruzizian	schists, phyllites, amphibolites, quartzites, phyllitic conglomerate, biotite schist
	mica schists and phyllites, graphitic or calcareous in parts
Lower Ruzizian	
	quartzites, arkoses, dark phyllitic schists quartzites

(after, Mendelsohn, 1981)

6.6 THE BANGWEULU BLOCK

This lies in the north-east of Zambia, and has a lithological suite analogous to that of a craton (Clifford, 1970; Kroner, 1977). However, recent findings indicate that it formed in mid-Proterozoic times (Andersen and Unrug, 1984). It consists of a gneissic basement, comprising schists, acid metavolcanic rocks, and granitoids of Ubendian magmatism. The schists occur in east-southeast-trending belts typically 75-100 km long and 10-20 km wide. Batholiths lie mainly between schist belts and bodies of lavas. Examples of these are the Mambwe and the Luchewe granodiorites, with Rb-Sr whole-rock ages of 1869±40 and 1824 Ma, respectively. They intrude the margins of the schist belts.

Two episodes of deformation are recorded by the schists. The first was very variable but the second forms structures parallel to the belts. Later granitoids are smaller and intrude the margins of the schist belts. They constitute a suite comprising gabbros, diorites, tonalites and large granites at the edge of the Ubendian belt, and they show a whole rock Rb-Sr intrusive age of 1838±Ma (Andersen and Unrug, 1984).

The Ubendian event does not seem to have been intense enough to reset the rocks completely. Since the oldest dates found are Ubendian, the basement of the block is considered to be of this age (Andersen and Unrug, 1984), and not Archaean (Clifford, 1970; Kroner, 1977; Drysdall et al, 1972; Cahen and Snelling,1966). The basement on which it may have formed is probably represented by the Jembia River Granulites of northern Malawi (Fitches, 1974). The cover sequence or Mporokoso Group is 5000m in thickness and comprises three formations referred to, in younging order, as the Kasama, Loitikila and the Luapula. These have Kibaran or more recent ages (Section 7.3.1).

6.7 SUMMARY

Pegmatite mineralisation of this age has not been reported. However, the presence of lithian mica in granites of this age from the Dete-Kamativi inlier (Table 6.2) indicates the probable presence of lithian pegmatites. Some of the granite-gneiss terranes and schist belts blocks found in the region such as the Bangweulu indicate these as areas of potential for deposits similar to those found in the Archaean blocks of the area.

7. UPPER PROTEROZOIC BELTS AND ASSOCIATED MINERALISATION

7.1 GEOLOGICAL OUTLINE

The upper Proterozoic belts were formed between 1400 and 1000 Ma. They are represented by the Namaqualand-Natal Metamorphic Province in the south, and the Kibaran and Irumide belts in central Africa (Fig. 3.1). The general geology of the Namaqualand Metamorphic Complex, for which information is readily available, is reviewed before outlining the other more northerly belts.

7.2 THE NAMAQUALAND METAMORPHIC COMPLEX

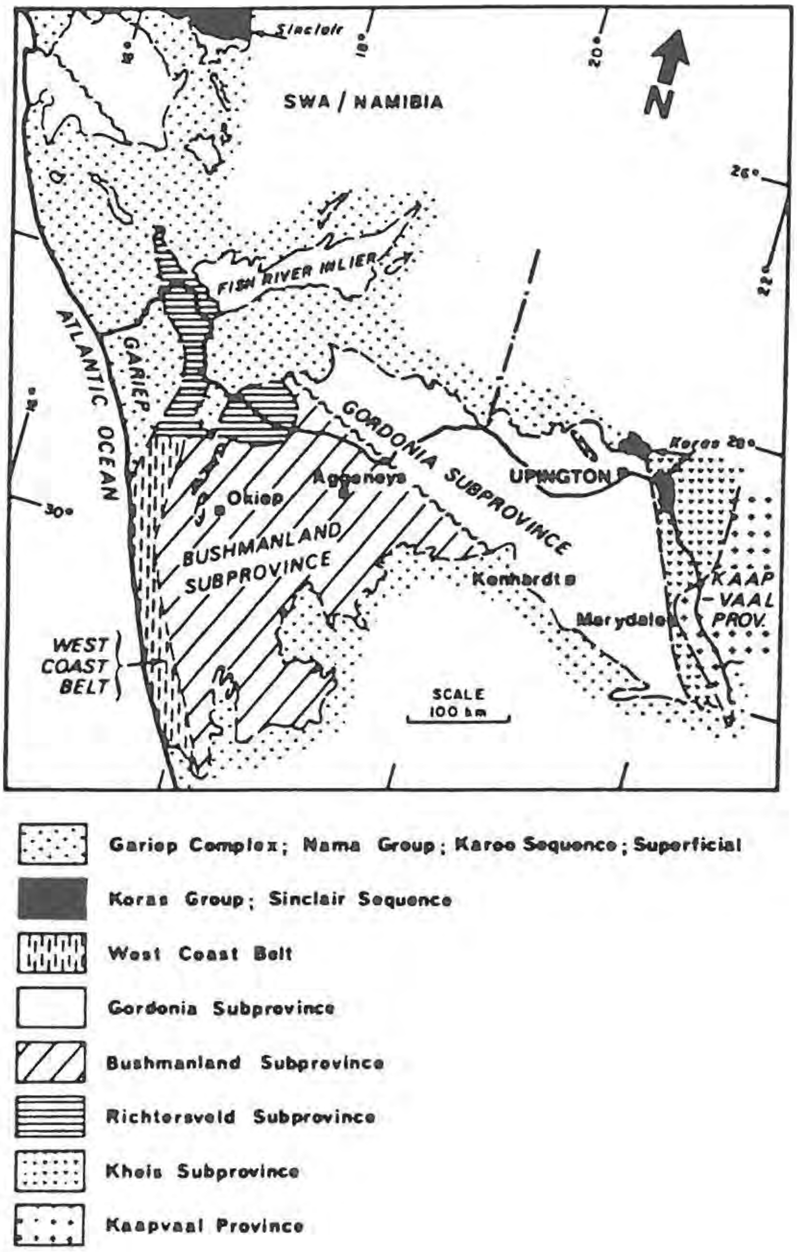

FIGURE 7.1 *Geotectonic subdivisions of the Namaqualand Metamorphic Complex (after Joubert, 1986 and Anhaeusser, 1991)*

7.2.1 Geological Outline

The complex consists of gneisses formed by several cycles and varying degrees of deformation and metamorphism. It appears to have been built by an amalgamation of several terranes which can be distinguished by their different structural, lithological and geochronological properties in addition to their geographic locations. Seven terranes are recognised (Fig. 7.1; Joubert, 1986). As the West Coast and Gariep belts have a strong Pan-African overprint, they are considered together with the Damara Orogen (Section 8.2.1).

A terrane model has been proposed, in which an active Andean-type continental margin (the Richtersveld Subprovince 1950-1800 Ma), which had hitherto been producing calc-alkaline volcanic and plutonic rocks due to westerly subduction of the Gordonia Subprovince, collided with an exotic micro-continent mass (the Bushmanland Subprovince) at about 1800 to 1750 Ma. Due to partial melting of the subducted continental mass, intrusive rocks being produced changed from dominantly calc-alkaline to leucocratic types. A protracted period of tectonism, associated granite intrusion, and high temperature metamorphism ensued between 1800 and 1100 Ma with the latter part constituting the Namaqua event (Moore, 1990) with which pegmatite mineralisation is associated (Hugo, 1969; 1986).

7.2.2 Pegmatite Mineralisation

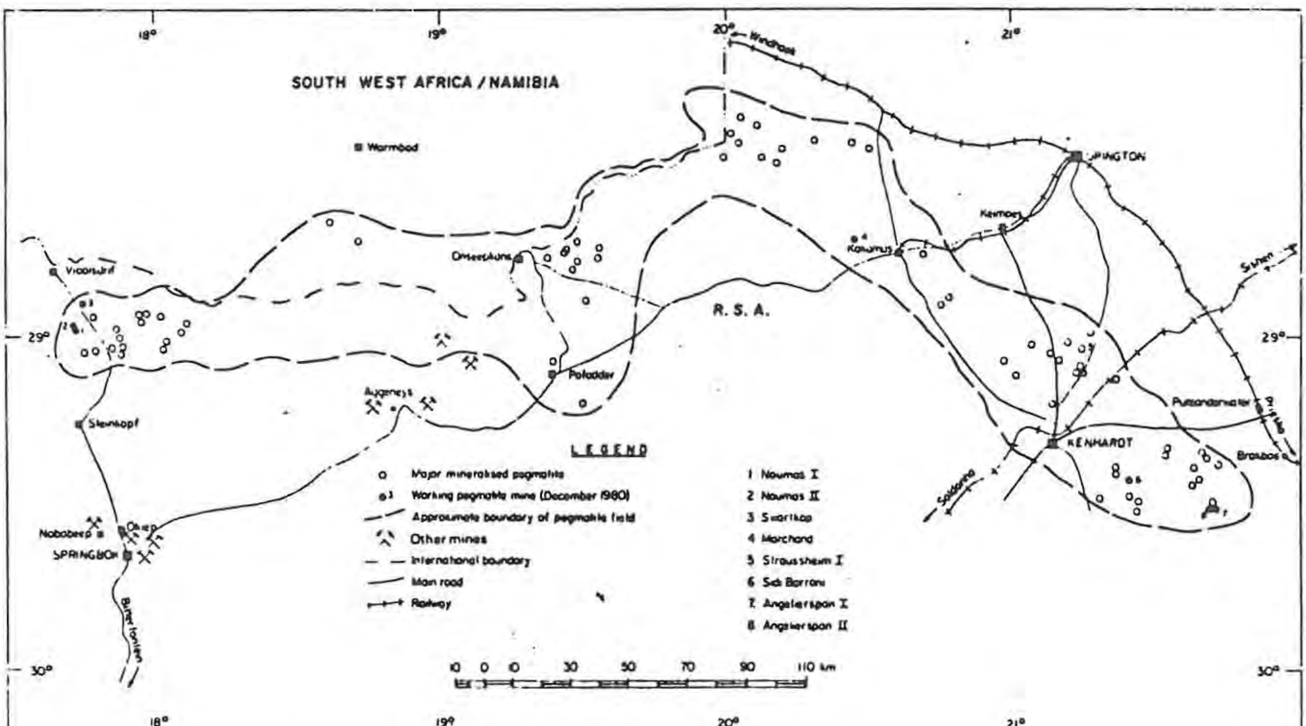

FIGURE 7.2 *Extent of the pegmatite belt in the North-western Cape Province and southern Namibia (after Hugo, 1986)*

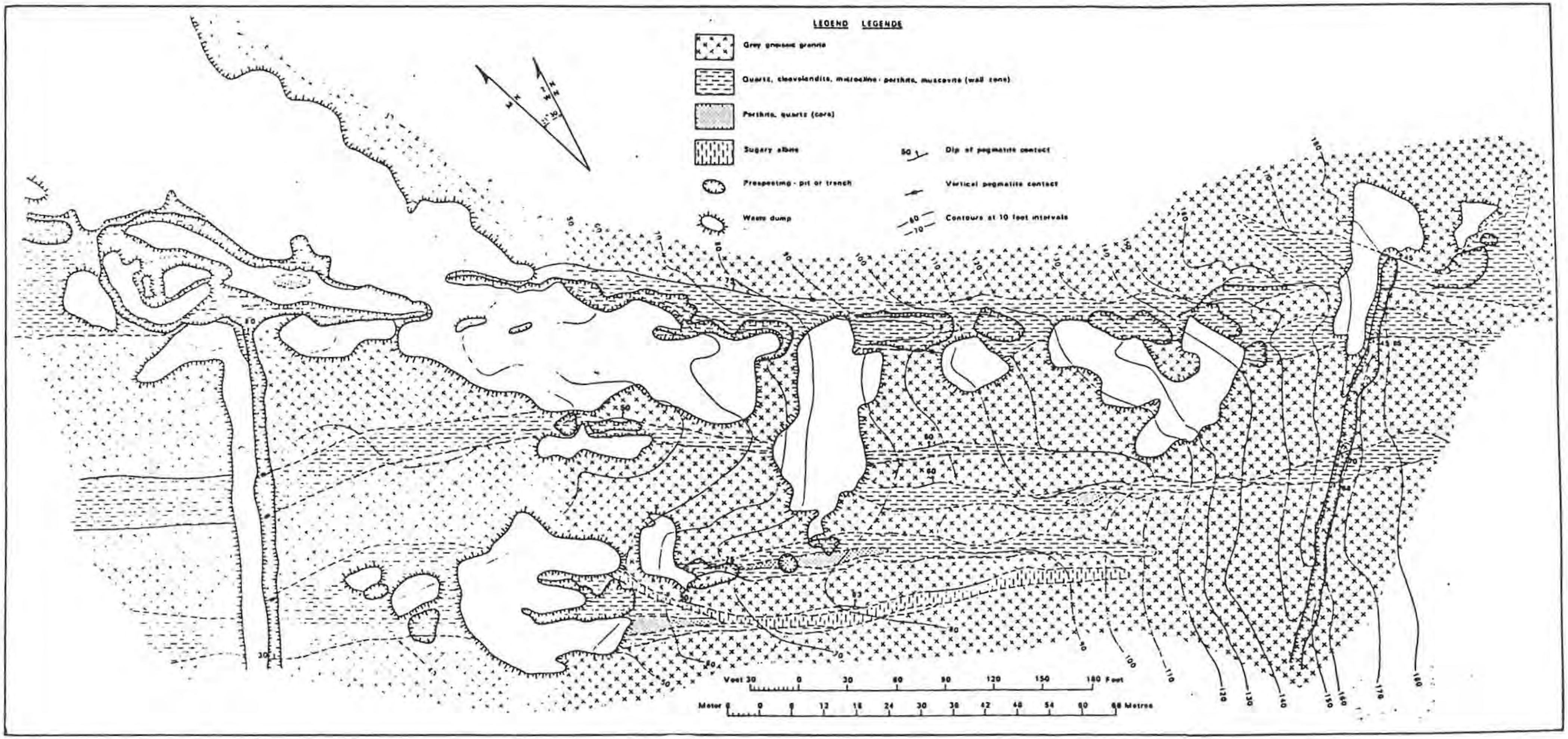

FIGURE 7.3 The Noumas II and other adjacent smaller pegmatites, Orange River Pegmatite Belt (after Schutte, 1972)

A major pegmatite field for the region of Africa under review occurs around Kenhardt, north of the Orange River and along it in a westerly direction, but remaining south of the Richtersveld Subprovince (Fig. 7.1 and 7.2). It is referred to as the Orange River Pegmatite Belt (Hugo, 1986). The belt is 450 km in length and 15 to 30 km in width. The main commodities obtained are feldspar, mica and beryl. Past supplies figures show that 30% of K-feldspar, 10% of mica, and virtually all South Africa's beryl and tantalite-columbite needs were met from mines of the area. Bismuth, and rose quartz were also produced. Untapped resources of these minerals are believed to be abundant (Hugo, 1986).

The deposits are set in pre-tectonic sediments and volcanic rocks of the Gordonia and Kenhardt districts (Fig. 7.1), and are concentrated in the syntectonic intrusive rocks of the Vioolsdrif, Hoogoor and Keimoes Suites (Joubert, 1986), and rocks in their immediate vicinities. The suites are considered as parental to the pegmatites. Two ages of pegmatites have been found with emplacement dates of 1000 Ma and 950 Ma, and these have been assessed to be syn- and post-tectonic types, respectively (Hugo, 1969). Difficulties are encountered with regard to positive identification of parental granites in the Orange River Belt since the pegmatites are of the type emplaced at great depth (Sections 2.3, 2.4, 2.10).

The pegmatites range in size from a few centimetres in thickness to very thick bodies of up to 3 km in length. Irregular masses are also known. All the basic units of pegmatites such as zones, replacement-bodies fracture-fillings are present. Homogeneous and inhomogeneous types are also known.

7.2.3 Feldspar: The Noumas II Pegmatite

This is hosted in biotite-granodiorite, is 250m in length and 10 to 25m in width, and plunges steeply eastwards and more gently northwestwards (Fig. 7.3). It is well zoned, with a 0.5 to 5m core zone consisting of K-feldspar and quartz in the ratio 4 : 1. The 2- to 3m-thick intermediate zone is composed of quartz, albite, K-feldspar, muscovite and beryl. The wall zone is 0.5 to 3m in thickness and comprises graphic granite consisting of albite-oligoclase, quartz and accessory muscovite with schorl. The core and intermediate zone are partly replaced by bodies of up to 3m in diameter and consist of greissens of muscovite, cleavelandite and some columbite-tantalite. The main economic mineral, microcline-perthite, is pure white and has a K_2O content of 13.5%. The amount of beryl and tantalite-columbite is not known while mica is of low grade. Similar pegmatites are the Swartkop, Marchand, Blomeras and Angelierspan II bodies.

7.2.4 Mica: The Noumas I Pegmatite

Dominantly mica-producing mines are based on pegmatite bodies at Noumas I and Straussheim. Mining of feldspar and beryl is also necessary in order to be economic (Hugo, 1986). The mica is recovered from the wall and intermediate zones. Brown, sheet-mica leaves reach 0.1 to 0.5m in diameter. The Noumas pegmatite is the largest of these pegmatites. It is dyke-like, at least 1000m in length, and 10 to 42m in thickness. It is hosted in foliated granite (Fig. 7.4). It was mined for beryl and bismuth with some tantalite and columbite between 1951 and 1960. Mica and spodumene were recovered during the 1960's, while tantalite-columbite was also mined in addition in the 1980's.

Zoning is well-developed, but is asymmetrical, with a border zone 10- to 150mm-thick, and consisting of microcline-perthite, quartz, muscovite and accessory garnet. The wall zone is 1 to 6m wide, and comprises microcline-perthite, muscovite, quartz and albite-oligoclase. Also present in this zone are beryl, bismuth, apatite, triplite, garnet and mica books up to 0.75m long. The intermediate zone is 0.6 to 8m in thickness, and consists of spodumene, cleavelandite, quartz and accessory microcline-perthite, beryl and tantalite-columbite. The core is composed of anhedral quartz and large crystals of microcline-perthite. Replacement bodies are 1.5 to 9m across and consist of lithian muscovite greissen with cleavelandite, microcline-perthite, tantalite-columbite, microlite, thorianite, orangite and gummite. To a depth of 30m, the Noumas II Pegmatite was estimated in 1986 to have 80,000 tons of mica, 6000 tons of beryl and 1200 tons of tantalite-columbite with 52 to 60% Ta_2O_3 (Hugo, 1986).

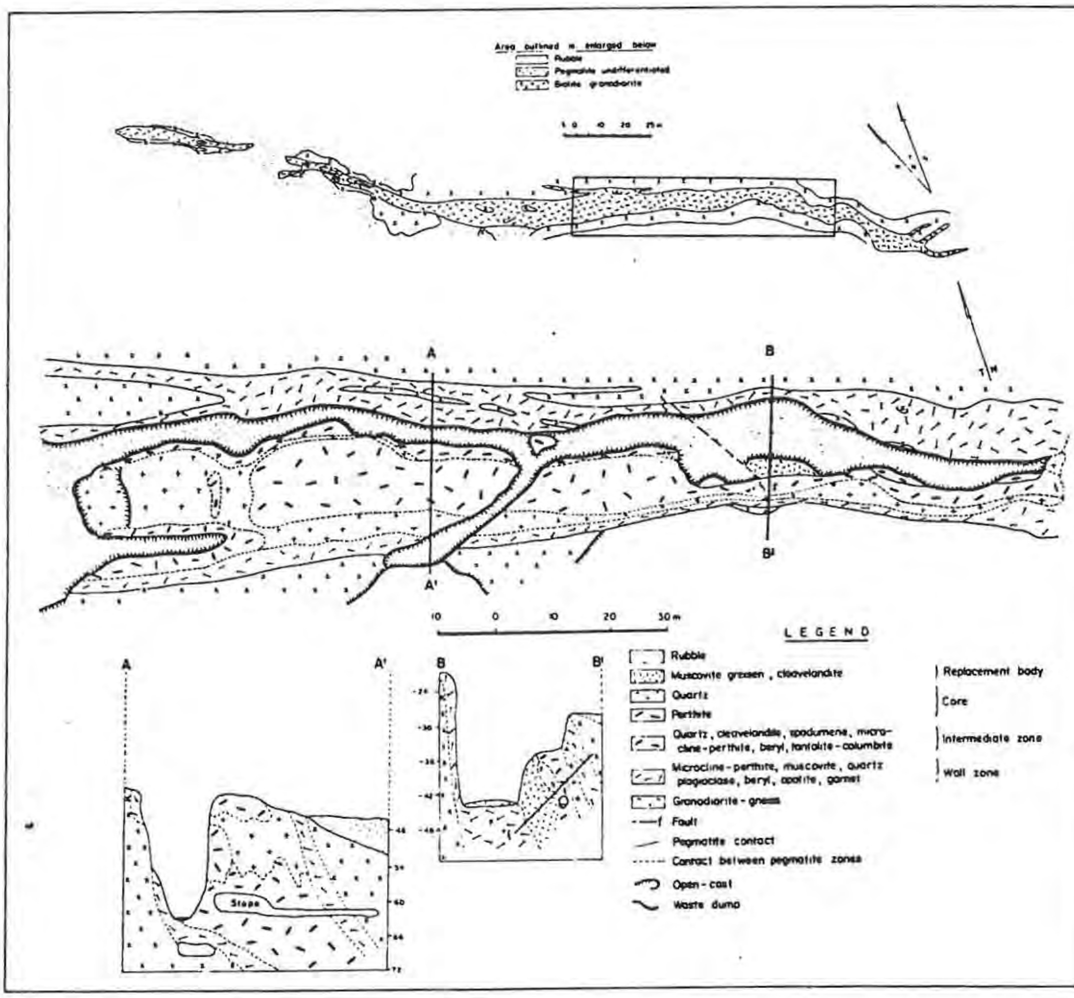

FIGURE 7.4 *Geological map and sections of the Noumas I pegmatite, Namaqualand District, North-western Cape Province, South Africa (after Hugo, 1986).*

7.2.5 Beryl: The Angelierspan I Pegmatite

This pegmatite dyke strikes north-south, has a length of 100m, a width of 8m with a known minimum depth of 13m. Trench workings did not reach the border zone. The wall zone is 8m in thickness and consists of albite-oligoclase, microcline-perthite and muscovite with schorl. Granophyre blocks range is sizes between 0.25 and 1.5m. The core consists of milky quartz and white or pink feldspar. Beryl is recovered from the core, and the boundary between the core and the wall zone. Its distribution is irregular, while bodies as large as 6m across are known. One body weighing 60 tons was mined. Replacement bodies lie inside the core, and consist of muscovite greissen with sugary albite and euhedral tantalite-columbite.

7.3 DAMARA BASEMENT

Bodies of basement crop out within areas of the Pan-African sediments of the Damara Belt (Fig. 3.1). The basement occurs as inliers in the north and south marginal zones, and in the median zone of the belt. It consists predominantly of granitoid and infolded supracrustal sequences. The Rehoboth Inlier, for example, is composed of 1400 to 900 Ma old Irumide-Namaqua granites, and has a cover sequence of volcanic and sedimentary rocks. Central Zone inliers lie in cores of basement mantled-gneiss-domes, and show dates ranging from 1925 to 300 Ma. Northern basement inliers lie in an arcuate zone from Grootfontein to Mocamedes, and consist of granitoid gneisses. Similar Irumide-Namaqua-Kibaran sequences are preserved in northwest Botswana. Despite the presence of volcanic rocks, real evidence for subduction in the area is lacking. The rifts are also separate and may have evolved intracontinentally.

Granites intruded during the Irumide event did not evolve sufficiently to yield mineralised pegmatites (Mason, 1981).

7.4 THE KIBARAN

7.4.1 General Geology

The Namaqua-Natal Belt extends north-northeast into central Africa where it branches into northwestward Kibaran and northeastward Irumide branches (Fig. 3,1, 4.11 and 7.5). In the type area, the sequence of lithologies deposited and affected by the Kibaran consists dominantly of fine clastic and sedimentary rocks with phyllites and schists as major components, and occasional conglomerate, basalt, and a limestone cap (Table 7.1; Cahen and Lepersohne, 1966).

TABLE 7.1 Sequence of Kibaran Sediments

ROCK TYPES	PROBABLE THICKNESS
Limestone; dolomitic, siliceous, stromatolitic	0500m
Schist, graphitic; phyllite; quartzite	3000m
Quartzite; conglomerate, phyllite, basalt	3000m
Phyllite; sericite or chlorite schist; local conglomerate; limestone; rhyolite	4000m

(After Mendelsohn, 1981)

Sedimentation is believed to have already begun by 1900 Ma and to have continued to the beginning of the Kibaran Orogeny, 1250 to 1300 Ma ago (Clifford and Gass, 1970). Accompanying igneous activity continued until 500 Ma ago, but occurred intermittently. In contrast to the Namaqua Belt to the southwest, metamorphism affecting the whole belt was typically low-grade. Thrusting and folding took place, also along northeast-trending lines but is aligned northwestward in the extreme north. Dome-and-basin structures were produced in the Karangwe-Ankole sequence.

Remnants of the Kibaran rocks or underlying basement reactivated by the same event occur in a belt from the northwest of Zimbabwe, and stretch northeastwards across Zambia, where the front extends from northwest of Malawi to about the middle of Lake Tanganyika (Fig. 7.5). In this area, it abuts against the northwest-southeast-trending Ruzizian-Ubendian Belt. Here, a lateral ramp is formed with northwest-southeast strike slip movements occurring between the Ubendian and Kibaran or Irumide (Daly, 1987). In the Zambia area, the belt is composed of quartzites, metavolcanic rocks, marble, graphitic schists, pelite, migmatitic sequences (Drysdall et al., in De Vletter et al, 1972 P 725).

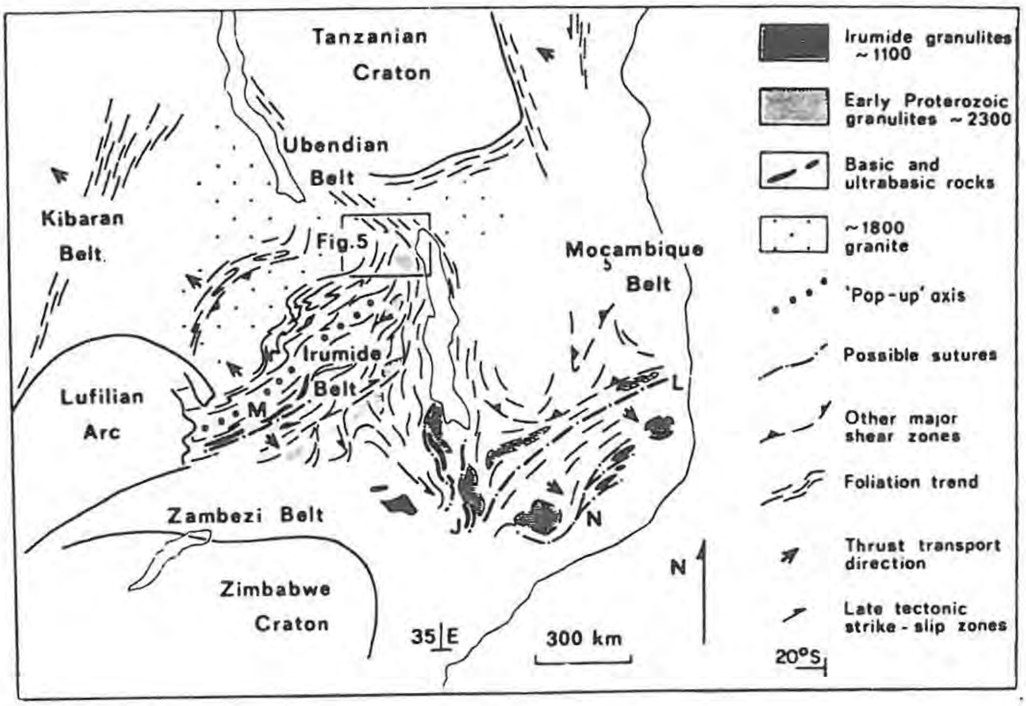

FIGURE 7.5 *Sketch map of the Ubendian, Irumide and Lurio belts showing regional and foliation trends and movement directions L - Lurio Belt; N - Namama foreland (after Daly, 1987)*

There is increasing evidence that areas which were believed to comprise Mozambique age rocks in fact consist of substantial remnants of the Irumide with little and in some cases no overprint by the younger event. This seems to be the case in northeastern Mozambique and southern Malawi, where the Lurian Belt appears to be a reappearance of the Zambezi Belt of northeastern Zimbabwe cropping out of the post-Cambrian strata (Andreoli, 1984,1989; Kirkpatrick, 1968; Morel, 1961; Sacchi et al., 1984).

The Precambrian sequence of the area has been determined to consist of a gneiss and migmatite basement with a cover of composite metasedimentary and volcano-detrital material. A granulitic unit was thrust over the basement and cover. The thrusting was partly accompanied and outlasted by intrusion of granitoids subdivided into three classes. The first are well foliated and have a minimum Kibaran age (1100 Ma) and low Sr_i ratios (0.7027). The second type are undeformed and are surrounded by migmatitic aureoles of Kibaran age and having low, ill-defined Sr_i ratios. The third class are the youngest and are circular, porphyritic granites with a weak and mainly linear fabric. They are suspected to have been derived, at least in part, from the earlier granites. Their Sr_i ratios are high (Sacchi et al., 1984).

Structural trends are Kibaran since they are northeast or east-northeast, as opposed to the north-south grain of typical Mozambiquan regions. Tilting has produced nappes in the forelands of the Lurio Belt as at, whilst thrusting has exhumed granulite facies rocks in southeastern Malawi at the root of the nappes (Andreoli, 1989; Sacchi et al., 1984).

In the northwest-southeast trending West Congo Geosyncline, the West Congolean sequence was deposited in one large depository. Clastic sequences include conglomerates, quartzites, amphibolites (metasedimentary) and mica schists. These were overlain by quartzites, black schists, mica schists and acid volcanic rocks. This is in turn overlain by a tillite before deposition of the succeeding Katangan (Stanton et al., 1963; Cahen and Snelling, 1966; Cahen and Lepersohne, 1967). Probable Kibaran age rocks also occupy the Franceville Sembe Quesso and Liki Bembe orogens, which appear to be northern equivalents of the West Congolean (Fig. 4.10). The Kibaran sequences were not strongly deformed or metamorphosed, except in a few isolated patches where they are associated with syn- to late-kinematic granites.

7.4.2 Deposits in General

The main Kibaran zone forms an important Sn-W-Ta province (Fig. 7.6). The Sn-W-Ta pegmatites occur in Kibaran and Ruzizian formations associated with 1000Ma granites, commonly in the domal areas. Sn-W associations occur in quartz-veins occurring separately from pegmatitic zones. Four granite suites are recognised. The first are syn- to late-tectonic, and intrude anticlinoria, causing contact metamorphism, and were intruded at about 1200 Ma (Pohl and Gunther, 1991). Afterwards, biotite-granites of alkaline chemical character evolved from mantle-derived material were emplaced, and were followed by many small granites intruded between 1000 to 900 Ma. The small granites were associated with pegmatites and pegmatitic granites, with the most evolved being Sn-bearing. They are sub-alkaline, strongly peraluminous, equigranular, biotite-muscovite granites with 5% normative corundum. They are also rich in Li, Cs, Rb, B, F and Sn, and show high Sr ratios. Mineralised granitoids occur in areas of considerably deep erosion.

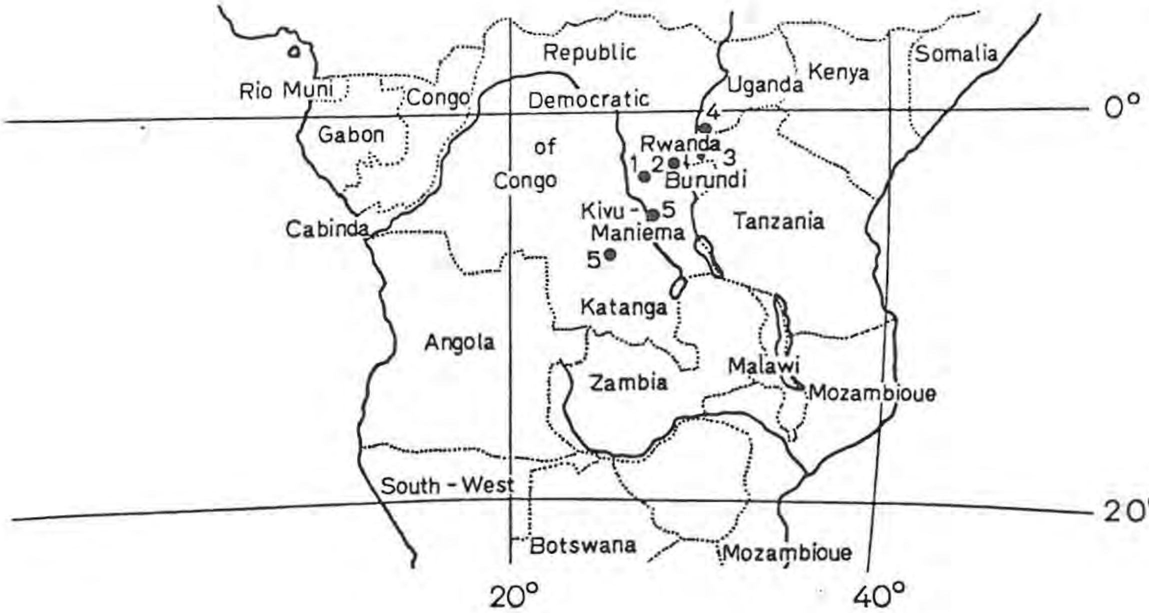

FIGURE 7.6 *The Central African Sn-W-Ta Province (after Varlamoff, 1972)*

Rocks in such areas are esimated to have been uplifted and eroded to expose sections which lay some 12 km depth (Pohl and Gunther, 1991). Pegmatite deposits of upper Proterozoic are also found in the Zambezi Belt segments, in its (?) extension in Mozambique and Zambia, and in Zaire.

7.4.3 Li-Sn-W at Dete-Kamativi, Zimbabwe.

Tin pegmatites have been known in the area since the 1920s while production began in the 1930s. Small workings produced about 4000 tons SnO_2 concentrates (Sn = 70%) by 1950.

At 1000 Ma, mica and Li-Sn-W-base metal sulphide pegmatite deposits intruded early to mid-Proterozoic supracrustal sequences of the Kamativi and Nshontanda formations in the Dete-Kamativi Inlier (Master, 1991; Broderick, 1981; Table 6.1). The pegmatites comprise mainly quartz, feldspar, mica, tourmaline and garnet, and occur in five clusters. Most are concordant with the foliation which is steep to vertical, but the mineralised ones are generally horizontal to sub-horizontal, implying a change in the stress field prior to their emplacement. On the basis of field relationships, three types have been recognised. These are (i) highly deformed quartz-rich veins, (ii) unzoned pegmatites of variable size, composition and texture and, finally, (iii) the mica-rich zoned pegmatites (Lockett, 1979).

Two classes of mineralised pegmatites are known. The first consists of discordant, sheet-like or saucer-shaped concordant of up to 30m wide which have been traced for up to 2km. They occupy domal areas of the structure and dips vary between 20 and 50° (Fig. 7.7) Within these locations, flat-lying joints were the chief sites of magma injection. The second type are dyke-like bodies up to 8m in thickness with steeper dips of 70° or more. These may be discordant to foliation, but strike parallel to the regional trend. The vertical type are regarded as the feeders to the more gently dipping class. Vertical dykes occur sporadically in the spodumene-producing areas.

The only deposit to have been mined on a large scale is the Kamativi Tin Deposit. The pegmatite is zoned and has a wall rock alteration band of about 10 cm in thickness marked by an increase in quartz and tourmaline, and depletion of feldspar in the schistose country rock (Fig. 7.8). It contains cassiterite with lithium minerals and tantalite-columbite (Rijks and Van der Veen, 1972). Wolframite veins of the are small, rarely exceeding 1m in thickness. They are widespread and occur in swarms of thin sheets or as irregular stockworks hosted in the mica schists. Most veins consist of milky-grey quartz, minor tourmaline and variable amounts of wolframite. The wolframite occurs as tabular crystals, partly altered to scheelite. Disseminated Cu- and Fe-sulphides may be present. Accessory amounts of K-feldspar, muscovite and cassiterite may also occur. Wolframite veins are taken as the final stage of the hydrothermal processes responsible for formation the Sn pegmatites.

The Dete-Kamativi pegmatites appear to show intermediate characteristics between those of the Namaqualand Metamorphic Province and those further north. This is seen in the abundance of Sn and W as opposed to mica-feldspar-beryl mineralisation in the southerly belt.

7.4.4 The Emerald Deposits of Zambia

The pegmatites are limited to the Muva Supergroup at the top of the Kibaran sequence of Zambia, and are spatially related to the Irumide granitoids (Fig. 7.9; Siliwa and Nguluwe, 1984). The deposits occur in three east-west trending zones consisting of talc-chlorite-amphibole-magnetite schists, referred to as the Northern, Central and Southern zones (Fig. 7.10). The Northern Zone comprises three schists belts (Miku, Dabwisa and Fibodele) while the Central Zone comprises four belts (Kamubeza, Perala, Fwaya-Fwaya and Libwente) and the Southern Zone of two (Nkabashila and Mitondo) (Fig. 7.10).

The pegmatites consist of feldspar, quartz and muscovite. Emeralds occur sporadically and at the contacts between schists and pegmatites. In these zones, the schists are altered from talc-chlorite-amphibole-assemblages to phlogopite-biotite-tourmaline types. Small amounts of emeralds occur within quartz-tourmaline veins. Rarely, isolated stones may be found in country rocks. Emerald occurrence, even where classified as rich, is erratic. Veins associated with mineralisation have high contents of Be, As, Ag and Bi, implying a low temperature of emplacement (Siliwa and Nguluwe, 1984). They also occur in country rocks with high contents of chromium.

The petrogenetic model to explain the emerald deposits is similar to that envisaged for the Gravelotte Emerald Mine (Section 7.1.4). It involves intrusion of late-tectonic Kibaran-Irumide granitoids into, or adjacent to the schist belts. An early phase of simple pegmatites was injected from the granites to produce a network of quartz-feldspar-tourmaline pegmatites and associated silicification. During a later pneumatolytic stage, emerald and beryl were formed, preferentially in voids created along glide-surfaces. Fluid inclusion studies indicate temperatures of formation from 300 to 500°C (Siliwa and Nguluwe, 1984). Cr was leached out of ultramafic portions of the schists. Subsequent Pan-African deformation has rendered some stones opaque.

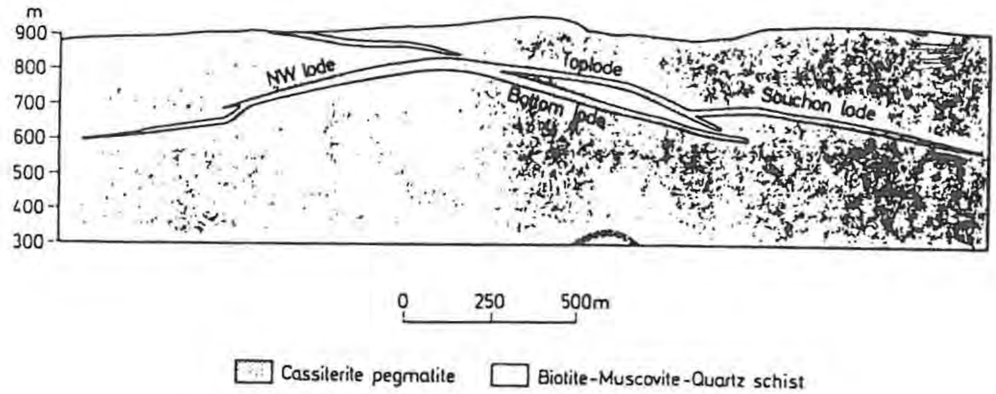

FIGURE 7.7 *Cross-section of a pegmatite from the eastern part of the Dete Kamativi Sn district showing the domal occurrence of of thick mineralised pegmatite bodies (after Rijks and Van der Veen, 1972)*

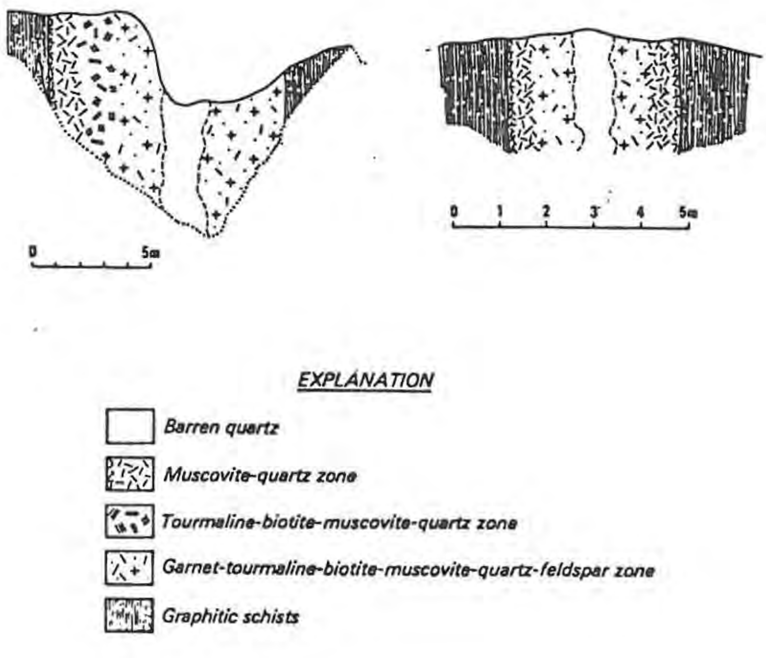

FIGURE 7.8 *Some features in the mica-pegmatites from the Dete-Kamativi Inlier in northwestern Zimbabwe (after Lockett, 1979)*

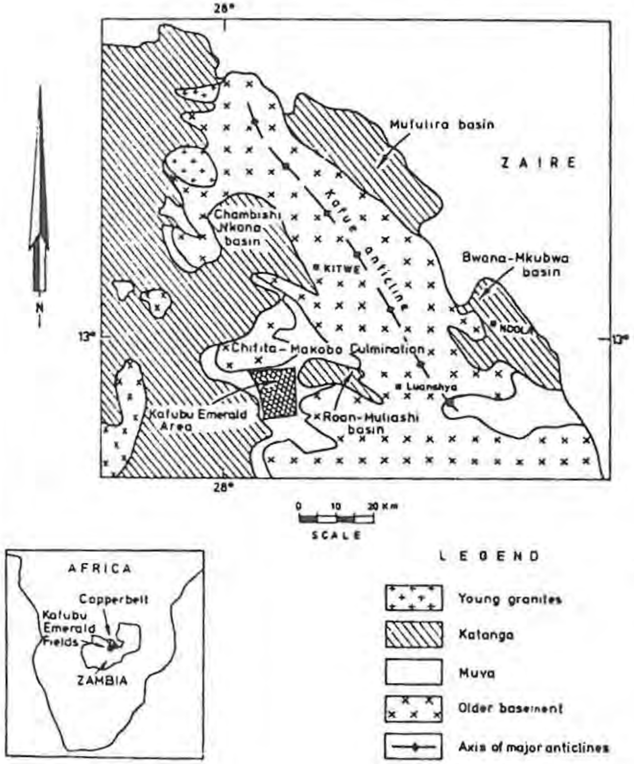

FIGURE 7.9 *Location and regional geological setting of the Kafubu Emerald Fields of Zambia (after Siliwa and Nguluwe, 1984)*

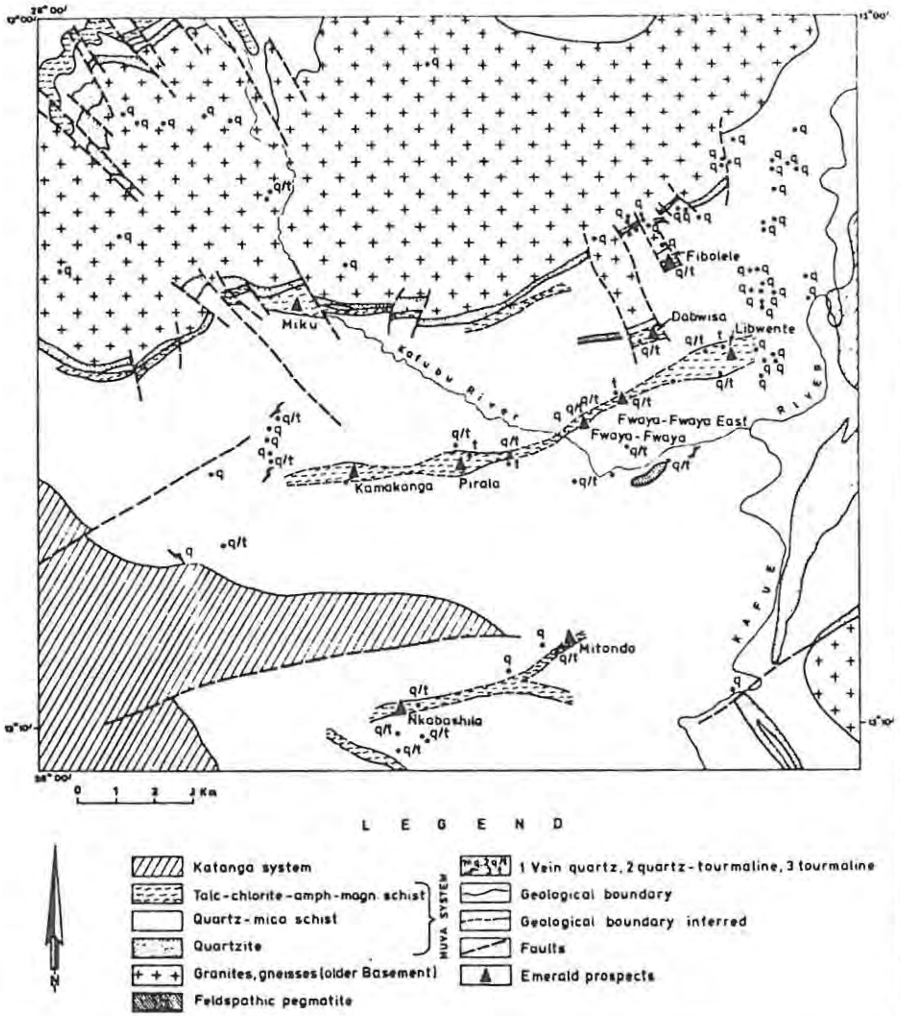

FIGURE 7.10 *Geological map of the Kafubu area (after Siliwa and Nguluwe, 1984)*

7.4.5 Li-Be-Ta-Nb Deposits of Alto Ligonha, Mozambique

Economically important bodies of pegmatites are situated in a zone 200km by 50km (Fig. 7.11), and in the Lurio and Nampula groups forming part of the basement types described above (Sections 7.1 and 7.4.1). The rare elements Li, Cs, Be, REE, Sc, Y, Th and Nb-Ta are predominant in the economically important pegmatites which occur in zones of up to 1km around equigranular granite intrusions. Regional zonation is shown by a northern zone of Be, Nb and Li mineral assemblages, and a southern zone of beryl and columbite-tantalite (Hutchinson and Claus, 1956).

The pegmatites display a wide range of strikes and dips and surface topographic expression. Boundaries with wall rocks are generally sharp and discordant. Alteration of country rocks is shown by deep weathering near pegmatite contacts. One pegmatite at Marige has an irregular and gradational boundary, and contains partly digested country rock xenoliths.

Zonation is well developed (Table 7.2). Border zones are a few cm to 0.7m in thickness and consist of an assemblage of sugary textured plagioclase-quartz-muscovite-black tourmaline-garnet rock. Wall zones are the thickest as they reach up to 7.5m at Muiane (Fig. 7.12). They are dominated by plagioclase with quartz, muscovite, and, in places, biotite. Perthite is sporadic and may form large euhedral crystals. It is often also graphic.

There are five main assemblages which form the intermediate zone. The first is the "book-mica-assemblage", which is a coarser and inner mode of the wall zone. Present in all pegmatites is the second "perthite assemblage". It is a very coarse-grained unit lying just inside the book mica assemblage if the latter has developed. In the next "quartz-plumose-muscovite-assemblage" plagioclase replaces quartz. The fourth, is the main "lithium-assemblage" and consists of plagioclase, spodumene, quartz and minor amounts of amblygonite and lepidolite. The plagioclase is often partly kaolinised cleavelandite. The innermost intermediate zone constitutes the fifth "lepidolite $\pm$ albite-assemblage" and is composed of of the two minerals in extremely variable proportions.

TABLE 7.2 Zone Types in Pegmatites of Alto Ligonha, Mozambique *(Hutchinson and Claus, 1956)*

	Muiane	Naipa	Marige	Nanro	Nahora	Macula	Nahia	Nihiri	Pitteia	Murapane	Nacasupa
(1) plagioclase-quartz-muscovite-tourmaline-garnet	X	X	X	X	X	X				X	
(3) plagioclase-quartz-muscovite: with or without perthite with or without biotite	X	X	X	X	X	X	X	X	X	X	X
(3a) plagioclase-quartz-book muscovite		X								X	
(3b) perthite with very minor quartz and muscovite	X	X	X	X	X	X	X	X	X	X	X
(3c) quartz-muscovite	X	X	X		X						
(5) plagioclase-quartz-spodumene: with or without amblygonite with or without lepidolite	X	X		X	X						
(8) lepidolite with or without cleavelandite and vice versa	X	X	X		X		X				
(11) quartz	X	X	X	X	X	X	X	X	X	X	X

The core comprises massive white quartz as a ridge or hill. Vugs 1 to 4m are known. Boundaries are often sharp but gradational contacts may occur. The distribution of economic minerals is such that columbite is concentrated in the perthite assemblage, with lesser amounts in the wall zone and the quartz-plumose assemblage. Beryl occurs mainly in the perthite and quartz assemblages just outside the quartz cores. The important lithium mineral is lepidolite which occurs in the inner intermediate zone. Small, but very rich pockets of native bismuth are found with quartz in the perthite unit at Marige, and near the quartz core at Naipa. Muscovite deposits are rare.

FIGURE 7.11 *Location of the Alto Ligonha pegmatite district in Mozambique*

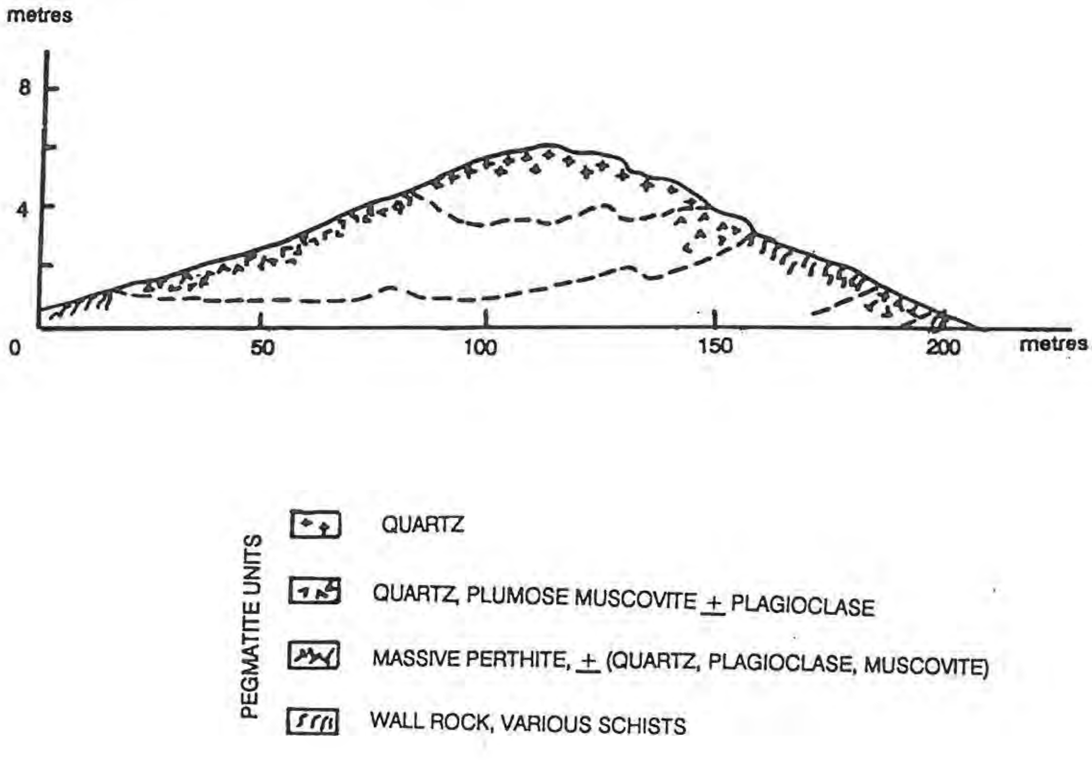

FIGURE 7.12 *Cross-section of the Muiane Pegmatite of the Alto Ligonha District in Mozambique (after Hutchinson and Claus, 1956)*

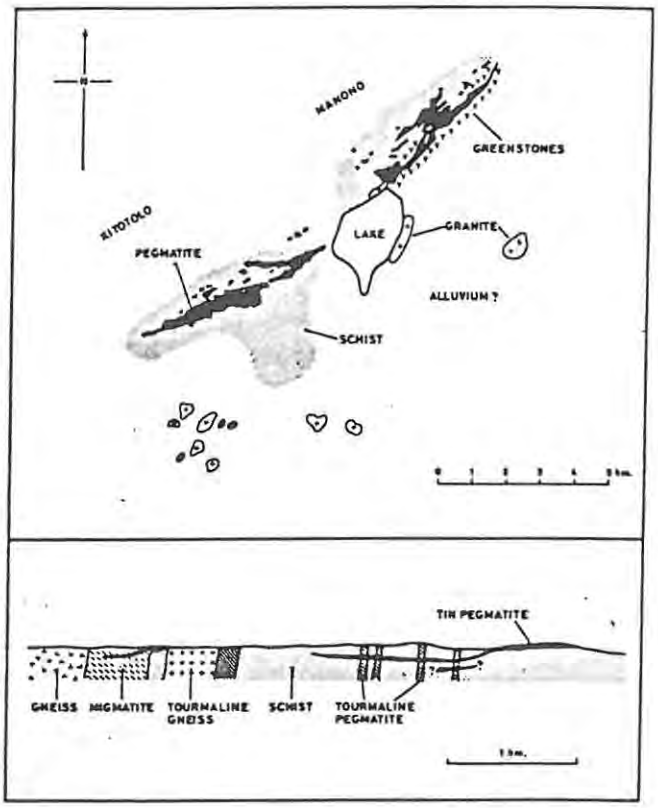

FIGURE 7.13 *The Manono and Kitololo pegmatite fields, Katanga Subprovince, Shaba, Zaire (after Pelletier, 1965)*

7.4.6 The Sn Deposits of Katanga, Zaire

In the Katanga subprovince, there are some post-tectonic, leucocratic and muscovite bearing granites that are Sn bearing. A variety of mineralised quartz veins and pegmatites are associated with these granites and constitute the extensive Katanga Tin Province. Perhaps one of the most spectacular Sn-Nb-Ta pegmatite is the Manono of northern Katanga (Fig. 7.13).

The combined length of two neighbouring pegmatites, the Manono and Kitololo is 14km while the joint width averages 700m. The two pegmatites occur *en echelon* and are concordant with the foliation. Zoning, though present, is complicated by the presence of numerous xenoliths of country rocks. Spodumene is a common mineral associated with quartz and albitic units. It is replaced by greenish, finely-laminated micaceous pseudomorphs known as killinite.

Cassiterite enrichment is associated with albitic zones and pockets of greissen along contacts, otherwise it is evenly distributed throughout the pegmatites as fine crystals. Accessory minerals include coloured tourmaline, apatite, beryl, some lepidolite and flourite. Beryl and amblygonite are scarce or absent.

7.4.7 Sn-W-Quartz Deposit at Rutongo, Rwanda

The Sn-W deposits of Rwanda are a southwesterly continuation of similar mineralised pegmatites of south-western Uganda (Von Knorring, 1970). The Rutongo deposit occurs in the east limb of the an anticlinorium intruded by one of the host/parent G4 granites. On outcrop, the granite is sericitised and strongly kaolinised. Siliceous rocks that intruded were silicified and tourmalinised, while metapelitic lithologies are sericitised, kaolinised and bleached. A wide halo of tourmaline also follows bedding-planes.

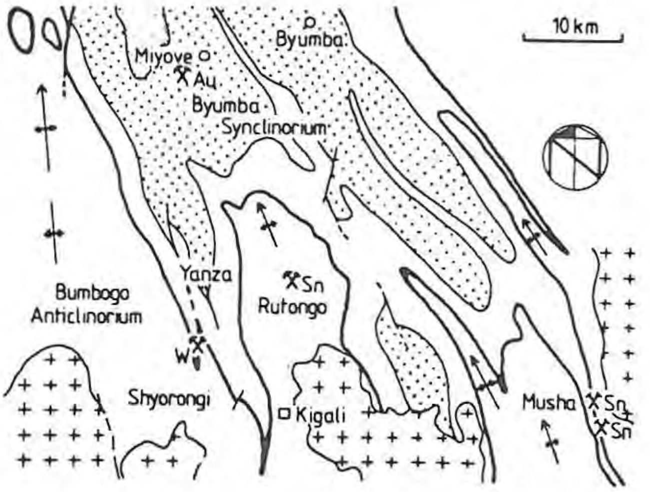

FIGURE 7.14 Geological map of showing position of mineral deposits north of Kigali, Rwanda:
Crosses - tin granite bodies (G4); White with black band - Lower group of Kibaran sediments; Coarse stippling - lower group, Fine stippling - upper group (after Pohl and Gunther, 1991)

Tectonic control is also apparent since the intrusions occupy zones of cross-folding and thrusting (Fig. 7.14 & 7.14). Lithological control is shown by the common occurrence of mineralised veins in country-rock sandstones as opposed to the interbedded pelites. Single veins range between a few cm to 1m, but are mostly around 10cm in width. They consist of quartz and cassiterite with minor quantities of muscovite, rutile, tourmaline, sulphides and secondary quartz (Pohl and Gunther, 1991).

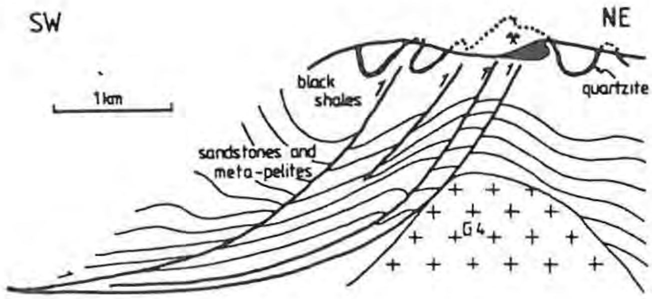

FIGURE 7.15 *Schematic section showing tectonic setting of the Nyakabingo tungsten depositand underlying G4 granite cupola (after Pohl and Gunther, 1991; N. B. key as in Fig. 7.14)*

Fluid inclusion studies of the cassiterite vein deposits (Pohl and Gunther, 1991) suggest several invasions by aqueous phases. The first was of high fCO_2 with moderate salinity, and was intruded at 2kb. The second, which was intruded at 200b or 2km depth had a low fCO_2, higher salinity and carbonaceous phases. These are interpreted as having formed during initial retrograde boiling of the parent granite, and as the fluids entered fractures just produced. The second fluid phase may represent efficient boiling in regions where gas escaped from the system, hence the low fugacity. Its higher salinity is attributed to leaching of apical parts of the parent granite. The presence of two fluid phases trapped under different conditions is a common characteristic of Sn deposits (Eugster, 1985; Haapala and Kinunen, 1982). A third fluid phase is even more saline and is also associated with cassiterite. It is suggested that this resulted from contamination of the magmatic fluids by meteoric waters (Pohl and Gunther, 1991).

The Sn is transported in the fluid as a chloride complex, and is precipitated due to lowering of pressure, temperature and pH, as might be expected when boiling occurs as shown below

$SnCl^+ + 2H_2O \dashrightarrow SnO_2 + 2H^+ + Cl^- + H_2$ (Eugster, 1985).

H^+-metasomatism results and explains the kaolinisation and silicification associated with the deposits. Au, As, CO_2, CH_4 and N_2 appear to have been derived from intruded sediments and was concentrated in the pegmatite fractionation process, while graphite may have been produced by oxidation of methane from the country rocks (Frost, 1979).

7.4.8 W-Quartz Deposits at Shorongi, Rwanda

Also in Rwanda at Shorongi is the Nyakabingo ferberite tungsten-quartz vein mine (Pohl and Gunther, 1991). This occurs in the eastern limb of a complex dome in the Bumbogo anticlinorium. It appears to be associated to a G4 granite which lies a few kilometres to the south. Alteration diminishes in the same direction. Granitic materials have been altered to kaolin, muscovite, quartz and K-feldspar. Country-rock shales and sandstones have been intruded by small veins, mostly parallel to bedding. The veins are typically 10 to 30cm, although thicknesses up to 1m are known. They consist of ferberite, reinite, muscovite, very little scheelite, sulphides, kaolin, tourmaline and, rarely, cassiterite.

Fluids from the wolframite veins show a similar thermodynamic history to those at Rutongo (Section 7.4.7). This is reportedly also typical of tungsten deposits (Ivanova and Naumov, 1989). Transportation of W is believed to be controlled by its tendency to partition into Cl- and P-rich solutions, as opposed to F, BO_3^{3-} and CO_3^{2-} fluids. Precipitation occurred due to changes in the chemical environment. This was triggered probably by sudden wider contact with the country rock, which was itself caused by implosion related retrograde boiling (Manning, 1985).

7.5 SYNTHESIS

The late Proterozoic pegmatite deposits of the region display a wide variety of the emplacement-depth-controlled classes (Section 2.3; Table 2.1 and Fig. 2.2). The southern deposits of Namaqualand and Dete-Kamativi are predominantly of deeper the mica-feldspar-beryl facies. This contrasts with the more northern deposits, with the Alto Ligonha pegmatites showing intermediate characteristics while those of Manono-Kitololo and Rwanda show predominance of shallower assemblages. The large scale variation is therefore suggestive of deeper erosion in the south of the region under review. Within a district, individual plutons and associated pegmatites are also emplaced at different levels, so that assemblages typical of shallow facies are found where deep facies assemblages are predominant. This was noted in the central African pegmatite province (Fig. 2.3) but is applicable wherever pegmatite deposits occur (Varlamoff, 1972).

8. PAN-AFRICAN BELTS AND ASSOCIATED MINERALISATION

8.1 GENERAL GEOLOGICAL OUTLINE

Deposition of the sequences affected by Pan-African events span the period from 800 to 600 Ma, while deformation and metamorphism lasted from 700 to 350 Ma. In the south of the continent, the events are best represented by the Damara Belt, with its type area in Namibia. Going northeastwards into Botswana, the Damara is covered by younger strata, but reappears in Zimbabwe and Zambia as the Katangan and Lufilian belts. The continental belts appear to constitute inland branches of the Mozambique Belt, which follows the east coast of Africa, and the West Congo-Gariep-Malmesbury Belt on the west and south coast (Hunter, 1981b; Fig. 3.1).

8.2 THE DAMARA MOBILE BELT

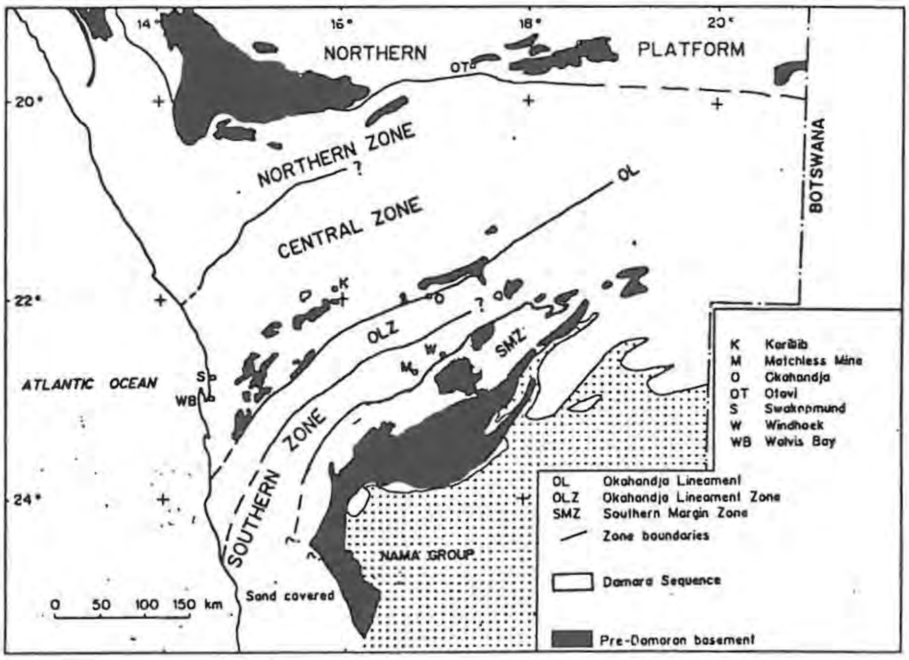

FIGURE 8.1 *The main branch of the Damara Orogen showing the pre-Damara basement and the main lithological, structural and metamorphic zones*

8.2.1 General Geology

The belt (Fig. 3.1) is divided into six zones on the basis of lithological sequence, structure and metamorphic characteristics (Fig. 8.1). Deposition in the Damara orogen began 800 Ma ago, when it occurred in separate depositional basins. Four stratigraphic groups are recognised (Table 8.1). The basal Nosib Group consists of volcanic to volcaniclastic lithologies and sedimentary rocks. Thick clastic wedges of up to 6000m at the base represent a high-energy mass-wasting environment. They were interlayered with volcanic materials of ryolitic to dacitic composition (Mason, 1981).

TABLE 8.1 Lithostratigraphy of the Damara Sequence

NORTH

GROUP	SUBGROUP	FORMATION	LITHOLOGY (MAX. THICKNESS)
MULDEN		OWAMBO	SHALE, MARL, SILTSTONE, SANDSTONE (1000 M)
		KOMBAT	SHALE, DOLOMITE LENSES
		TSCHUDI	QUARTZITE, CONGLOMERATE, ARKOSE, ARGILLITE (3000 M)
UNCONFORMITY IN NW			
OTAVI	TSUMEB	HÜTTENBERG	DOLOMITE WITH CHERT, SHALE, LIMESTONE, STROMATOLITES, OOLITES (900 M)
		ELANDSHOEK	DOLOMITE WITH CHERT, STROMATOLITES (1100 M)
		MAIEBERG	DOLOMITE, LIMESTONE, SLUMP BRECCIA (950 M)
		CHUOS	MIXTITE, DOLOMITE, LIMESTONE, SANDSTONE, ITABIRITE, OOLITE, CHERT (700 M)
LOCAL DISCORDANCE			
	ABENAB	AUROS	DOLOMITE, LIMESTONE, MARL, SHALE (450 M) STROMATOLITES
		GAUSS	DOLOMITE, LIMESTONE, OOLITIC CHERT, SANDSTONE (750 M)
		BERG AUKAS	DOLOMITE, LIMESTONE, STROMATOLITES, ARKOSE, GREYWACKE (525 M)
LOCAL DISCORDANCE			
NOSIB		VARIANTO	MIXTITE, TUFF, ITABIRITE
		ASKEVOLD / NAAUWPOORT	RHYOLITE, TUFF, AGGLOMERATE, ANDESITE, - EPIDOSITE, BOSTONITE (6000 M)
		NABIS	QUARTZITE, ARKOSE, CONGLOMERATE

CENTRE

GROUP	SUBGROUP	FORMATION	LITHOLOGY (MAX. THICKNESS)
SWAKOP	KHOMAS	KUISEB	QUARTZ - BIOTITE SCHIST, BIOTITE - GARNET - CORDIERITE SCHIST, AMPHIBOLE SCHIST, QUARTZITE, MARBLE, CALCSILICATE ROCK (3000 M)
		KARIBIB	MARBLE, BIOTITE SCHIST, QUARTZ SCHIST, CALCSILICATE ROCK (700 M)
		CHUOS	MIXTITE, MARBLE, QUARTZITE (700 M)
LOCAL DISCORDANCE			
	UGAB	RÖSSING	MARBLE, QUARTZITE, CONGLOMERATE, BIOTITE SCHIST, BIOTITE - HORNBLENDE SCHIST, CALCSILICATE ROCK (700 M)
LOCAL DISCORDANCE			
NOSIB		KHAN	CALCSILICATE ROCK, AMPHIBOLE - PYROXENE GNEISS AND QUARTZITE (1100 M)
		ETUSIS	QUARTZITE, ARKOSE, CONGLOMERATE SCHIST, RHYOLITE (3500 M)

SOUTH

GROUP	SUBGROUP	FORMATION	LITHOLOGY (MAX. THICKNESS)
NAMA GROUP PARTIALLY ?			
SWAKOP	KHOMAS	KUISEB	BIOTITE SCHIST, BIOTITE QUARTZITE, GRAPHITIC SCHIST, CALCSILICATE ROCK, AMPHIBOLITE (MATCHLESS MEMBER) (10.000 M)
		AUAS	QUARTZITE, SCHIST, MARBLE, AMPHIBOLITE, ITABIRITE (1800 M)
		CHUOS	PEBBLES SCHIST, MIXTITE, QUARTZITE, SCHIST, ITABIRITE, AMPHIBOLITE, CALCSILICATE ROCK (1650 M)
	KUDIS	BLAUKRANS	GRAPHITE SCHIST, QUARTZITE, QUARTZ - MICA SCHIST, CONGLOMERATE, ITABIRITE (1700 M)
		? HAKOS	QUARTZITE, SCHIST (2000 M)
		CORONA	DOLOMITE, SCHIST, CONGLOMERATE (400 M)
LOCAL DISCORDANCE			
NOSIB		DURUCHAUS	PHYLLITE, QUARTZITE, CONGLOMERATE, LIMESTONE (5000 M)
		KAMTSAS	QUARTZITE, ARKOSE, CONGLOMERATE (6700 M)

(after Porada and Wittig, 1983)

Succeeding dolomitic carbonates dominated the sequence in the north, but graded into a sequence of carbonate and clastic components in the centre and south. Dolomitic carbonate precipitation continued in the north, while the deposition in the centre and south was dominated by aluminous or schists of a basaltic origin, with some siliclastic rocks, carbonates and amphibolites. A distinct Khomas Trough developed between the southern foreland and the Central Zone. It is possible that the basin developed to the extent that ocean spreading occurred for a short time before closure (Miller, 1983c). The Damara sequence is capped by an erosive and clastic sequences.

Granite generation was restricted mainly to the west of the Central Zone and to the north of the Khomas Trough (Fig. 8.1). It occurred from about 750-650 Ma to 450 Ma (Kroner et al., 1978; Clifford, 1967). Syntectonic granites were formed in cores of mantled-gneiss-domes, while post-tectonic granites and associated pegmatitic granites were produced as sheets. Three main types of granite are recognised (Miller, 1983a). The Red Gneissic suite appears to have been produced by partial melting of the Nosib Group and is confined to the western part of the Central Zone. Salem-type granites are granodioritic and syn- to post-tectonic. Their intrusion was initially coeval and eventually outlasted that of the Red Gneissic Granites. They may have been produced from remobilised basement. The third category were post-tectonic and were intruded between 450 to 550 Ma. These were dominantly adamellitic to granitic. The presence of contact aureoles indicates that they were intruded into cold rocks.

Metamorphism is of low grade in the palaeo-shelf areas, where mild flexuring also took place. Near mantled-gneiss-domes, amphibolite and granulite facies metamorphism were reached. Dislocation is intense on the southern margin.

8.2.2 Rare Metal Pegmatite Deposits

Four groups of the Salem-type of granites (Miller, 1983a) are distinguished. These are early syn-tectonic, syn-tectonic, late-syntectonic and post-tectonic (Richards, 1986). Deposits of Sn, Be, Rb, Ta-Nb and Li mineralisation in addition to mica and feldspar are associated with the post-tectonin intrusions.

Stratigraphically, pegmatites with economic mineralisation are limited to parts of the Uis Formation of the Khomas Subgroup. A lithological association with Sn-rich sediments, and nodular and spotted schists or "knotenschiefer" has also been claimed (Richards, 1986). The pegmatites are limited tectonically to graben structures believed to have formed due to intersection of northeast- and southeast-trending normal faults (Fig. 8.2). It is likely that these formed due to updoming in the area during development of the three arms of the Damara Orogen which also caused the widespread anatexis responsible for remobilisation of pre-existing formations (Miller, 1983a). The pegmatites occupy *en echelon* structures produced during subsequent shearing.

Regional zonation in the country rocks is well displayed in the Uis Belt (Fig. 8.3). There is an inner garnet zone in contact with the parent leucogranite, and an albitised, greissenised and mineralised zone outer zone. A tourmaline (schorl) zone often separates the two zones. More rarely, a tantalite-columbite zone may also separate the mineralised and the tourmalinised zones.

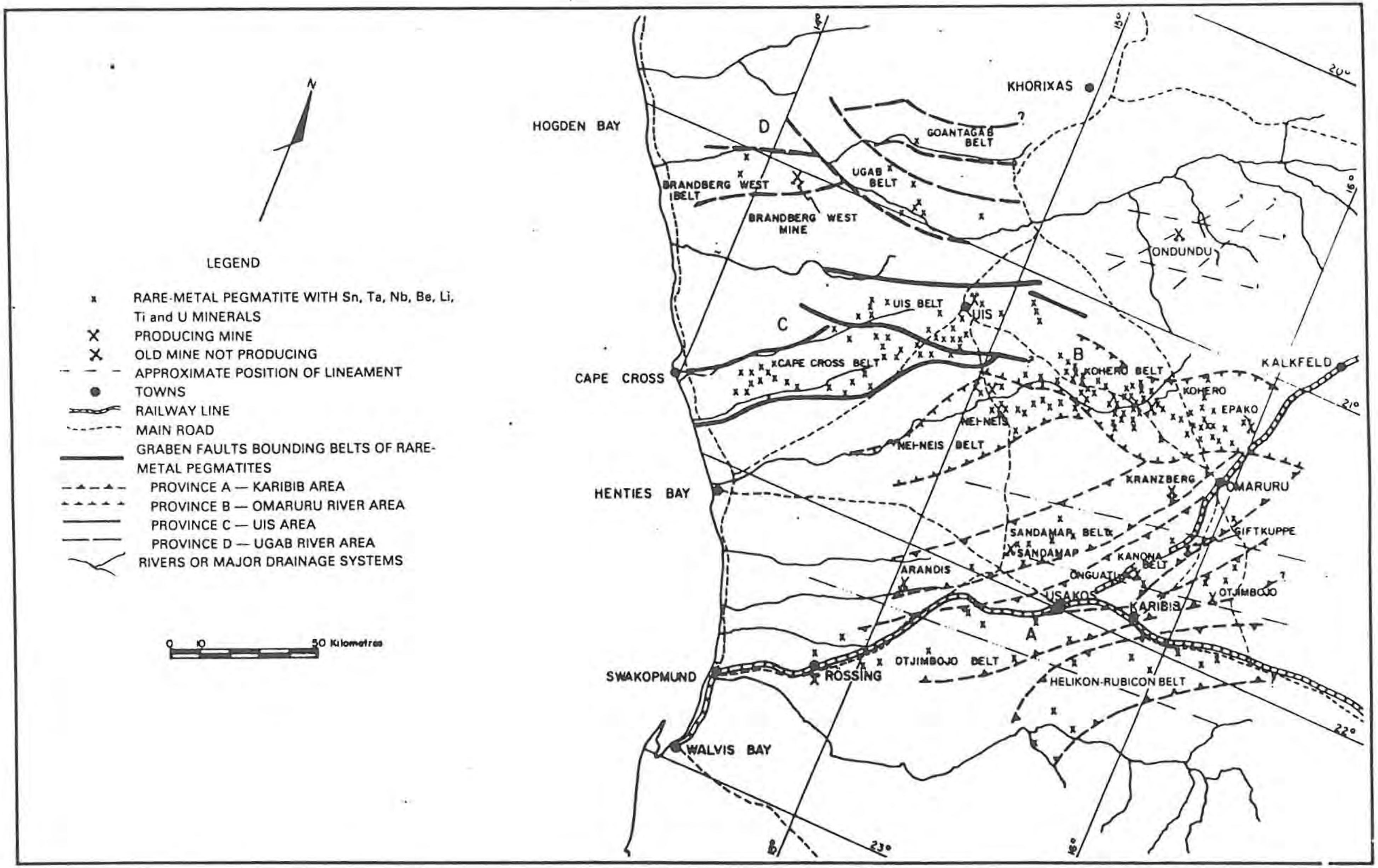

FIGURE 8.2 *Pegmatite fields of the western Damara Orogen, Namibia Orogen (after Richards, 1986)*

Haloes of intensely altered, bleached and fractured country rock extend typically 20 to 50m, but can occur as far as 300m from the contact with the pegmatite (Richards, 1986). Assemblages formed show an increase in tourmaline which peaks at 50m, where it can form up to 60% of the country rock (Fig. 8.3). The alteration zone is tourmaline-free within 5m of the pegmatite. Sericite, and muscovite are most abundant at the margin of the pegmatite, but thereafter follow curves similar to that of tourmaline. Biotite, quartz and feldspar show an antithetic pattern (Fig. 8.3).

In the Cape Cross-Uis belt, four types of pegmatite have been distinguished: (i) cassiterite-bearing types 1 to 3m in thickness, occurring individually and generally generally unzoned, (ii) Nb-Ta-rich types which display typical Li-Na-K mineralisation and albitisation of feldspar (iii) Li-rich, amblygonite-, spodumene- and petalite-bearing types which occur in north-south-trending swarms with cassiterite bearing veins occupying the margins of the belt while they occupy the middle and (iv) a simple category which hosts no mineralisation (Diehl, 1986).

The muscovite-sillimanite and garnet-cordierite/biotite isograds of pelitic assemblages, and muscovite/plagioclase-diopside isograds in calc-silicate rocks place constraints on the physical conditions of the country rocks at the time of emplacement. The temperature was between 500 and 600°C and the pressure was approximately 500M*p*a (Richards, 1986).

Pegmatite-hosted mineral deposits are also found in the West Coast Belt, where they are associated with the Younger Granites (Sohnge, 1950; Hugo, 1986).

8.3 THE KATANGA BELT

In central Africa, sedimentary rocks of the Katangan are preserved in the Central Congo Basin, the West Congo Geosyncline, and in the Zaire-Zambia area before extending southwestwards to Namibia. The basal part has a minimum age of 888-840 Ma, determined using vein mica from beds low in the sequence. Syngenetic Pb gives a date of 1040-1055 Ma, while a maximum possible age is given by the Kibaran Orogeny itself (Cahen and Bartholomew, 1974).

TABLE 8.2. Sequence of Katangan Sediments

Upper Katangan (Kundulungu)	Upper Kundulungu	Calcaire Rose Petit Conglomerat (tillite)
	Lower Kundulungu	Kakontwe limestone Grand Conglomerat (tillite)
Lower Katangan (Roan Mine)	Mwashia (Upper Roan)	
	Upper Roan/Middle Roan	
	Lower Roan	

(After Mendelsohn, 1981)

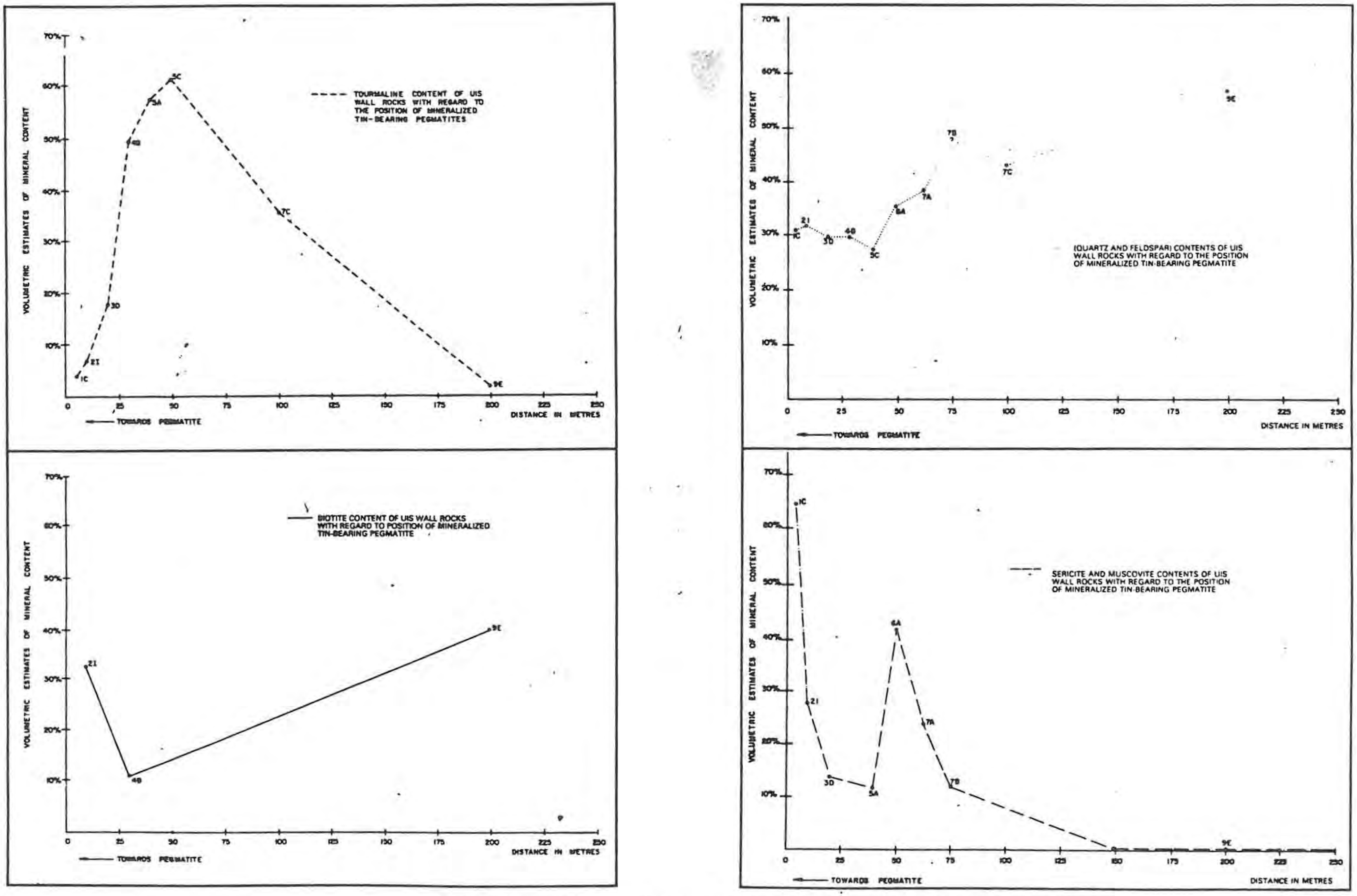

FIGURE 8.3 Zonation around a pegmatite due to alteration in the Damara Orogen, Namibia

(after Richards, 1986)

Table 8.2 shows the lithologies of the type area of the Katangan sequence. The presence of two tillites or tilloids in the Katangan sequence has enabled correlation of the Pan-African deposits across central Africa. These are found in the Liki-Bembe outlier, Ubangi-Lindi (West Nile Complex), the Bushimay and Golfe de May (Mendelsohn, 1981; Fig. 4.10).

Deposition occurred throughout the area, and took place between 1000 Ma and 640 Ma, terminating at the onset of Katangan tectonism. The deposition is believed to have been confined to intracratonic basins, whose outlines had already been marked out by Kibaran sedimentation. The environment was essentially shallow and quiescent. Igneous activity was mild.

Structure in the Central Congo is essentially flat-lying, with faults and folds being of local importance only. The West Congo Geosyncline shows a southeastward trend. In Zambia and Zaire, the Lufilian Arc (Fig. 7.5) was formed. Metamorphism was of a low grade, and seems to have occurred in two episodes; a Katangan proper event at 640 Ma, a Mozambiquan event at 520 Ma which involved higher grade of metamorphism. In places such as the the Lufilian Arc, a hotter and more intense tectono-thermal Mozambique event has overprinted the Katangan structures. The events have been attributed to plate-tectonic processes, with linear jostling of plates after an initial collision (Mendelsohn, 1973).

The Katangan and Mozambiquan events were widespread in the Pangeaic orogenic systems and followed a stable phase between 900 and 700 Ma (Hurley, 1974; Sawkins, 1976).

Significant pegmatite-hosted mineral deposits of the Katangan and rocks of similar ages from the central African subregion are not known. Deposits at present recognised as of upper Proterozoic age were wrongly thought to be of Pan-African (Mozambique) age in some cases in the past, as was the case for the Alto Ligonha pegmatites of Mozambique (Hutchinson and Claus, 1956; Sacchi et al., 1984).

8.4 THE ZAMBEZI BELT IN NORTHEASTERN ZIMBABWE

8.4.1 Geological Outline

The middle to upper Proterozoic Lomagundi and Piriwiri groups (Sections 6.3, 7.4.1 and 7.4.2) are unconformably overlain by the Makuti Group of 825 ± 55 Ma ago, which is correlated across the Zambezi River with the Katangan (Broderick, 1976). The Makuti group is pelitic at the base, while succeeding units are predominantly composed of feldspathic quartzite, calc-silicates, micacious and aluminous schists and amphibolites of igneous origin. The sequence was metamorphosed during the Katangan and the grade of metamorphism increases southwestwards from Makuti (greenschist facies) to Vuti (granulite facies). The Rushinga Group to the north of Mount Darwin and in the Chimanda area (Fig. 4.8) has been correlated with the Makuti Group.

8.4.2 Mica Pegmatites

Mozambiquan (400-650 Ma) granitisation was associated with pegmatite generation and intrusion at . Some of the pegmatites contained economic amounts of beryl (Wiles, 1961). Two types have been recognised: (i) mica-pegmatites associated with sillimanitic and amphibolitic terranes in highly micaceous country-rocks (ii) intrusive pegmatites that are more widespread and are hosts to the beryl deposits (Wiles, 1961).

The mica occurs in wall zones and in large books with unspotted plates which are developed where a quartz core has formed. Beryl crystals which reach up to 6m in length, occur in the core or core-margins of short and lenticular pegmatites. The mineralised cores are associated with microcline or perthite in inner assemblages of intermediate zones.

8.5 THE MOZAMBIQUE BELT

The 700 to 400 Ma Pan-African belt has been previously thought to stretch for 4000 km from south Egypt to Mozambique after passing through Sudan, Uganda, Kenya, Tanzania, Zambia and Malawi. It was also thought to represent a cycle of events from subsidence with deposition, uplift, folding, metamorphism and plutonism. Geological relationships are complex as the Pan-African event was the last stage of a polyphase evolution. The belt is ill-defined in the Mozambique-Malawi area where deposits of this age are not preserved. In east Africa, it represents east-directed subduction. Activity at the time appears to have caused thermal resetting of the Irumide-Kibaran-Lurian sequences of areas of Zambia, Malawi and Mozambique (Shackleton, 1987)

In the Malawi, the basement affected by the Mozambique event is equivalent to the basement to the Bangweulu Block (Section 6.6) while its Irumide cover is represented by the Mchinji and Mafingi Groups (Section 7.1; Fig 4.11). The basement is subdivided into northern and southern subprovinces.

From its high Na, Mg, Ca and Na > K, the southern subprovince appears to represent a metasedimentary succession although the Si content is in many places much less than is usual for such sequences. Prekinematic orthogneisses include meta-dolerites meta-anorthosites and many synkinematic basic and ultrabasic rocks. The southern subprovince underwent granulite facies metamorphism of Kibaran (Irimide) age followed by retrograde metamorphism. Amphibolite metamorphic facies rocks, however, are the norm (Clifford, 1974; Bloomfield, 1968).

The northern subprovince is has undergone a similar history, but has no calcareous metamorphic rocks. Sillimanite bearing cordierite-granulites are common, as are phyllitic zones where cataclasis has affected micacious lithologies.

Structural and other geological relationships are complex since the rocks have been involved in Ubendian, Irumide and Mozambiquan events, with the northern subprovince with the northern subprovince occurring at a triple-point. Three ill-defined Mozambiquan episodes of deformation are recognised. The second was associated with pegmatite generation which have radiometric dates ranging from 470 to 380 Ma.

In the extreme south of Malawi and adjacent parts of Mozambique, the current view is that juxtaposition of granulite and eclogite, and associated amphibolite facies rocks represent assemblages produced during prograde metamorphism, and that secondary overprinting was insignificant (Morel, 1961; Andreoli, 1984; Andreoli, 1990).

Similar findings have been noted in northeast Mozambique. A resulting alternative model proposed for the area involves collision of a Lurio Craton in northeast Mozambique, against the Niassa Craton in northwest Mozambique, with southern Malawi being the island arc complex in between, to produce the Lurio Belt (Andreoli, 1990). As formation of the belt shows Irumide ages, the Mozambiquan event is believed to have caused only rejuvenation.

Pegmatite mineralisation of this age is not reported.

9. GEOLOGICAL SETTINGS AND POTENTIAL FOR PEGMATITE MINERALISATION IN CENTRAL AND SOUTHERN AFRICA

9.1 INTRODUCTION

This section is a summary of the findings of Sections 4 to 8 and reviews the major tectonostratigraphic provinces outlined in the report, followed by a discussion of their prospectivity with respect to pegmatite-hosted mineralisation. Assessment is based mainly on the reported presence or absence of mineralisation, and is therefore essentially historical.

9.2 ARCHAEAN SETTINGS

The Kaapvaal and Zimbabwe cratons have amply indicated the presence of deposits, and that the main concentrations occur in association with the margins of the greenstone belts. There are wide areas on both cratons where farther study may be considered. In addition to the two cratons, the less well understood Angola-Kasai Craton, is also likely to host deposits. The Gabon Shield, the basement to the Western Nile Complex and the Tanzanian Craton may also host similar deposits, but the search in these areas may have been discouraged by the availability of Late Proterozoic and Phanerozoic deposits which may have been easier to find and mine.

9.3 ARCHAEAN TO EARLY PROTEROZOIC MOBILE BELTS

The Limpopo Mobile Belt consists of deep crustal rocks in which pegmatite materials must have been generated but not preserved. Thus mineralisation in the area in rocks coeval with the main thermo-tectonism is extremely unlikely. However, younger granites which intrude the belt should be studied as it is possible that they could have been intruded later, after the belt had been lifted into colder regions where pegmatites could be preserved after formation.

The outcome of this review regarding pegmatite mineralisation in Early to Middle Proterozoic mobile belts is rather discouraging since no deposits of this age have been found. Apart from a possible occurrence in the Dete-Kamativi inlier in Zimbabwe, rocks of this age appear not to contain any pegmatitic mineralisation. This includes suites constituting the Namaqua basement, the older parts of the Damara basement, the western segment of the Zambezi Belt, the Western Nile Complex, the West Congo Geosyncline and the Ruzizian Belt (Sections 6 and 7.3). It should be noted, however, that belts of similar ages and settings on the other continents are mineralised. Thus the belts cannot be dismissed on the basis of age alone, before regional geochemical surveys and assessment of potential parental granites (Sections 10.2, 10.3, 11.2, 11.3). This is encouraged by the reported occurrence of lepidolite mica in granites of this age at Dete-Kamativi (Broderick, 1981).

9.4 LATE PROTEROZOIC BELTS

This was the golden age of pegmatite mineral deposit formation in the subregion, with deep mica-feldspar facies occurring mainly in the southern Namaqua Metamorphic Complex, and the Zambezi Belt , while moderate- to shallow-depth mineralisation occurred farther north in the Katanga subprovince in Zaire, and in Zambia and Rwanda, and in the Lurian Belt of Mozambique (Section 7). Even shallower-level deposits are reported just outside the area of interest in the Katangan rocks of Tanzania (Malisa and Muhongo, 1984). It is the author's view that the deposits reported in the north of Malawi are of this age too. Zones of Namaquan, Irumide, Kibaran and Lurian activity are therefore priority regional targets for exploration.

9.5 PAN-AFRICAN BELTS

Terranes of this age are also attractive since they contain mineralisation in the Damara and the eastern segment of the Zambezi Belt. However, apart from occurrences in the Damara area, they appear to be associated with isolated granitoids in older belts. Since there is widespread development of deposits of this age in nearby parts of western Africa (Kuster, 1990), the belts of this age, especially those in Angola, in the West Congo Geosyncline and West Nile Complex, form suitable targets.

9.6 SYNTHESIS

All the belts may host pegmatite mineral deposits. The highest chances of finding ore bodies are in the belts classified in this study as Late Proterozoic, with those of the southern parts of the area being better endowed in minerals of the deeper facies of emplacement, while those farther north are hosts mainly to mineral deposits typical of shallower depths of emplacement. The exception is the Limpopo Mobile Belt. Attention to this belt should be given if suitable granites of younger age can be identified.

10. SUMMARY OF TECHNIQUES FOR REGIONAL PEGMATITE EXPLORATION

10.1 INTRODUCTION

This section summarises the techniques normally employed during exploration for pegmatite mineral deposits. It assumes that a part of the belt has been selected already, and the aim is to begin limiting the size of the area. Briefly the techniques involve, firstly, the identification of possible areas where parental granites may occur through regional geochemical studies and determination of the relative age of emplacement relative to deformation. Secondly, the occurrence of pegmatite bodies with potential for mineralisation is ascertained. The next phase attempts to identify regional zonation around individual parental granites, whether the granites crop out or not. The final stage of regional exploration involves preliminary examination of individual pegmatites in order to find out if detailed mapping, geophysical sounding or drilling should be conducted. The next stage of evaluation of individual pegmatite bodies will not be entered into except in very brief outline.

10.2 REGIONAL STREAM-SEDIMENT GEOCHEMISTRY

Presuming that part of a belt, probably limited by international boundaries, has been chosen for purposes of exploration, it is necessary to obtain some idea of its mineralisation potential. This is rapidly achieved by reconnaissance techniques such as stream-sediment sampling or float-chasing. Stream-sediment sampling will yield two-fold information. It allows determination and isolation of the mineralised area, and secondly, it can be used for direct isolation of a potentially mineralised target, thereby avoiding examination of similar but barren counterparts (Trueman and Cerny, 1982).

In this approach the explorationist is at an advantage because pegmatite-related heavy minerals and resistates may be present in samples. Commonly, cassiterite and niobium-tantalum minerals will report to the heavy mineral fractions in panning or heavy-liquid separations. Associated with these are the resistates which may report to other size fractions in screening and/or be discarded in the heavy-mineral concentrating procedures. This rejected fraction should be examined visually for tourmaline, beryl or spodumene.

10.3 IDENTIFICATION OF PARENTAL GRANITES

Granite plutons form the fundamental unit which is looked for in pegmatite exploration (Trueman and Cerny, 1982). Suitable ones have been noted in the foregoing review to be generally late- to post-tectonic (Trueman and Cerny, 1982). This was certainly the case for the G4 granites associated with tin deposits of Swaziland (Sections 4.1.2 and 4.1.3; Davies, 1964), and the Mashishimala Granite to which are closely related the various deposits occurring in the northeastern Transvaal in the Kaapvaal Craton (Sections 4.1.4 to 4.1.9; Anhaeusser, 1976, 1991; Robb and Robb, 1986; Sohnge, 1986; Van Eeden et al., 1939). It is also the case for the Chilimanzi Granite Suite of the Zimbabwe Craton (Section 4.2; Martin, 1964). This theme is repeated throughout the development of the central and southern belts as reference to sections (6 to 8) will show. The main distinguishing criterion is their crosscutting contacts with country rocks. Their relatively light colour is also characteristic. These features are often discernible on remote-sensed images including aerial photographs.

The remotely sensed data review is followed up by regional sampling and geochemical analysis. Relative to standard granite (Daly,) the granites, and in particular their cupolas show:

1. lower contents of Ti, Fe, Mg and Ca,
2. high quantities of Na, K, Be and Sc with low K/Rb, K/Cs, Mg/Li, Rb/Sr and variable K/Ba ratios,
3. possibly low Al/Cr, Zr, V and Zr/Hf

(Kuzmenko, 1976 in Trueman and Cerny, 1982).

If a specific type of mineralisation is being sought, comparison of the major and minor element contents, ratios and mineral composition of the granite under study with those of known parent granites may give further clues (Table 10.1). Trace element content and ratios have also been reported to show geochemical signatures of granites parental to Ta deposits. The most important elements for this assessment are Rb, Li and Mg. Rb and Li should not fall below 25 ppm while the Mg/Li ratio should not exceed 100 (Table 10.2; Beus et al., 1968 in Cerny, 1982).

TABLE 10.1 Geochemical Indicators of Some Parental Granites and Their Specific Mineralisation Types

	Li		Rb	Sn	V	Mg/Li		Zr/Sr	
Intrusion Type	< 25	> 100	< 150	> 20	> 4	< 30	> 100	< 30	> 100
Parental to apogranites Ta-bearing	< 1%	64%	< 1%	28%	1%	100%	< 0.1%	83%	< 0.1%
Parental to Ta-bearing pegmatites	< 1	41	5	26	1	39	0.4	93	14
Parental to Ta-poor pegmatites	26	27	24	16	1	27	30	87	5
Parental to Ta-free greisens	26	23	30	21	14	44	12	46	70
Barren granites	54	2	43	1	30	0.1	98	7	48

(after Beus, 1968 in Trueman and Cerny, 1982)

10.4 REGIONAL ZONATION OF PEGMATITES

After a prospective region has been defined, there is often need to limit attention to field zones in which a particular type of pegmatite mineralisation being sought is likely to be found, thereby keeping expenditure low. This is done by utilising the zoned nature of pegmatite mineralisation (Sections 2.6 and 2.7; Fig. 2.2 and 2.3). On the ground they are defined as areas in which

1. the number of <u>internal zones</u> of individual pegmatite bodies is similar;

1. the number of internal zones of individual pegmatite bodies is similar;

2. the numbers and varieties of metasomatic assemblages are the same;

3. certain mineral assemblages occur;

4. the physical forms of selected minerals are the same;

5. the specific compositions (as members of isomorphous series) are the same.

These mineral typomorphisms may be summarised as follows

1. plagioclase: the more sodic varieties are found in more fractionated pegmatites emplaced nearer to the surface;

2. rose quartz: usually found in barren pegmatites with respect to Be-Ta-Nb mineralisation;

3. muscovite: brown or dirty-green varieties are associated with barren bodies while yellow-green-silvery coloured types often imply the presence of Be$\pm$Ta, Nb and spodumene pegmatites;

4. tourmaline: black varieties associated with simple and barren pegmatites although Be, Nb and Ta occur in small quantities; blue and green types in albitised pegmatites suggest Sn-Nb-Ta mineralisation; pink or colourless ones imply Li-Rb-Cs mineralisation;

5. beryl: coarse, columnar greenish yellow or brownish crystals in simple pegmatites imply no rare metal mineralisation; white to pink, stubby and tabular crystals imply rare-metal, Li and F enrichment with Ta, Li and Cs;

6. columbite-tantalite: occurs in moderately fractionated and therefore poorly mineralised pegmatites in association with microlite, tapiolite and Nb-rutile; occurrence with more Ta-Nb mineral species indicates in general more fractionation and higher contents of the Nb-Ta;

7. spodumene: mineralisation high in Li with low Nb-Ta-Be is associated with green, small and columnar crystals occurring throughout the pegmatite bodies; white columnar spodumene near quartz cores indicates diverse rare-metal mineralisation which is generally poor in Li;

8. holmquistite: in intruded country-rocks indicates the presence of Li-pegmatite-

9. biotite and phlogopite: in country rocks indicate albitisation;

10. fluorite: in calcareous country rocks also indicates the presence of Li-pegmatite.

The zonation is controlled by the chemical character of fractionated material and the conditions of emplacement, evolution and consolidation (Trueman and Cerny, 1982). An outer quartz vein zone might also be discernible.

Where texture and mineralogy fail to reveal zonation, it may be of value to conduct systematic sampling of minerals such as K-feldspar and muscovite. K/Rb ratio of K-feldspar and K/Rb, K/Cs and Mg/Li ratios of muscovite show fractionation trends. This purely chemical approach is fraught with difficulties where the relative distances of different samples of pegmatite from their parental granites, which could also be different, is unknown.

11. DISCUSSION: APPLICATIONS TO THE REGION

11.1 INTRODUCTION

An attempt is made in this section to use the geological information available on pegmatite mineral deposits of the region (Sections 3 to 9) in order to develop some proposals as to how the standard techniques for regional exploration just outlined (Section 10) might be utilised in the search for more deposits.

11.2 REGIONAL STREAM-SEDIMENT GEOCHEMICAL SURVEY

This has been conducted in most of the area by the geological surveys of countries of the region over the years. In general attention was focussed on investigating the potential for base metal types of mineralisation as opposed to rare-metal types. It is not realistic to review the surveys in any detail within the limits of this dissertation. However, a review of work carried out in the Namaqualand Metamorphic Complex may illustrate how the existing base-metal exploration data may be collected and utilised when pegmatite-hosted mineralisation is being sought.

In the Namaqualans Metamorphic Complex, various grainsizes of stream-sediment were analysed for FeO, MgO, P_2O_5, MnO, Cr, Ni, Cu, Zn, Nb, Ce, Pb, and Th (Beeson et al., 1975). The aim was to set thresholds, and to determine the sample variability, population distributions and lithological control on stream-sediment composition. In addition the effectiveness of sieved and panned portions as indicators of mineralisation under the climatic and topographic conditions of the region was examined. The survey indicated that for granitic areas, 200-mesh Ce contents varied between 91 and 108 ppm, while Nb and Th values were from 18 to 25 and 16 to 26, respectively. These values could then be used to locate areas underlain by granite of high contents of Nb, Th and REE (Ce), which are also likely to be parental to pegmatite mineralisation (Beeson et al., 1975). Similar methods might be employed in other granitic terranes of the region.

In some areas, only the content of the target base metals and FeO with MnO were analysed for. Correlation of the base and rare metal contents of the Namaqualand Metamorphic Complex might be useful in other parts of the region where the geochemical data has this limitation.

11.3 PARENTAL GRANITES

11.3.1 Introduction

The information available on the granites which are either known or suspected to be parental to adjacent pegmatite deposits in central and southern Africa is summarised. The aim is to identify the general characteristics which can then be utilised during exploration in the region.

11.3.2 Archaean Settings

Within the granitic-gneiss terrane of the Kaapvaal Craton, identification of suitable alkaline granites would require the ability to distinguish late- to post-tectonic granites from earlier ones. These have been noted in the foregoing review to be leucocratic true granites as opposed to the surrounding trondjemites. They are also more uniform in grainsize and have circular outlines as opposed to the older granites which are sheeted. This applies even where the granite follows a pre-existing structure in general, as the intrusive character can be detected using aerial photographs of a suitable scale (Newton and Joubert, 1973). Distinguishing of the granites was clearly possible in mapping of the G4 granite suite in the Swaziland Tin Belt (Davies, 1964).

For granites bordering greenstone belts, the task is made much easier for two reasons. Firstly, the granites intrude the margins of the belts, with which they have a sharp lithological contrast. A better regional exploration strategy for deposits of the Swaziland Tin Belt and Murchison Greenstone Belt type might therefore start with granites of the greenstone belt margin. The tone of the potential granites on aerial photographs or satellite imagery, and the geophysical and geochemical attributes should be determined and used in the more difficult granite-gneiss terrane.

The lack of definite parental granites at Bikita (Robertson, 1981) complicates the approach just outlined in the search for deposits similar to those of the Bikita Pegmatite Field. However, it is generally agreed that granites of the Chilimanzi Suite have the right attributes, and that a similar body hidden at depth may have been the source of the pegmatite mineralisation of the area, as opposed to the closely associated Sesombi Suite granites, which are adamellitic. Texturally, granites of the former suite are porphyritic, although contact zones may have sheared textures. Boundaries with the country rocks are generally sharp and cross-cutting, with poor or no development of a contact aureole (Wilson and Harrison, 1977). In contrast, the Sesombi Suite granites are equigranular and tonalitic. These attributes would be of limited but nevertheless some usefulness in demarcating areas where farther isotopic and geochemical analysis should be done.

The Angola-Kasai Craton poses a more fundamental problem. Too little is known about its geology, mainly due to superficial cover and, until recently, civil strife. Most of the age determinations fall in the early to middle Proterozoic range (Cahen et al., 1984), which, on the basis the study of the adjacent areas, seems to suggest that the terrane is unsuitable for rare-metal mineralisation. The reported mica, beryl, quartz and uranium deposits (Section 4.3; Anhaeusser, 1981) are probably of Damaran age since Kibaran dates are rare (Cahen et al., 1984). This view is also supported by the absence of mineralisation in the coeval and adjacent Kibaran portion of the Damara basement (Mason, 1981). Thus, detailed mapping and the use of such indirect methods as geophysics, geochemistry and remote sensing are necessary before it will be possible to assess the potential, let alone suggest an exploration programme, for the area. In view of the reported presence of mineral deposits (Section 4.3), the need for the grassroots geological study is overdue.

Lack of alkali granites in the Limpopo Mobile Belt precludes the area as a target for pegmatite-hosted mineral deposits (Sections 5 and 9.2).

11.3.3 Early to Middle Proterozoic Belt Settings

Early to Middle Proterozoic granites of the Hudsonian (1800-1600 Ma) in the Canadian Shield have pegmatite mineralisation (Cerny, 1982). However, this review has shown a lack of mineral deposits of this age in the central and southern Africa region. The geochronologic comparison should not be followed too far, especially as sources, modes of mobilisation, concentration and emplacement have not been investigated. Nevertheless, it is likely that the apparent absence in the latter area is related to intense subsequent movements and heating. Zones where late- to post-tectonic potassic granites of the age occur still need to be assessed for their pegmatite mineral potential. That mineralisation cannot be discounted at present is shown by the presence of high-Li mica from one of the granites of this age determined in rocks of the Dete-Kamativi Inlier (Section 6.2; Table 6.1; Rijks and Van der Veen, 1972). Study of the source granite might reveal the nature of the granite types of this age which could be associated with mineralisation. Where similar granites occur in the region, they should be examined for possible rare metal mineralisation.

11.3.4 Late Proterozoic Belt Settings

Parental granites for deposits of the Namaqua Metamorphic Complex have not been determined positively. Difficulties are expected since deep-seated pegmatite mineralisation commonly occurs far from source granitoids. However, the pegmatites are spatially associated with the intrusive components of the Vioolsdrif, Hoogoor and Keimoes suites (Hugo, 1969, 1973, 1986). Starting on this premise, it is recommended that chemical and geophysical characteristics, as well as reflectivity be studied to find out if they can be readily distinguished. The knowledge gained can be used in further assessing the pegmatite potential of granites already determined to be late to post-tectonic.

A method of mapping out of individual granite bodies using aerial photographs has already been successfully used in the area (Newton and Joubert, 1973). This indicates that it is possible to distinguish individual plutons, although some work is still necessary in order to determine the characteristics that would help positive identification of the fertile granites.

Parental granites of the deposits at Alto Ligonha in Mozambique are also late-tectonic. They are undeformed and show a low, ill-defined Sr_i-ratio value, probably suggesting that mixing took place between mantle and crustal material. Additional detail on their chemical and other attributes is not available.

In the Kibaran of Rwanda and Zaire, Sn and W mineralisation is related to granites rich in Li, Cs, Rb, B, F and Sn, which also have high Sr_i-ratios. The granites are post-tectonic and generally pegmatitic and equigranular. They are also subalkaline and strongly peraluminous with a normative corundum content of up to 5%. They are, in this respect, similar to tin granites of Erzgebirge in Germany (Tischendorf, 1974). Their high Sr_i-ratio shows subcrustal to lower crustal derivation (Pohl and Gunther, 1991).

In the Zambian part of the Kibaran Belt, granitoids related to mineralisation may be the syenitic and granitic suite represented by the 1134 Ma Lisenga and Nkamba Bay (Andersen and Unrug, 1984). These appear to be represented by the Lwakwa and Mwenga granites of northern Malawi (Thatcher, 1974), and detailed analysis may show the little-known semi-precious stone deposits of northern Malawi to occur in similar settings to those of Zambia.

11.3.5 Pan-African Belt Settings

In the Damara, proposed parental granites are syn- to late-tectonic and occur as regionally discordant sheets. Their granitic to granodioritic chemical trends distinguish them from the associated Red Gneissic Suite (Jacob, 1973). The Mozambique age granites of the Zambezi Belt in northeast Zimbabwe have not been studied in any detail. However, they show discordant relationships with country rocks whose structure is dominantly Katangan (Wiles, 1976). It should therefore be easy to distinguish the granites.

11.3.6 Summary

Only the timing of emplacement of parental granites with respect to deformation is sufficiently documented to permit preliminary elimination of areas with sterile granites. There is some information on major element chemistry and Sr_i-ratio for a few parental granites. Since identification of potential parental granites is the first stage in the search for more pegmatite deposits, it is important that the trace element, textural, tonal and geophysical characteristics of known parental granites be assessed in addition to the time-tectonic, major element and Sr_i-ratio properties (Section 9).

11.4 REGIONAL ZONATION

11.4.1 Introduction,

Two approaches to regional exploration may be considered. Firstly, there may be variation in the type of mineralisation along a belt. This has already been noted for the Late Proterozoic Namaqua-Irumide-Kibaran-Lurian megabelt, the southern parts of which host deeper facies of mineralisation, while northern parts have shallower forms of mineralisation (Section 7.5). This aspect is not given any serious attention in this review.

Particular attention is given to the second approach in which the zonation considered is that centred on a parental pluton (Fig. 2.3). If mapped out, it will show where the principal economic mineral assemblages are concentrated, and permit early reduction of the area where detailed examination will be necessary (Trueman and Cerny, 1982).

11.4.2 Archaean Settings

Regional zonation is not known in the granitic-gneiss deposits of the Swaziland Tin Belt. It is likely to exist, however, and might be revealed by mapping beginning where mineralisation occurs, and the information obtained can then be utilised in similar areas around potential parental granites. The particular attributes which should be assessed on outcrops of pegmatites or associated country rocks such as numbers of mineral assemblages and zones, etc., have already been outlined (Section 10).

Examples of greenstone terrane deposits whose parental granites are reasonably well known have been cited for the Kaapvaal Craton. Although it was noted that the southern belt is better endowed (Robb and Robb, 1986; Anhaeusser, 1976; sohnge, 1986) attention appears not to have been given towards mapping around the Mashishimala Granite to determine the regional variation of the content of minerals. Such a study along the lines proposed in Section 10.3 is recommended. Its principal benefit would be to indicate where more or better pegmatite deposits may be expected. Regional patterns found can also be utilised in the exploration in the rest of the craton, as well as in the other cratons of the subregion under study.

Regional zonation in the Bikita Field in the Zimbabwe Craton is not reported on (Rijks and Van der Veen, 1972).. It is more likely, however that the presence of abundant reserves has affected the level of interest in determining any regional patterns. In a craton whose geology is dominated by granites, it is desirable to carry out such a study and to extend the results to other granites with potential for pegmatite mineralisation.

11.4.3 Early to Middle Proterozoic Belt Settings

Mineralisation has yet to be found in this setting in the subregion, hence proposals for the study of regional zonation studies would be premature here.

11.4.4 Late Proterozoic Belt Settings

Zonation around proposed parent plutons of the Namaqualand Metamorphic Complex has been demonstrated. An example is centred around a biotite-granodiorite body occurring in the vicinity of Styr-Kraal. Only a few internal pegmatites are rare-element bearing. Around the pluton, and occurring as a marginal suite, larger pegmatites contain allanite, gadolinite, euxinite and monazite. To the south and east, and occurring externally, the intrusion is bordered by larger bodies of inhomogeneous pegmatites with beryl, triplite and columbite-tantalite in addition to monazite, cyrtolite and euxinite. The Witkop pegmatite is an example of this. The zoning was noted in at least five other pegmatites including those at Kakamas, Kenhardt and Bokvasmaak (Hugo, 1969). As it has been demonstrated that mapping of individual plutons is possible in many instances (Newton and Joubert, 1973), it should also be feasible to identity pegmatite zones around an individual fertile pluton, which should greatly assist in homing in on the desired category and its best grades.

In the Alto Ligonha area, regional zonation is shown by a northern zone of Be, Nb and Li mineral assemblages, and a southern zone of beryl and columbite-tantalite (Hutchinson and Claus, 1956). Zonation is not documented for the other Late Proterozoic deposits reviewed. It is likely to be present and its assessment would have applications as outlined in Section 10.

11.4.5 Pan African Belt Settings

Spatial zonation is well displayed in the Damara orogen. It is defined by the types of economic minerals contained in the pegmatites and also as metamorphic assemblages in host rocks close to the body of pegmatite. In the Cape Cross-Uis Belt, economic mineral zonation is shown by the presence of Nb-Ta-rich and Li-Na-K-rich pegmatites in the middle and cassiterite-bearing veins at the edges (Diehl, 1986). Metamorphic zonation is shown by an inner garnet zone, and an albitised, greissenised and Ta-Nb-mineralised zone farther away. Combined with the observation that the pegmatites are commonly associated with Sn-rich sediments, the above criteria form powerful exploration tools for use in the orogen.

11.5 LITHOLOGICAL SETTINGS

11.5.1 Introduction

After identification of parental granites and the areas of interest in the regional zones, the next step is to determine areas where suitable lithological settings commonly known to host mineralisation occur. This is particularly applicable to deposit types in which reaction between pegmatite and intruded country rocks of specific compositions is necessary for mineralisation to occur.

11.5.2 Corundum and Emerald Deposits

Corundum and emerald deposits in the Murchison Belt occur in areas where the parental Mashishimala Granite invade greenstones. The association and likely participation of ultramafic lithologies during the formation of such deposits has also been discussed (Sections 4.1.6 and 4.1.7). Within the region, this association has also been noted at Kafubu in Zambia (without the corundum) (Siliwa and Nguluwe, 1984) and in parts of the Zimbabwean greenstones near Bikita. After locating fertile granites, exploration should therefore be directed at areas where the lithological association: "rare metal pegmatite + ultramafic schist" occurs.

11.5.3 Rare Metal Deposits

In exploration for rare-metal deposits, attention need only be directed at the search for pegmatites whether in schists or granite since this type of mineralisation seems not to depend on participation of chemical components other than those already contained in the pegmatitic magma. This is likely to apply in the case of mica and feldspar deposits as well. Proof of this is the presence of economic concentrations of rare metal, feldspar and mica in various lithologies. They occur in schists (Shangoni, Rwanda), greenstones (Bikita, Zimbabwe), schists and parental intrusive rocks (Orange River Pegmatite Belt), metasediments and later schists (Dete-Kamativi, Zimbabwe and other central African late Proterozoic deposits) or migmatitic basement cover rocks older than the pegmatites (Alto Ligonha, Mozambique). The lack of a strict relationship between pegmatites and country rock lithologies has been stated categorically for the Namaqualand belt (Hugo, 1969).

11.5.4 Relationship to Schist

In some cases, pegmatites have been observed to bear rare-metal mineralisation if they are hosted in schists closely associated with source granites. This is well displayed in the deposits of corundum and beryl in the Kaapvaal Craton (Sections 4.1.5 to 4.1.7), the rare metal deposits of the Alto Ligonha (Section 7.4.4) and the tin and tungsten deposits of Zaire and Rwanda (Sections 7.4.5 to 7.4.7). This is not always the case, as the Namaqualand deposits are known to occur within the suspected parent granites as well (Hugo, 1969; Section 7.2).

It is likely that the rare-metal deposits normally travel farther from parental granites since they are richer in volatle matter, while the mica and feldspar types are very likely to crystallise near or within the source. This simple pattern is complicated if the pegmatites intrude other alkali granites with similar properties to parental granites within the terrane. This gives the impression that the pegmatites are of the internal type (Fig. 2.3).

From the foregoing, field investigations should not be limited to external pegmatites hosted in schists intruded by parental granites, but should extend over the whole area intruded by the granites and their immediate surrounding. However, where corundum or emerald is sought the relationship is important (Sections 4.1.6, 4.1.7 and 7.4.3) and its existence needs to be established early in the exploration programme. Even where the other forms of mineralisation are sought, other features such as local deformation patterns and differences in rock competence (lithology) might favour the intrusion of pegmatite into the schists. In short, no general rule can be set and the local setting of potential mineralisation needs to be assessed at each potential site of mineralisation.

11.6 STRUCTURAL SETTINGS

11.6.1 Introduction

Also within or around the parent granite bodies, mineralised pegmatites are concentrated in favourable structures such as faults, domes or hinge zones of folds. Attenuated fold limbs and shear zones are also preferentially intruded by pegmatites. In exploration, identification of structural controls of pegmatite emplacement and mineralisation will assist in reducing the areas in which detailed examination is necessary. In this section, an attempt is made to identify such structural controls in the deposits of the subregion.

11.6.2 Early Precambrian Settings

As a rule, Archaean craton rare metal deposits occur along margins of greenstone belts where their localisation is probably controlled by deep-reaching marginal faults. This seems to be the case for the Sutherland and Murchison deposits (Section 4.1). The search for new mineral deposits around the numerous remains of greenstone belts associated with fertile granites in the Kaapvaal Craton and in the rest of the granite-greenstone terranes of the region needs to take this into account.

Structures which control the localisation of mineral deposits on the local scale of fields of pegmatites (i.e around parent granites) have not been studied in the Kaapvaal and Zimbabwe cratons. This is also true of individual fractures hosting pegmatite bodies. Their systematic examination is likely to reveal a close association as that observed in younger deposits such as those of the Damara orogen (Diehl, 1986; Richards, 1986) or the deposits of Rwanda (Pohl and Gunther, 1991). Such a study is necessary as it would help by limiting the area in which farther search should be conducted, and indicate the optimum direction for field exploration traverses i.e. at right angles to their trend.

The structural setting of the Bikita pegmatite bodies has not been studied either. However, the crude parallelism of the major dykes (Fig. 4.8) is unlikely to be coincidental. For similar reasons as those outlined for the Kaapvaal Craton deposits, examination of this aspect is necessary.

11.6.3 Late Proterozoic Belt Settings

In the Namaqua Metamorphic Complex, pegmatite emplacement and structure was controlled by the fractures and joints of the associated intrusive rocks (Hugo, 1969). The pegmatites may also show concordant relationships with respect to the grain of the country rock. As a result they may be hook-shaped, crescentic or form other types of curved bodies (Hugo, 1969).

The Dete-Kamativi Inlier pegmatites are emplaced in a domal structure which dips outwards at 10 to 20° (Rijks and Van der Veen, 1972). Bodies of pegmatites are thickest near the domes and grow thinner outwards (Fig. 7.7). It is possible that the location of the thicker parts is related to the relatively reduced pressure experienced in the domal parts. This suggests targetting of exploration in the Dete-Kamativi area, or other areas with similar structures, to the domal structures. The proposal is also valid in the Rwandan deposits where mineralisation occurs in analogous structures (Fig. 7.14 and 7.15; Pohl and Gunther, 1991; Varlamoff, 1972).

The Rwandan deposits just mentioned also illustrate another structural relationship in which mineralisation occurs along fractures such as shear or thrust zones (Fig. 7.15). Exploration can therefore be targetted along fractures within the domal structures.

Structural control is not reported at the Alto Ligonha deposits in Mozambique (Hutchinson and Claus, 1956). This is also the case at the emerald deposits of Zambia, although both areas underwent at least two phases of deformation.

11.6.4 Pan-African Belt Settings

The Cape Cross-Uis pegmatite belt of the Damara Orogen occupies a graben zone (Diehl, 1986). This has also been demonstrated for all the deposits of the area (Fig. 8.2; Richards, 1986). The fractures were utilised by the pegmatites even after the tectonic regime had changed and become compressive with the normal faults of previous graben structures becoming thrust faults and shear zones, thereby resembling those of Rwanda (Pohl and Gunther, 1991).

11.6.5 Summary

Where studies have been conducted, there is a clear relationship between local structure and pegmatite emplacement. The deposits occupy graben structures, faults, shear zones and hinges, domes, or areas where faults cross. This is illustrated for the deposits of Rwanda and the Damara orogen. The relationship is present also at the local scale. This is expected since it is easier for intruding magma to follow zones of crustal weakness or relatively reduced pressure. They would also tend to occupy pre-existing voids. The relationship of pegmatites with local structure therefore needs to be established to farther reduce the number and sizes of target areas for detailed work.

11.7 ALTERATION ASSEMBLAGES

11.7.1 Introduction

The chances of a deposit having formed and of its subsequent preservation and discovery are enhanced if evidence of late stage hyrothermal alteration can be demonstrated. This is because pegmatite mineralisation is essentially a hydrothermal process involving separation of late-stage fluids of felsic magmatic fluids, concentration of contained metals, followed by emplacement and consolidation. The mechanism in which mineralising fluids facilitate metal concentration and attain chemical characteristics which lead to alteration of country rocks have been outlined in Sections 2.12 and 2.13, and were illustrated by the deposits of Rwanda (Pohl and Gunther, 1991; (Sections 7.4.6 and 7.4.7). In this section, examples of alteration within or around mineralised pegmatites reviewed are recalled, and their possible applications to exploration highlighted. Caution has to be exercised where alteration haloes in country rocks are to be used. These may be very narrow or absent even if deposits are present, since pegmatite formation and mineralisation is essentially a closed-system process.

11.7.2 Early Precambrian Settings

Alteration is absent in the Swaziland Tin Belt. As zoning within pegmatites is also poorly developed or non-existent, deposits of this type are difficult to assess quickly without chemical analysis of individual pegmatites. The kaolinised and sericitised parts of the mineralisation are known to host the higher concentrations and grades of younger deposits as is the case in the Damara orogen and the deposits of Rwanda (Richards, 1986; Pohl and Gunther, 1991). It is therefore possible that this portion was eroded and redeposited as the alluvial and eluvial tin deposits which have been worked out (Sections 4.1.2 and 4.1.3; Davies, 1964). Areas where the equivalents of the eroded parts of the belt and associated high-grade mineralisation may still be present may be found if any fertile granites present in the granite-gneiss terrane are identified and follow up studies conducted in the way proposed (Sections 10.1 to 11.4).

In the corundum deposits of the Kaapvaal Craton, alteration is localised. Along the boundaries of pegmatites, amphibolites have been transformed to talcose schists or "amorphite" (Section 4.1.6). A similar setting is reported for the Gravelotte Emerald Mine in which pegmatite fluids seem to have altered ultramafic and mafic rocks of the Murchison Belt to talc-chlorite-biotite schists (Section 4.1.7). Here it is also associated with albitisation (Robb and Robb, 1986; Van Eeden, 1939). Thus pegmatites with an alteration halo of this type suggest the likelihood of mineralisation. Finally, alteration is not considered as an important attribute of the dominantly quartz-feldspar-mica deposits.

11.7.3 Late Proterozoic Belt Settings

Alteration in the country-rocks in the vicinity of individual pegmatites has not been reported in the Orange River Pegmatite Belt, and is assumed to be unimportant since very thorough study has been carried out in the area (Hugo, 1969). It is likely that the high pressure during emplacement at great depth confines fluids to the pegmatite bodies better than at shallower depths. This would severely limit access of fluids to the country rocks and result in the lack of alteration.

Alteration assemblages are poorly developed in the biotite schists which host the pegmatite deposits of the Dete-Kamativi Inlier in Zimbabwe. Near to pegmatite bodies, schist lenses are stopped from the wall rocks and are present as inclusions in the Sn lodes. The lenses tend to have bleached extremities and biotite flakes situated in a light-grey, fine grained mass of sericite. At a more advanced stage, only sericite and quartz remain, indicating that greissenisation did occur. Cassiterite-bearing pegmatites also develop a thin contact zone less than 2 cm in thickness and consisting of muscovite and quartz along the edges of mica schist inclusions (Rijks and Van der Veen, 1972). Local almandine and black tourmaline enrichment occurs in the country rocks in minor and accessory amounts. These might form a halo around mineralised areas as seen in the Damara orogen (Section 11.7.4).

Marginal alteration of schists bordering pegmatites of emerald deposits have been mentioned in connection with the Gravelotte and Bird Cage Camp deposits (Section 11.5.2). A similar situation occurs in the Proterozoic emerald deposits at Kafubu in Zambia (Siliwa and Nguluwe, 1984), and serves to emphasise the common alteration assemblage regardless of the age setting.

The Manono and Kitololo deposits of Zaire (Hutchinson and Claus, 1956; Von Knorring, 1970), and the Rutongo and Shorongi Deposits of Rwanda show a close association with albitisation and greissenisation. The greissenisation and silicification is present also in the associated pelites and psammites (Pohl and Gunther, 1991). Alteration zones may be present in the deposits of the Alto Ligonha area but have been weathered out (Von Knorring, 1972).

11.7.4 Pan-African Belt Settings

In the Damara orogen, alteration of the country rocks has also occurred, and consists of haloes of intensely altered, bleached and fractured rock typically extending 20 to 50m, and in places as far as 300m from the contact with the pegmatite. Assemblages formed show a horizontal increase in tourmaline which peaks at 50m, where it can form up to 60% of the country rock (Fig. 8.3). None occurs within 5m of the pegmatite. Sericite, and muscovite is highest at the margin of the pegmatite, but thereafter simulates the tourmaline curve. Biotite, quartz and feldspar show an antithetic pattern (Fig. 8.3).

The application of the alteration assemblages to exploration in the Damara orogen would involve (i) identification of zones of general alteration and, in these, (ii) determination of the direction in which tourmaline, sericite and muscovite is increasing. This should be accompanied by declining amounts of biotite, quartz and feldspar and in the direction of the pegmatite.

11.7.5 Summary

The process in which rare-metal mineralisation occurs involves highly reactive aqueous fluids, so that alteration assemblages are formed in adjacent rocks. In most cases, it is better if this is limited since injection of large amounts would also cause loss of rare metals to intruded rocks by the forming the pegmatite. In the case of Sn and W mineralisation, the process is in fact necessary for cassiterite and wolframite to precipitate (Manning, 1985). In either type of mineralisation, a field under examination is likely to have some pegmatites which did in fact release fluids to the country rocks. These form useful indicators of the possibility of mineralisation which should be searched for in areas of potential.

12. CONCLUSIONS

Mobile zones of all ages from the Archaean (greenstone belts) to the early Phanerozoic host pegmatite-based mineralisation in the central and southern region of Africa. Main concentrations are associated with the Archaean greenstone belts and all the late Proterozoic belts of the region.

Archaean occurrences include various deposits of the Sutherland and Murchison greenstone belts in the Kaapvaal Craton, northeastern Transvaal, South Africa. They include ore bodies of corundum, beryl, corundum-tantalite, mica, feldspar and tungsten. Dominantly Sn and Li deposits occur at Bikita in Zimbabwe. Here they intrude Bulawayan age greenstone belts, and are associated with late-tectonic granites of the Chilimanzi Suite. In the cratons, areas where the greenstone belts are intruded by late-tectonic alkali and peraluminous granites, form suitable exploration targets for pegmatite-based mineralisation. The tin granites and pegmatites of Swaziland may be similar to the Sutherland and Bikita types, although the associated greenstone belts are not known. In addition to the Kaapvaal and Zimbabwe cratons, areas of Angola may have remnants of cratonic material which, if identified positively, should be assessed for pegmatite deposits.

In the very high-grade terrane of the Limpopo Belt, pegmatite components of the rocks must have been generated, but were driven up by the heat and pressure during granulite facies metamorphism. Potential for mineralisation is limited to mildly deformed and metamorphosed remnants of greenstones which are found in the belt's marginal zones, and the mode of mineralisation is likely to be similar to that outlined for the cratonic regions.

Remnants of the early to middle Proterozoic are poorly studied, due mainly to masking by polycyclic metamorphism and deformation during younger events. The potential in terranes of this age therefore remains unassessed. The presence of lithian mica in early to middle Proterozoic granites in the Kamativi area is an indication of the possible potential of granites of this era. However, it is possible that the granites were remobilised to provide a source for the late Proterozoic and early Phanerozoic deposits of the area.

Late Proterozoic deposits were the most extensive and include all types in terms of economic minerals contained. In the Namaqualand Metamorphic Complex to the southwest, mineral assemblages typical of deep regions of emplacement are predominant. Shallower assemblages are the more common northeastward into the Irumide, Katanga and Lurian belts. This probably indicates regional variation in the depth to which uplift has occurred, or differing levels at which rich pegmatites were emplaced. Syn-, late- to post-tectonic alkali granites of this age probably have the highest potential in the area reviewed.

Apart from the Damara Orogen, mineralisation of Pan-African age is uncommon in the region studied. Detailed studies also show that deposits and host terranes previously considered to be of this age are, in fact, late Proterozoic. Syn-, late to post-orogenic Pan-African alkali granitoids in the Malmesbury and Mozambique true Mozambique belts need to be examined for potential as sources of pegmatite-based mineralisation, especially where late Proterozoic sequences are remobilised, as rare-metals they contained may have been concentrated during the reworking, as is suspected to have occurred for the Damara pegmatites

For all the areas, techniques for preliminary assessment of the mineralisation will involve early identification of possible parent granitoids, followed by mapping out of pegmatitic regional zonation before evaluation of individual pegmatites bodies or fields.

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