GRANITOID RELATED Sn-W MINERALISATION WITH SPECIAL REFERENCE TO SOUTHERN AFRICA, THE VARISCAN BELT IN EUROPE, AND THE MALAY PENINSULA

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FRONTISPIECE: DAMARA TIN PROSPECTORS, OUSIS PEGMATITE FIELD, BRANDBERG NORTH, NAMIBIA

ABSTRACT

A review of the geotectonic settings of granitoids and various tin-tungsten provinces in Europe, Malaysia and southern Africa shows a close spatial and temporal association of mineralisation to S-type ilmenite series granitoids. Granitoids with these affinities are derived from crustal anatexis and are most commonly found in continental collision and different ensialic, intraplate orogenic settings, (e.g. SW England, Malaysia, Namibia) as well as in association with anorogenic magmatism (Nigeria, Brazil, South Africa). Tin-tungsten mineralisation is related to late- to post-tectonic granites, emplaced into areas of substantial tectonic thickening. Crustal anatexis leads to an observable calcalkaline chemical trend, with a source of gabbroic or amphibolite composition through anatexis to; mafic-intermediate enclaves, para-autochthonous anatectic granitoids (tonalite, granodiorite), to intermediate level quartz monzonite, granodiorite, biotite-granite, to late-tectonic highly fractionated muscovite-bearing granites, and high level porphyry intrusions. Mineralisation is spatially related to apical protrusions of the youngest most differentiated granite. Various mineralised environments are recognised, including endogranitic veins, primary disseminations, pegmatites and pipes, and exogranitic stockwork and fissure veins, and replacement bodies. A common factor to all these deposits is the inherent greisen environment, characterised by post-magmatic metasomatic alteration and mineral deposition. Common alteration mineral assemblages include albite, quartz, muscovite, tourmaline, and fluorite $\pm$ topaz. Ore mineral assemblages commonly display a paragenetic sequence of oxides (cassiterite, wolframite, scheelite), followed by sulphides (molybdenite, pyrite, pyrrhotite, chalcopyrite, sphalerite, arsenopyrite/loëllingite, Pb-Bi(Ag) sulphosalts) and then lower temperature carbonates (calcite, siderite, ankerite).

Analysis of Pan African orogenic provinces in southern Africa (Damara and Saldanian Provinces) shows there is good potential for applying integrated exploration techniques in search of endo-exogreisen Sn-W systems. Careful analysis and interpretation of granitoid geochemistry (K_2O, Na_2O, FeO/Fe_2O_3, F, B, Sn, W, Mo, Cu, Rb, Sr, Ti, Zr) should aid delineation of Sn-W and Mo-Cu metallogenic provinces in these regions. Magnetic susceptibility determinations should also aid distinction of S-type ilmenite series (less than 1×10^{-4}emu/g) from I-type magnetite series (more than 1×10^{-4}emu/g) granitoids.

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1.0 INTRODUCTION

The aim of this dissertation is to present an overview and comparison of the geotectonic settings and mineralised environments of several tin-tungsten provinces. Aspects of related granitoid genesis, typology and emplacement are discussed, with emphasis on their geotectonic setting. Case studies of more notable European deposits related to the Variscan orogeny (e.g. Cornubian, Iberian Peninsula), and the Malay Peninsula Sn-W belt are presented as a basis for comparison with known and potentially mineralised regions in southern Africa. The latter comparison emphasises the Pan African mobile belts (Damara, Saldania Provinces) as opposed to the anorogenic Bushveld Complex. Implications for tin and tungsten exploration in southern Africa are assessed.

Data and ideas presented are largely from literature research, although a substantial input pertaining to the southern Africa setting is from the writers own observations and discussions with other geoscientists during the 1984 MSc Exploration course.

2.0 ASPECTS OF THE GENESIS, EMPLACEMENT, METALLOGENY AND GEOTECTONIC SETTING OF GRANITES

This section is but a brief overview of a broad and complex topic. The principal aim is to elucidate some details and classifications of granites relative to their geotectonic setting, as well as aspects of Sn and W geochemistry relative to particular mineralising environments. All these factors form a basis for later comment and discussion.

2.1 Granite typology and classification

There are many different types of granite, whose different origins, may broadly reflect different tectonic settings. A great range of mineralogical and geochemical parameters are now available on which to base a reasonably precise granite typology (Chappell and White, 1974; Hine et al., 1978; Takahashi et al., 1980; Ishihara, 1981; Didier et al., 1982). Granites can be formed by a variety of processes, including fractional melting, fractional crystallisation and even metasomatism (Pitcher, 1983), and different source rocks may be involved. The granites are often marked by subtle compositional differences, which reflect the diverse evolutionary paths. Furthermore the environments of granite generation vary, and may be related to variable orogenic/anorogenic settings. A broad categorisation by Pitcher (1979a) of orogenies into Alpinotype, Andinotype and Hercynotype (Fig. 2.1) serves as a basis to discuss the genetic connection between granite type and tectonic environment. Tables 1A, B broadly categorise granite types,

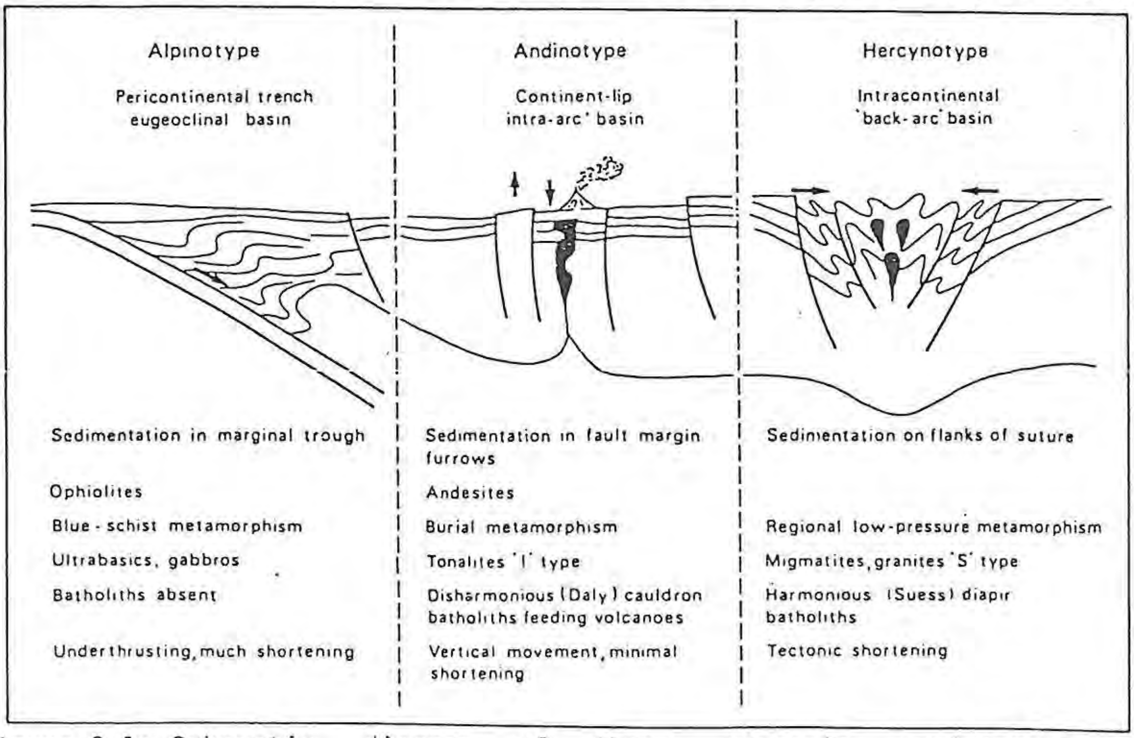


Figure 2.1: Schematic diagram of three orogenic environments to illustrate the differing environments of granitic rocks (Pitcher, 1979b)

Table 1A: Granite types, characteristics and geological environments (Pitcher, 1983)

M-type	I-type (Cordilleran)	I-type (Caledonian)	S-type	A-type
Plagiogranite subordinate to gabbro	Tonalite dominant but broad compositional spectrum — diorite to monzogranite — with wide SiO_2-range. Major association with gabbro	Granodiorite-granite in *contrasted* association with minor bodies of hornblende diorite and gabbro	Granites with high but narrow range of SiO_2. Leucocratic monzogranites predominate but granitoids with high biotite content locally important. Autometasomatic variants	Biotite granite in evolving series with alkalic granite and syenite. Highly contrasted acid-basic relationship
Hornblende and biotite; pyroxene, magnetite	Hornblende and biotite; magnetite, sphene	Biotite predominates; ilmenite and magnetite	Muscovite and red biotite; ilmenite, monazite, garnet, cordierite	Green biotite. Alkali amphiboles and pyroxenes in alkalic types; astrophyllite
K-feldspar interstitial micrographic	K-feldspar interstitial and xenomorphic, pink in hand specimen	K-feldspar generally interstitial and invasive, pink in hand specimen	K-feldspar often as megacrysts with protracted history, white in hand specimen	Perthites
Basic igneous xenoliths	Dioritic xenoliths, may represent restitic material	Mixed xenolith populations	Metasedimentary xenoliths predominant	Cognate xenoliths, also basic magma blebs
Mol $Al/\left(Na+K+\frac{Ca}{2}\right) < 1.0$	$Al/\left(Na+K+\frac{Ca}{2}\right) < 1.05$	$Al/\left(Na+K+\frac{Ca}{2}\right)_{ca}\ 1$	$Al/\left(Na+K+\frac{Ca}{2}\right) > 1.05$	Often peralkaline, relatively rich in F
Initial $^{87}Sr/^{86}Sr$ ratio ≤ 0.704	< 0.706	> 0.705 < 0.709	> 0.708	Considerable range 0.703–0.712
Small, quartz diorite-gabbro composite plutons	Great multiple, linear batholiths with arrays of composite cauldrons	Dispersed, isolated complexes of multiple plutons and sheets	Multiple batholiths, plutons and sheets, less voluminous and more commonly diapiric than I-types	Multiple, centred, cauldron-complexes of relatively small volume
Associated island-arc volcanism	Associated with great volumes andesite and dacite	Sometimes associated basalt — andesite lava 'plateaux'	Can be associated with cordierite-bearing lavas but characteristically lacking in voluminous volcanic equivalents	Associated with caldera-centred alkalic lavas
Short, sustained plutonism	Very long-duration episodic plutonism	Short, sustained plutonism; post-kinematic	Sustained plutonism of moderate duration; syn- and post-kinematic	Short-lived plutonism
Oceanic island-arc of Fijian-type	Andinotype marginal continental arc (and some island-arcs, e.g. New Guinea)	Caledonian-type post-closure uplift	Hercynotype continental oblique collision. Also encratonic ductile shear-belts	Post-orogenic or anorogenic situations
Open folding: burial-type metamorphism	Vertical movements, little lateral shortening; burial-type metamorphism	Dip-slip and strike-slip faulting; retrograde metamorphism	Much shortening; low-pressure metamorphism in slate belts; associated with a *Granite Series*	Doming and rifting
Porphyry-Cu, Au, mineralization	Porphyry-Cu, Mo, mineralization	Rarely strongly mineralized	Sn and W-greisen and vein-type mineralization	Columbite, cassiterite and fluorite

Table 1B: Comparison of I-type/Magnetite-series and S-type/Ilmenite-series granitoids

I-type or magnetite-series granites Mantle source	S-type or ilmenite series crustal source
Tend to be the acid end of a broad compositional spectrum from basic to acid	Tend to occur in restricted ranges of only acidic compositions
High Na_2O contents:- over 3.2% in felsic varieties, over 2.2% in mafic varieties	Rel. low sodium contents: less than 3.2% Na_2O with rocks $\pm$5km K_2)
Mol. $Al_2O_3/Na_2O+K_2O+Ca_2O$ less than 1.1. Low initial $^{87}Sr/^{86}Sr$ ratios (less than 0.708). Normal range of Delta^{18}O values (approximately 6-10%, SMOW)	Mol. $Al_2O_3/Na_2O+K_2O+Ca_2O$ over 1.1 High intial $^{87}Sr/^{86}Sr$ ratios (over 0.708) Enriched in ^{18}O (greater or equal to 10%, SMOW)
C.I.P.W. normative diopside or less than 1% normative corundum	C.I.P.W. normative corundum over 1%
$\pm$ Linear variation diagrams	Irregular variation diagrams
Magmas with relatively high fO_2, relatively high Fe^{+3}/Fe^{+2} ratios (esp. in biotite), bulk Fe_2O_3/FeO over 0.5, magnetite present	low fO_2, low Fe^{+3}/Fe^{+2} ratios, bulk Fe_2O_3/FeO less than 0.5, ilmenite present
Hornblende and sphene common	Muscovite, monazite, cordierite and garnet common
Magnetite 0.1-0.2% by volume, with accessory ilm, hm, py.	Ilmenite less than 0.1% by volume, with accessory po, graphite, muscovite.
Depletion in lithophile elements	Enrichment in lithophile elements.

their characters and geological environments. Obviously such categorisations are a simplification, especially as each mobile belt (e.g. such as the Variscan and Damaran) is always, to a certain degree, unique. Generally in ocean-continent plate-margin orogens, which involve external, volcanogenic troughs, floored by oceanic crust and undergoing high pressure - low temperature metamorphism, granite magmas are rarely developed. On the other hand those involving internal volcanogenic troughs, overlying continental crust and undergoing non- deformative depth metamorphism, support vast multiple batholiths consisting of a broadly compositional series of I-type magnetite series granitoids (Table 1A, B) emplaced over a long period. These granitoids are typical of volcano-plutonic arcs. They differ signficantly from those generated within intraplate orogens, which include continental-plate collision and intracontinental rift settings, and support low P-low to high-T metamorphism, as well as generating smaller batholiths over lesser time intervals. These batholiths are compositionally dominated by S-type ilmenite series granitoids (Table 1B). The contrast between I-type magnetite and S-type ilmenite series granitoids is further enhanced by general differences in the metalliferous ore association-porphyry Cu or Sn-W greisens and vein-type respectively.

Both these types of magmas may experience fractionation leading to more evolved compositions, and both may be contaminated before, during or after fractionation. Thus a wide range of compositions can be effected, and discrimination between any given granites must be approached with caution (Beckinsale, 1979). None of the distinctions in Table 1A can be taken as absolutely rigid, although they certainly are helpful for applying criteria constraining possible source regions (mantle vs. continental crust), and likely environment and/or style of mineralisation.

In essence M-type granitoids (Table 1A) include the plagiogranite of oceanic-island arcs and which grades into an I-type (Cordilleran) representing the voluminous gabbro-quartz diorite-tonalite assemblage of active continental margins. I-type magnetite series (Cordilleran) is separate from an I-type (Caledonian) representing granodiorite and granite of the immediately post-orogenic uplift regimes. S-type ilmenite series granitoids are sharply contrasting, incorporating the peraluminous granite assemblage of encratonic and continental-collision fold belts, and an A-type (Loiselle and Wones, 1979) which includes the alkalic granites of both stabilised fold belts and the domal swells and rifts of cratonic terranes.

Many of the contrasts between I-type magnetite series and S-type ilmenite series are a consequence of different mechanisms of crustal deformation, because these determine the duration of the processes, the degree of participation of mantle-derived basic rocks, and even the nature of the lower crust (Pitcher, 1979b). Aspects of the genesis of these magmas are discussed below.

2.2 Genesis of granitic magmas

The typology and classification of granites outlined above is largely conditional to the actual mechanism of magma genesis. Classic debates have considered granites to be derived by

1) differentiation or contamination of basaltic magmas,
2) partial fusion of continental crust, or
3) metasomatic transformation of the crust by aqueous emanations or solid state diffusion.

Source materials were either mantle peridotite for basalt (and perhaps the aqueous emanations) and the base of metamorphosed volcano-sedimentary piles in the continental crust. A third potential source of silicic magmas (the subducted oceanic crust) was introduced with the advent of convergent plate boundary tectonics, in particular accounting for the eruption of andesites and the intrusion of calc-alkaline granitoid batholiths in island arc settings.

Each of the three magma sources, mantle peridotite, continental crust, and subducted oceanic crust, has a distinct geochemistry (Wyllie, 1983b). This has led to many of the above classifications of granites, including those proposed by Chappell and White (1974), White and Chappell (1977), (I- and S-type granites), and Ishihara (1977, 1981) (magnetite- and ilmenite-series granitoids) (Table 1B).

These two "series" of granitoids are considered (see Ishihara, 1981) to have resulted from the prevalence of different oxygen fugacities during the evoluton of granitic magmas. Disassociation of water in the hydrous magmas is a main oxidising agent for I-type magnetite series magmas, whilst incorporation of crustal carbon is the most essential reducing media for S-type ilmenite series magmas.

Wyllie (1983a, b) assesses the generation of granites in the subduction setting. He concludes that granitoid intrusions, especially those dominated by tonalitic and dioritic compositions, include mantle and crustal contributions. Importantly most granitic liquids are undersaturated in H_2O and cannot be derived from primary melts from either mantle peridotite or subducted oceanic crust because these compositions cannot yield liquids rich in SiO_2. Recent geochemical investigations (Atherton and Tarney, 1979; Green, 1980; Kay, 1980) all favour complex multi-stage models and also advocate a mixed mantle-crust origin for many granitic rocks. The calc-alkaline batholiths and eruptive rocks in island arc settings are thus seen as the end products of deep-seated, complex, multi-stage processes (Fig. 2.2).

In accounting for this, Wyllie (1983a, b) invokes a model whereby regional metamorphism, migmatites and granites are produced by massive influxes of heat carried largely by fluids, but partly by magmas, rising from the mantle. This applies especially to subduction zones where wet crustal wedges are dehydrated and partly melted. The flux of fluids, especially H_2O and CO_2, efficiently carries heat upwards and the combination of an enhanced heat flow with increased water promotes partial melting (anatexis) of the crust to produce granite magma. H_2O undersaturated basic magmas are produced by partial melting of mantle peridotite above dehydrating subduction zones, and the basic magma rises up into the crust at high temperatures, hybridising with crustally produced magma to give typical batholithic complexes composed of tonalite to granite (I-type magnetite series) with variable crustal and mantle derived components.

These concepts can equally be applied to the generation of granites beneath continents. If anhydrous mafic magma, derived from the mantle, is emplaced into the lower crustal rocks of continental regions, large quantities of H_2O-bearing melts will form due to equilibration between the magma and crust, leading to anatexis and assimilation. The melts produced range from dioritic to granitic, depending on the nature of the source rocks and their H_2O content.

The H_2O content of the magmas is an extremely critical facet of granite genesis, and is considered to have direct effects on the mode of emplacement as well as on the metallogenic association. In this respect

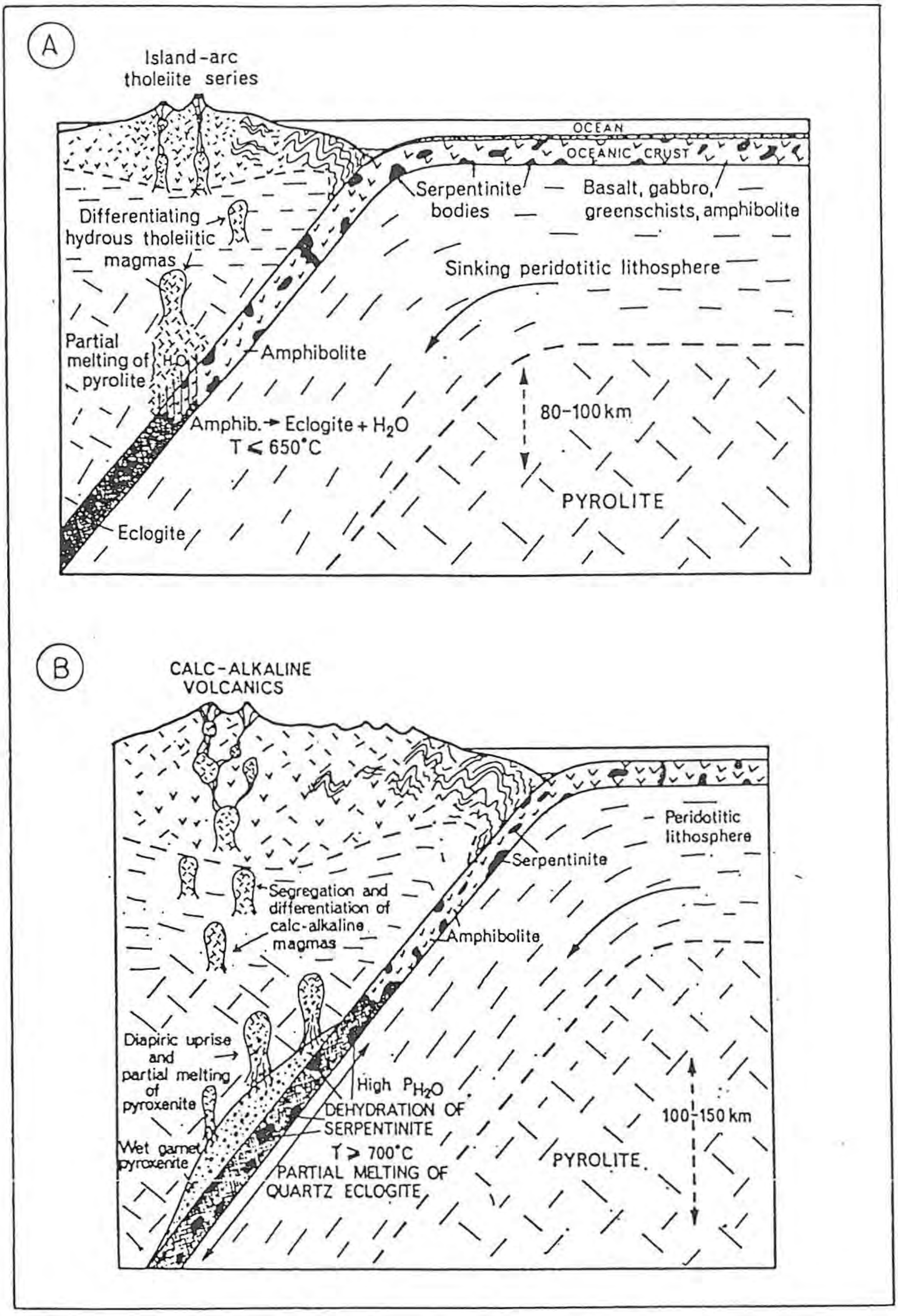


Figure 2.2: Early and late stages in the petrological evolution of an island arc. (a) Involves dehydration of subducted amphibolite, introduction of water into the overlying wedge and generation of island arc tholeiite igneous series. (b) Involves partial melting of subducted oceanic crust and reaction of liquids with mantle above Benioff zone leading to diapiric uprise and formation of calcalkaline magmas (after Ringwood, 1974)

Strong (1981) envisages two environments for mineral deposits of granitic affinity

a) where there are large volumes of H_2O at deep levels, with primary muscovite-bearing, silica-rich minimum melts which foster Sn-W mineralisation, and

b) where H_2O- poor, intermediate composition magmas yield amphibole-biotite bearing subvolcanic intrusions with porphyry style mineralisation (Fig. 2.3).

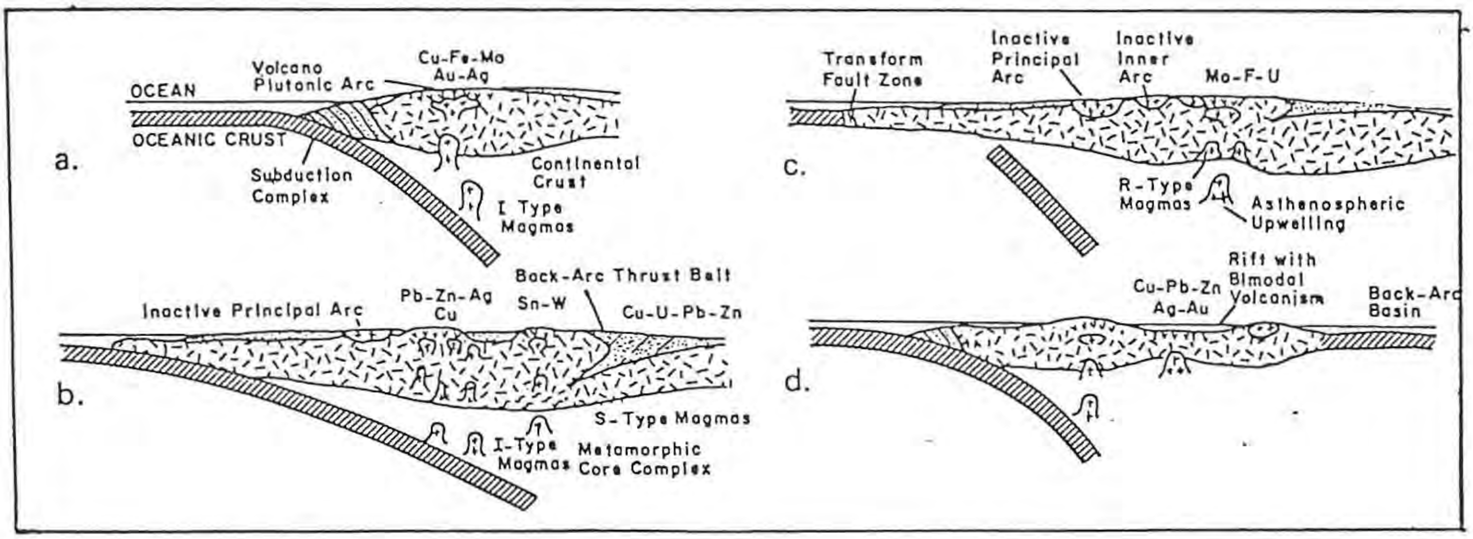


Figure 2.3: Schematized relationships between styles of subduction and arc metallogeny. (a) Moderate to steep subduction with emplacement of Cu-Fe-Mo-Au-Ag deposits in principal arc; neutral stress regime. (b) Shallow subduction with emplacement of Pb-Zn-Ag-Cu deposits and Sn-W-U in inner arc belts; compressional regime. (c) Rapid steepening of detached, freely sinking slab due to transition from subduction to transform margin, with emplacement of lithophile element-type deposits; extensional regime. (d) Steep subduction and commencement of intra-arc rifting with emplacement of Kuroko-type deposits in a submarine setting; extensional regime. (from Sillitoe, 1981a; in Sawkins, 1984)

Both Strong (op. cit.) and Hyndman (1981) advocate the control of granite melts by the breakdown of hydrous phases such as muscovite, biotite and amphibole at successively higher temperatures. Highest H_2O availability would be for muscovite, and lowest for amphibole, and as a consequence advanced melting at higher temperatures will produce less silicic melts or magmas. These are hornblende or biotite bearing, (constituting I-type magnetite series) as opposed to the former which would be muscovite bearing (S-type ilmenite series). I-type magnetite series magmas are capable of rising and intruding to higher levels in the crust before they reach their solidus curve and crystallise. Hence they are capable of venting at surface to produce volcanic-plutonic complexes. These high-level subvolcanic plutons commonly are host to porphyry-type mineralisation. By contrast, in S-type ilmenite series granitoids the presence of primary muscovite limits the crystallisation conditions of the water-saturated magmas, and leads to crystallisation before the melts rise to lower pressure regimes (Hyndman, 1981). These muscovite bearing leucogranites commonly host Sn-W deposits, and are more typical of deeper level metamorphic terranes, commonly with associated alteration and mineralisation assemblages inherent to the greisen environment (see section 2.4).

2.3 Emplacement mechanisms

It is widely accepted that most granite batholiths consist of multiphase plutons intruded during an extended episodic thermal regime. Some of the differences in various settings in terms of volcanic association, composition and genesis have been outlined above, and can be carried to include the mode of emplacement. The nature of the crust may determine whether the magmas fill predominantly cauldron or diapir batholiths, the former related to tensional regimes such as at continental margin/subduction zones, the latter in compressive, ductile regimes such as within intracontinental orogens.

There are fundamental differences in these concepts, including the actual fluid chemistry of the magmas. Magmas filling cauldrons, especially the relatively 'dry' magmas of destructive plate margins, are rapidly emplaced in narrow conduits, creating a narrow heat plume within a cool upper crust. These hot magmas are derived at least in part from the mantle, and their sufficiently low viscosity allows differentiation within chambers, producting volcanic derivatives (Pitcher, 1979a; Strong, 1981).

Diapiric magmas occur within the broad heat plume of a warm, ductile crust, forming globular bodies which diapirically expand within their ductile envelopes. These magmas are significant for their 'wet' nature, i.e. H_2O over 2.0% (Burnham, 1979). Initially the diapirs ('globules' of magma) ascend in a bell-shaped form, with constriction of the base and inflation of the upper parts, forming a mushroom shape. Further ascent of the magma causes the ductile envelope to close in around the tail of the ascending diapir. At high levels the ductility of the crust decreases and the magma stiffens due to cooling, shouldering aside or updoming the adjacent country rock (Holder, 1978). This 'piercement' mechanism may culminate in the pluton punching through the crust where there is little overburden (Pitcher, 1979a). Such forceful intrusion is characterised by smoothly concordant contacts, upward drag of the surrounding envelope and radial and concentric fracture patterns in the roof zone.

Alternatively, in some high-level plutons, the magma may remain mobile, and concordant diapirs then evolve towards the angular discordant cauldrons as stoping becomes the dominant means of uprise (Holder, 1978; Pitcher, 1979b). Deformational and thermal recrystallisation effects in the country rock are not pronounced.

Deeper-seated plutonism, in which the ductility contrast between the pluton and envelope is small, may result in the piercement mechanism being modified by the mechanism of radial distension, in which the diapir expands outwards in pulses, like a balloon, due to the input of new magma into its core. In these cases there is generally a lack of evidence for upward drag, and thermal metamorphic minerals (e.g. chlorite, cordierite, andalusite) may show considerable flattening, imparting a schistosity to the periphery of the pluton and the envelope.

The reader is referred to Bateman (1984) for a recent discussion of aspects concerning diapirism and granite magmas.

2.4 Aspects of Sn and W geochemistry in mineralising environments

Both Sn and W are incompatible elements that concentrate in highly fractionated melts, and, due to partitioning effects, may become preferentially enriched in hydrothermal solutions. This section briefly considers aspects of the geochemistry and transport processes of these elements, and the consequent mineralising environment that they are most commonly found.

Tin

The geochemical specialisation of Sn-bearing granites (i.e. their enrichment in Sn, F, Li, B etc) is due to enrichment in residual liquids thought to be controlled by the oxidation state of Sn in the melt in respect to conditions of high or low oxygen fugacity (Lehmann, 1982). In I-type magnetite series magmas Sn can be tetravalent (Sn^{+4}), and in the early crystallisation stage of differentiaion is capable of substituting for Ti^{+4} or Fe^{+3} in femic minerals such as sphene, magnetite, ilmenite, epidote-group minerals and hornblende (Ishihara et al., 1979; Ishihara, 1981). This substitution limits the availability of Sn to concentrate in residual fluids. In S-type ilmenite series magmas, on the other hand, Sn may be divalent (Sn^{2+}), and has no suitable substitution site in the rock-forming minerals. Thus Sn is more readily available for accumulation in residual melts. Consequently, if an original magma is specialised in Sn and transport media (e.g. F, Cl, see below), tin deposits can be formed in the apical parts of the granitic magma. The presence of S-type ilmenite series granitoids is thus considered a prerequisite in major tin fields (Ishihara et al., 1979).

The means and conditions of Sn transport in hydrothermal fluids are varied and complex. Available partition coefficients (Holland, 1972; Burnham, 1967, 1979) indicate that early formed hydrothermal fluids tend to be enriched in Cl^- or CO_2. The relative proportions of CO_2 and Cl^- present in the fluid phase depends on their initial concentrations in the melt, and on the confining pressure. Fluid phases formed at high pressure are likely to be CO_2-rich and halogen-poor, while later and/or lower pressure fluids would be halogen-rich and CO_2-poor (Burnham, 1979). Consequently complexes involved in the transport of metals such as Sn (and W) in such contrasting fluids may well be different.

Cassiterite (SnO_2), due to its greater stability, is the most common Sn-bearing mineral in hydrothermal mineral deposits. This mineral is commonly associated with fluorite and/or tourmaline, which is largely a consequence of the nature of the Sn-bearing solutions and the mode of Sn transport in these solutions. A variety of mechanisms have been suggested for the transport of Sn in hydrothermal solutions (mostly under highly alkaline conditions and assuming the valency state of tin to be Sn^{+4}, e.g. Klintsova and Barsukov, 1973; Klintsova et al., 1975). Patterson et al. (1981) consider that in a reduced sulphur field and at elevated temperatures over geologically reasonable pH ranges, divalent tin complexes are the likely form of tin transport. Sn^{2+} readily complexes with F^-, OH^- and Cl^-, and the role of these ligands is probably a critical aspect of tin transport.

Sn concentrations due to hydroxyl complexing $Sn(OH)^o$, $Sn(OH)^-$, $Sn(OH)^+$ are extremely low at all but very low fO_2 conditions or at extremes of pH, and therefore hydroxyl complexes seem unlikely to make a significant contribution to hydrothermal Sn transport. Concentrations of Sn due to F complexing ($Sn\ F^+$)are also low except at low pH and/or high fluoride concentrations, although higher fluoride complexes may be more significant, analogous to chloride complexes. F appears likely to be involved in Sn transport in only very saline solutions. Sn is considered to be readily transported as a stannous chloride complex ($SnCl^-$) in conditions adequate for the formation of Sn mineralisation (low pH, low fO_2, Patterson et al., 1981). Similar mechanisms would precipitate cassiterite and accompanying sulphides, namely an increase in fO_2, an increase in pH, a decrease in temperature, or some combination of these. Sn concentrations would be increased at higher chloride concentrations, lower pH, and/or higher temperature (Patterson et al., 1981).

Conditions causing precipitation of cassiterite (above) are very similar to those for wolframite (see below), and it appears that in a Sn/W bearing solution cassiterite (and wolframite) would precipitate earlier in slightly more acidic conditions with scheelite later at slightly higher (neutral-alkaline) pH levels.

Tungsten

Tungsten is a transition metal element with complex chemistry, capable of encompassig a wide range of oxidation states. In nature however it is found almost exclusively in a hexavalent (W^{+6}) state within oxygen compounds, and otherwise as the sulphide species tungstenite (WS_2). Two principal groups of oxotungstate compounds are known as minerals, typified by wolframite (Fe, Mn)(WO_4) and scheelite ($CaWO_4$). Other less common minerals, which form a solid solution series with wolframite, are ferberite ($FeWO_4$) and hübnerite ($MnWO_4$) (Evans, 1974).

In a crystallising magma that is specialised in tungsten, the tungsten is concentrated in H_2O-rich residual solutions (Krauskopf, 1970). At subcritical temperatures (less than 500°C) in solutions containing silica, transportation may be provided by W-bearing heteropolyacids (e.g. $H_8Si(W_2O_7)_6$, Ivanova and Khodakovsky, 1968).

Experimental data (Foster, 1977) indicate that it is possible to mobilise W in chloride- bearing hydrothermal solutions in sufficient quantity to allow later deposition in economic concentrations. Investigation of scheelite solubility in dilute KCl solutions (0.5M, 1M) buffered with the solid phase assemblage K-feldspar-quartz-muscovite showed that solubilities increase exponentially from approximately 60ppm at 300°C to more than 1000ppm at 550°C (Foster, 1977). Molecular hexahalides (WCl_6) probably persist at near magmatic temperatures, but after partitioning molecular H_2WO_4 may become the dominant W species

(Foster, 1977). At sub-critical temperatures the major species are ionic, controlled largely by pH (Fig. 2.4). Four main parameters are postulated to control the transport and deposition of tungsten, namely temperature, fO_2/fS_2, $a(Ca^{2+})/(a(Fe^{2+}) + a(Mn^{2+}))$, and pH (Foster et al., 1978). Scheelite deposition at supercritical (over 500°C, 1000 bars) and near-critical temperatures is a function of temperature variation, and aqueous activity of the ratio $a_{Ca^{2+}}/a_{Fe^{2+}}$. Assuming an initial Cl-rich brine, slightly alkaline solutions (i.e. $a_{Ca^{2+}}$ greater than $a_{Fe^{2+}}$) favour the deposition of scheelite. However, at high fO_2 levels and where $a_{Fe^{2+}}$ is greater than $a_{Ca^{2+}}$, wolframite will be deposited, while at lower fO_2 levels Fe-sulphides may form (Foster, 1977). Higgins (1980) however suggests, from fluid inclusion evidence, that at high pressures fluids are dominated by CO_2 with very low Cl^- values and that W is transported in carbonate/bicarbonate complexes.

The role of F in the hydrothermal transport of W is difficult to assess (Foster, 1977). At lower temperatures (about 300°C) oxyfluoride complexes may increase W solubilities, especially where the concentration of F^- is exceptionally high (Ivanova and Khodakovsky, 1968). Studies of acidity - salinity diagrams in terms of mineral equilibria in greisen (see below) and porphyry deposits indicates that scheelite is formed under <u>low</u> HF and KF activities, while wolframite occupies a field of <u>higher</u> HF and KF values (Burt, 1981). This explains the common association of fluorite (CaF_2) and topaz ($Al_2SiO_4(F, OH)_2$ with wolframite (and cassiterite) occurrences. Burt (op. cit.) also shows that as the F content of a system increases the anorthite content of plagioclase and other calcic phases (e.g. sphene) decrease, or become unstable, making Ca available to form fluorite. In some systems, where F appears low, this Ca is sometimes taken up to form scheelite (Pirajno, in press). Westra and Keith (1981) further illustrated this apparent disassociation between F and scheelite, noting that scheelite is common (but without cassiterite) in calcic magma series, which have low F values, whereas in calc-alkaline magmas wolframite or hübnerite occur together with Sn and Mo, and F is enriched.

As will be seen from later sections, significant Sn and W mineralisation is commonly related to transport mechanisms and partitioning effects that are inherent to greisen environments. It is pertinent here to lay a basis for further discussion by briefly reviewing aspects of this environment.

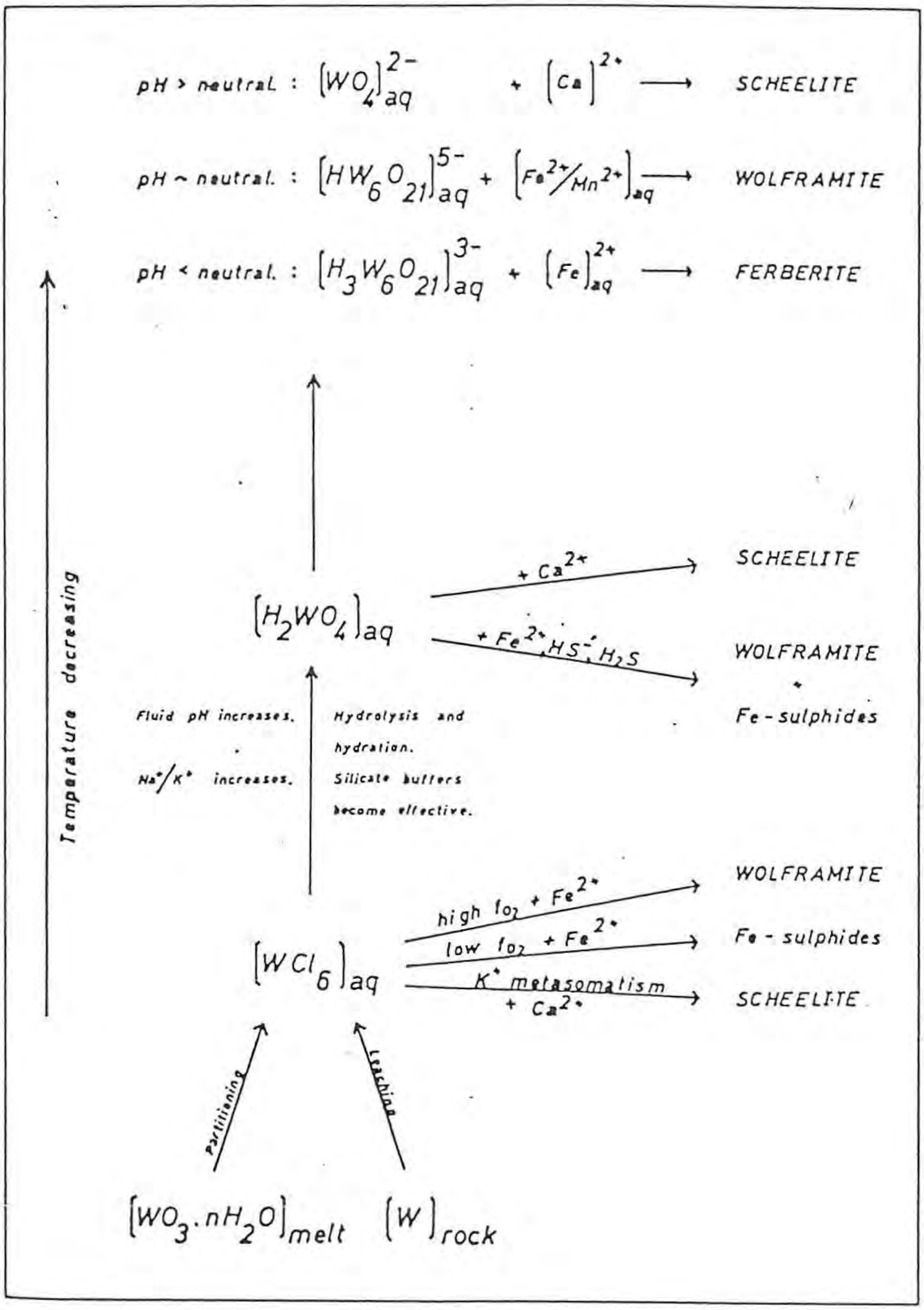


Figure 2.4: Tungsten species and crystallisation of tungsten minerals (Foster et al., 1978).

2.4.1 The greisen environment

Greisen rocks are formed as a result of post-magmatic metasomatic and hydrothermal activity on a crystallised granitic body (endogreisen) and adjacent country rocks (exogreisen) (Shcherba, 1970; Pollard, 1982; Pirajno, in press). Generally greisenisation affects high level granitic stocks or sheets emplaced in favourable structures within argillaceous-arenaceous rock sequences. The variety of mineralised settings discussed here can be assessed in terms of the probable depth of emplacement of the granitoid intrusions (Fig. 2.5), coupled with the potential interaction of mineralised fluids with favourable lithological horizons (Figs. 2.6, 2.7, 2.8). It has already been emphasised that Sn-W mineralisation is commonly spatially and temporally related to S-type ilmenite series granites. A fractional crystallisation sequence from hornblende granodiorite through biotite to biotite-muscovite granite is advocated for the Sn-rich granitoids and which occurs in Europe, Malaysia (Main Range) and southern Africa.

The Brandberg North area (Damara Province) appears distinct from others discussed in that granite cupolas associated with mineralisation have formed conspicuous circular structures, and within which, after probable polyphase diapiric emplacement and subsequent collapse, late stage Sn-W bearing fluids have migrated upwards along planes of weakness. Granitoid terranes in Europe, Malaysia and South Africa don't appear to have documented circular features, but as they occur eleswhere in the world (e.g. New Zealand) it is feasible that they are an inherent feature to there regions either not readily discernable, not described, or since eroded away.

Greisenisation (qtz-musc-tourmaline $\pm$ F, Li-bearing minerals) and oxide and sulphide mineralisation are localised in the apical portions of cupolas where the most intense metasomatic activity occurs, effects of which decrease with depth (Taylor, 1979). The emplacement of such cupolas (endogreisens) invariably is accompanied by veining, of limited vertical extent, depending on the nature of pre-existing fractures. The vein systems are often superimposed sheeted systems (common to all the deposits described here) due to the combination of radiating fractures from intrusion and subsequent collapse. Endogreisen systems appear to be emplaced at shallow levels (2-6km, Fig. 2.5), where epizonal emplacements have associated vein systems, or else may evolve into

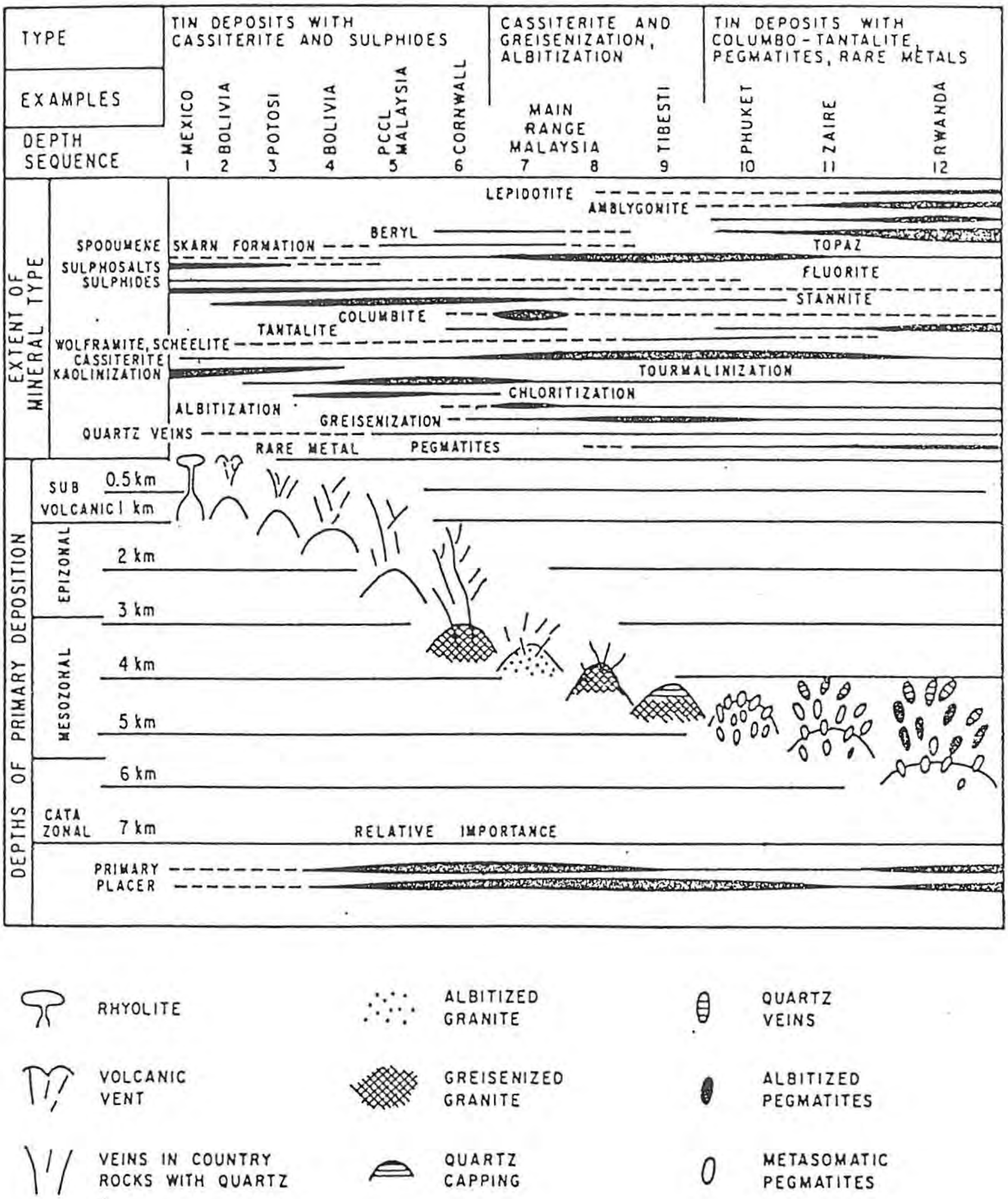


Figure 2.5: A classification of tin deposits according to depth of emplacement of associated granitoids (after Varlamoff, 1978), and also showing the associated mineral associations and types of alteration (Hutchison, 1983)

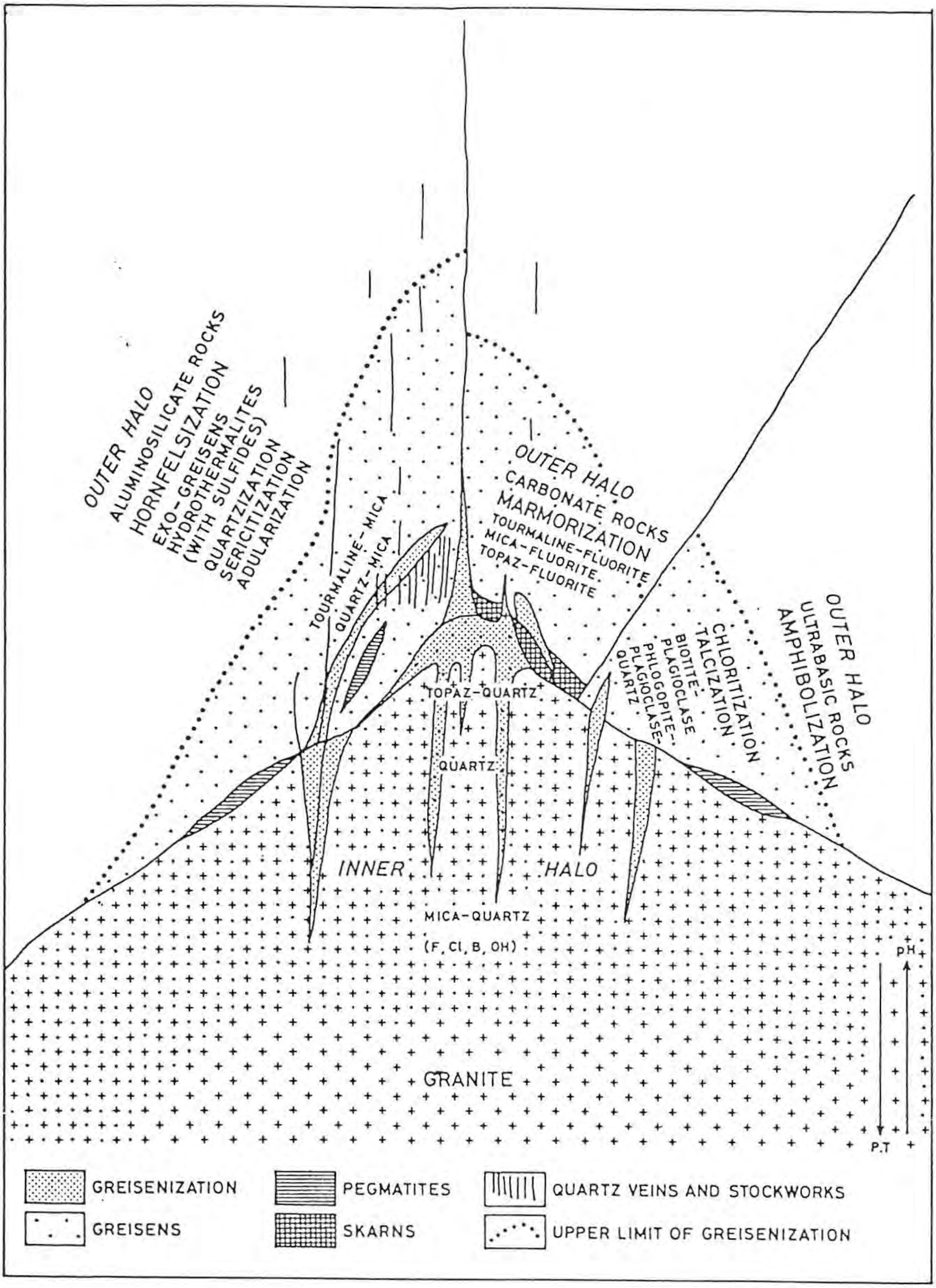


Figure 2.6: The greisen environment, illustrating summary concepts concerning greisen tin-tungsten systems in silicate, carbonate, and ultra basic environments (Shcherba, 1970)

porphyry type intrusions with allied mineralisation. It is significant that some of the Bolivian and Mexican Sn deposits are hosted within porphyry systems, whereas Cornubian, Malaysian, Iberian, Central-Western Europe and southern African deposits are all deeper seated deposits with related pegmatites and more importantly greisens (Fig. 2.5).

The critical factor in the evolution of mineralised greisen, porphyry and similar hydrothermal systems is therefore the development of a cupola or magma chamber above a parent batholith, essentially filled with high silica magma enriched in lithophile elements (Mutschler et al., 1981). These domes or chambers may be 0.5-1.5km in diameter and may extend more than 2-3km above the parent batholith, to within 1-3km of the surface (Mutschler, et al., 1981). Lithostatic confining pressures at the top of the cupola would be about 0.3-1.0kb with about 2.5-4.00wt% H_2O content (Burnham, 1979). Processes that may enhance concentration of Sn and W, plus other lithophile elements in these apical zones include:

i) Magma differentiation by fractional crystallisation - McCarthy and Groves (1979) report that prolonged crystal fractionation of a quartz-monzonite magma can result in the formation of a highly differentiated granite melt enriched in volatiles and lithophile elements.

ii) Liquid-state thermogravitational diffusion (Shaw et al., 1976; Hildreth, 1979), which allows migration of elements to take place in response to the thermal and gravitational field within a water-undersaturated stationary layer 1-2km thick at the top of a convecting magma column. A cupola may form which is enriched in W, Sn, Mo, Nb, Ta, U, Th, Rb, and volatiles including F, B, Cl, and depleted in Mg, Sr, and Ba (Hildreth, 1979).

iii) Vapour Phase Partitioning: Following vapour-saturation in the magma numerous constituents (including W, Sn, Mo) may be partitioned into the vapour phase (Holland, 1972). Gravitational rise of the vapour phase and rapid diffusion of magma constituents through the vapour phase may concentrate volatiles, alkalis and incompatible elements at the top of the magma column (Jahns and Burnham, 1972), and introduce these components to an overlying hydrothermal system (Burnham, 1967, 1979).

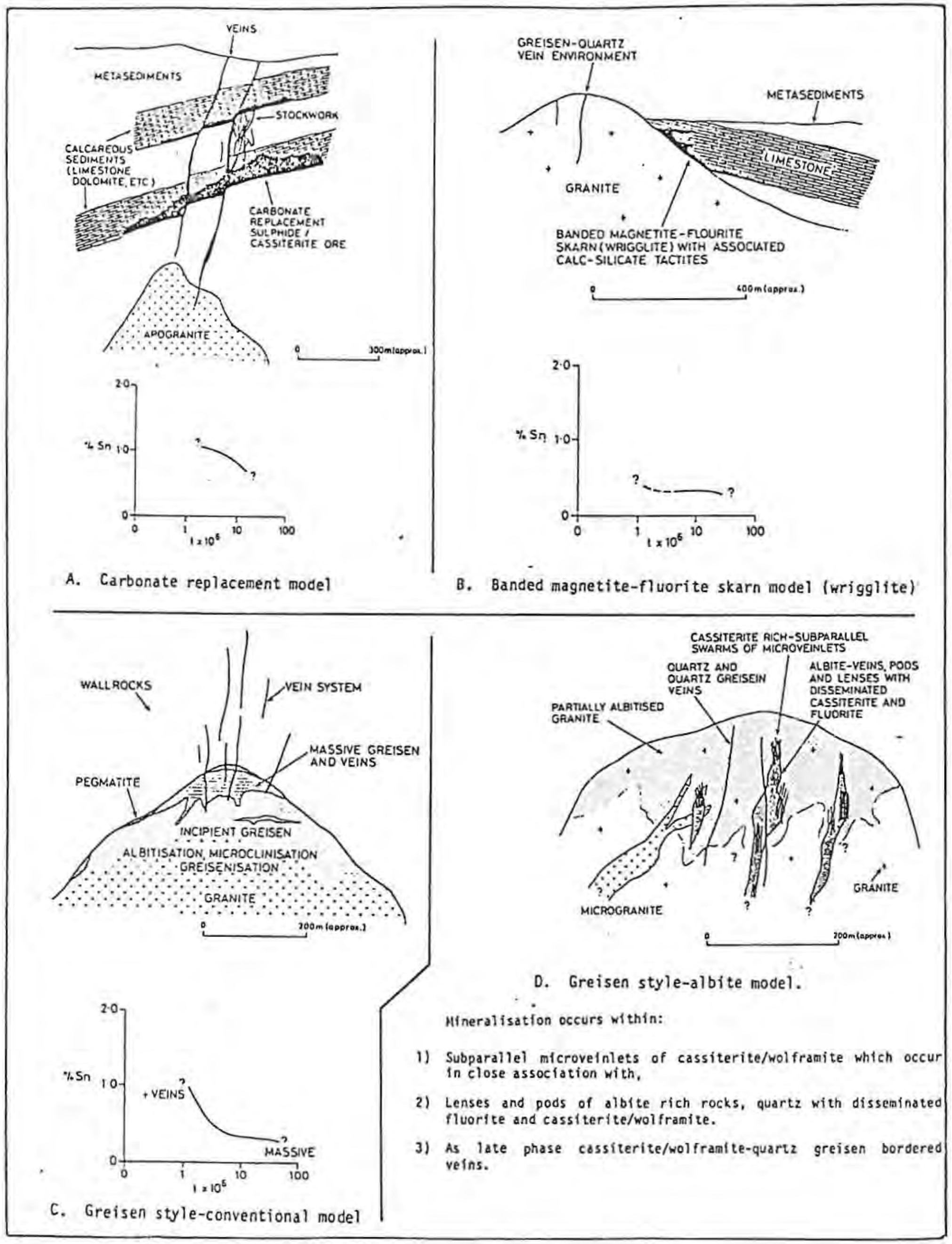


Figure 2.7: Aspects of the greisen environment, and likely tonnage grade curves (Taylor, 1979)

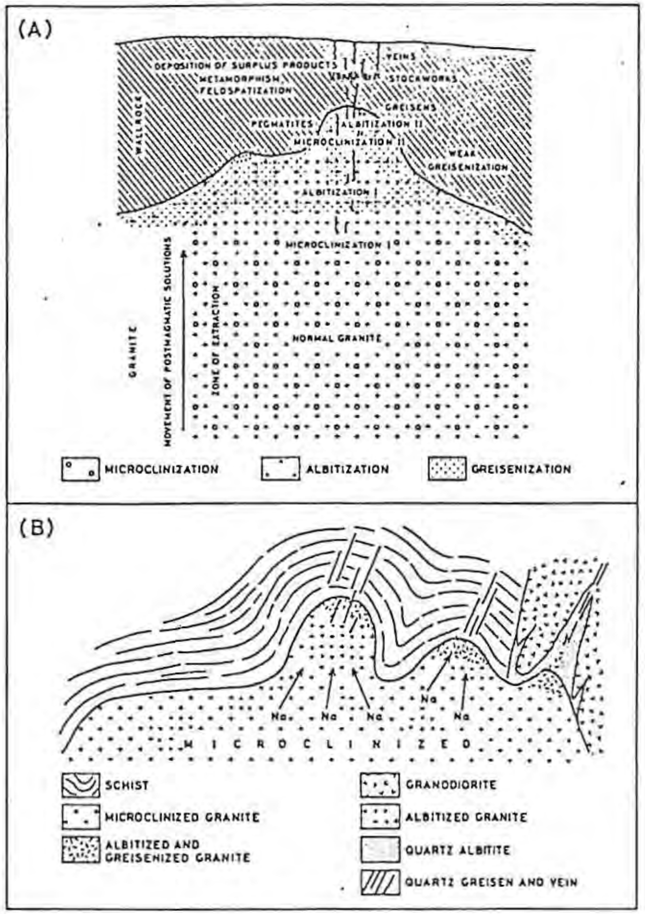


Figure 2.8: Aspects of the greisen environment

A) Alteration effects and vein systems (Shcherba, 1970)
B. Development of post-magmatic metasomatism (Beus and Zalashkova, 1964)

iv) Magma Convection was proposed by Whitney (1975) as a means of concentrating metal - and sulphide-bearing hydrothermal fluids derived from a large magma volume at the top of a magma chamber.

As cooling and crystallisation proceed, a vapour-saturated cap forms in the roof portions of the cupola, and partitioning of a supercritical phase occurs - the process of second, or resurgent, boiling (Burnham, 1979; Burt, 1981).

This is a critical stage in the evolution of porphyry and/or deeper seated greisen deposits, as substantial increases in fluid pressures may occur, which upon exceeding the lithostatic confining pressures, may cause hydraulic fracturing and failure of the roof zone - either allowing silicic magma to rise upwards, or the explosive escape of fluid and vapour phases. The ascent of the supercritical fluid marks the initiation of a magmatic-fluid dominated hydrothermal stage, a process leading to ore deposition. The formation of these hydrothermal fluids by partitioning (often by boiling) of a supercritical fluid phase and a vapour phase results in lighter species such as HF, HCl and H_3BO_3 being partitioned into the vapour phase whilst heavier saline species such as KF, KCl, and K_3BO_3 remain in the brine or melt (Burt, 1981). In the vapour salinity decreases markedly and there is a concomitant increase in acidity - the reverse applying to the residual (melt or brine) fluid phase. The result is a clear separation of chemical properties, causing potassic alteration (K-feldspar) and greisenisation (muscovite- topaz) of roof-zone rocks (Burt, 1981). The acid fluid is gradually neutralized, possibly by the interaction with wallrock, condensation into a low salinity fluid, or dilution by low-temperature meteoric waters (Burt, 1981). All of these processes could lead to deposition of W, Sn and associated sulphides.

Greisens with granitic pluton affinities are generally deeper seated than porphyry deposits. Associated W/Sn mineralisation with accompanying muscovitisation and greisenisation corresponds to the sulphide deposition in porphyries at the boundary between the potassic-alteration core and the outer sericite-pyrite shell (Mitchell and Garson, 1972).

Burt (1981) distinguished between porphyry-type greisenisation and vein-type greisenisation. The former occurs in a relatively near surface environment (1-3km) and boiling of brine or magma is the dominant early greisenisation process. Vein-type greisenisation (plutonic affinities) is more deeply seated (2-4km), and involves relatively low salinity aqueous fluids for which evidence of boiling (e.g. fluid inclusions show an increase in acidity in the vapour phase, and a decrease in salinity in residual brine or magma) is commonly lacking. Greisenisation is due to cooling and decompression at depth of moderately saline aqueous fluids at a temperature of around 500°C. Alteration assemblages are marked by broader chlorite and biotite fields, and muscovite occurs more commonly than topaz or Al-silicates. Garnet and magnetite are typically absent, and rhodochrosite and siderite or pyrite are the typical Mn- or Fe-phases (Burt, 1981).

Ore assemblages superimposed on the greisen alteration may display metallic zonation, with a core of Sn-W $\pm$Mo mineralisation, and sulphides of Cu, Pb, Zn, Bi-Ag extending vertically and laterally. Invariably this zonation is complex (as at Cornwall for instance), with two or more pulses of mineralisation and alteration superimposed. For a given pulse of mineralisation a vertical zonation may develop along the lines presented by Yan et al. (1980) for tungsten vein deposits in China (Fig. 2.9), and may be applicable to those discussed here. Furthermore there appears a common ore depositional sequence involving an oxide phase (cassiterite-wolframite-scheelite), followed by sulphide deposition with pyrite, pyrrhotite, chalcopyrite, arsenopyrite, molybdenite/sphalerite and lower temperature sulphosalts (e.g. Pb-Ag-Bi, 330°-180°C), and a final carbonate stage (calcite, siderite, ankerite, fluorite, pyrite, quartz and scheelite). Deposition of the various stages will be controlled by parameters such as fluid-wall rock interaction, fO_2, fS_2 and temperature.

2.5 Distribution of Sn-W bearing granites relative to geotectonic settings

The generation (and classification) of granites in various tectonic environments has been discussed above. This section outlines briefly the relative distribution of Sn(W) mineralisation and geological features of given terranes associated with the various granite types. The geotectonic subdivisions are after Mitchell (1979) and Mitchell and Garson (1981).

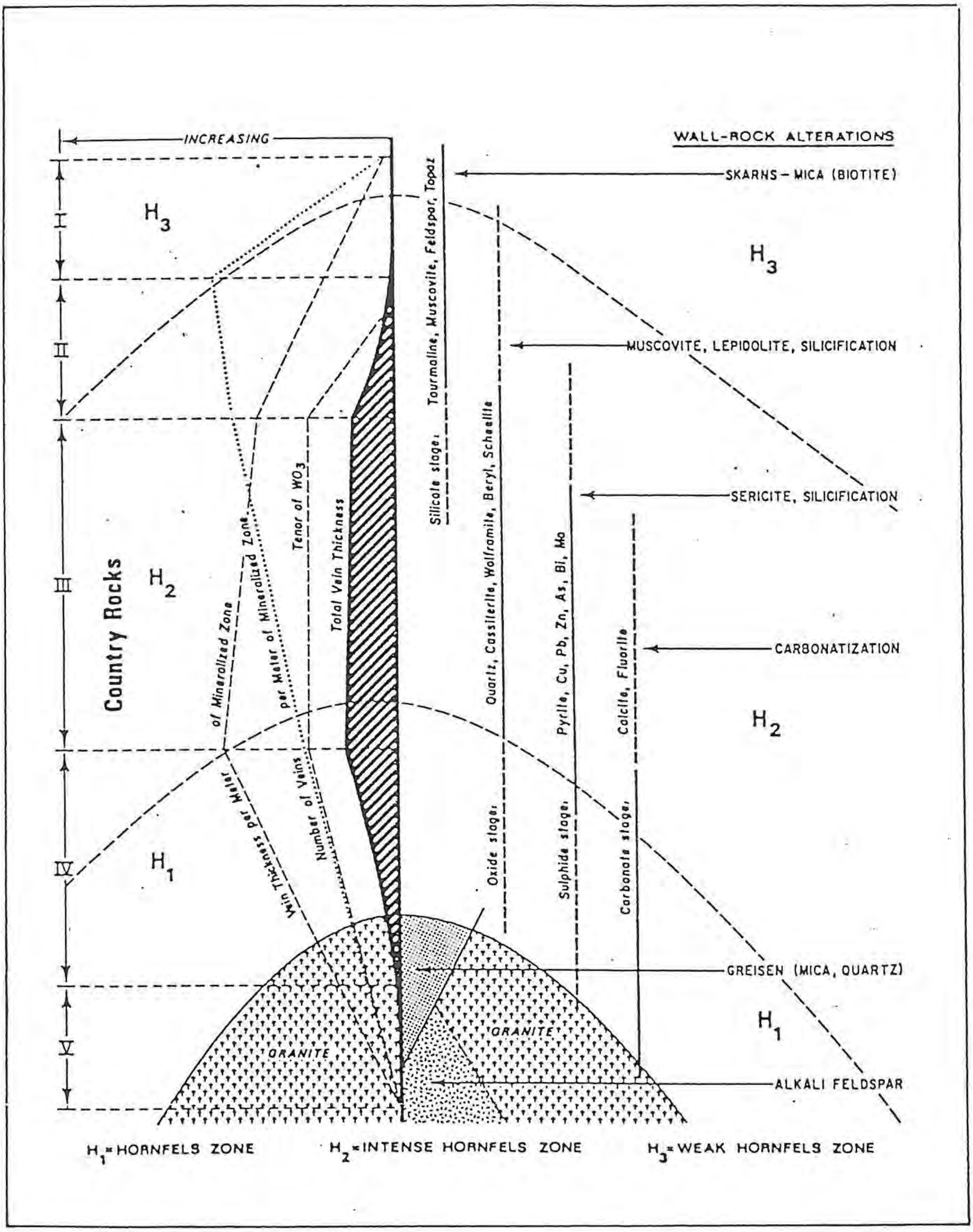


Figure 2.9: Vertical zoning pattern for tungsten vein deposits in China, that may be applicable to many of the areas discussed in this work (after Yan et al., 1980)

Rift-related mineralisation (Fig. 2.10a, 2.11a; e.g. Jos Plateau, Nigeria; Rondonia, Brazil)

Sn-W mineralisation is associated with granitic rocks mostly emplaced in the earliest stages of intracontinental rifting. Pre-intrusion rocks are usually either metamorphic rocks of an older orogeny or rift-related continental sediments and silicic volcanic rocks slightly older than the intrusive plutons. The mineralised igneous rocks are high-level granitic bodies, usually occurring as subvolcanic ring complexes, and some mineralisation may be hosted by the overlying roof rocks. Sn-W bearing plutons are mostly biotite-granites $\pm$syenites, peralkaline granites. They may be classified as anorogenic granites, with little deformation (usually only faulting) or metamorphism associated with their intrusion and associated mineralisation. Pegmatites, tourmaline and Be-minerals are uncommon.

Subduction related mineralisation (Fig. 2.10b, 2.11b)

In magmatic rocks Sn-W mineralisation is rare, and if present is associated with biotite or two-mica granites, adamellites, and rarely granodiorites, and rhyolitic to dacitic volcanic rocks (usually marked by strong hydrothermal alteration). The granitic belts are dominated by granodioritic rocks of I-type magnetite series affinity, which are syntectonic, with initial strontium ratios indicative of a lower crustal and upper mantle origin. The pre-pluton host rocks commonly include a thick succession of volcanic rocks erupted subaerially during pluton emplacement. Examples of Sn mineralisation in this environment are the Alaska-Aleutian Arc, and NE Honshu, Japan.

Outer arc granites (Fig. 2.10b) intrude very thick successions of deformed flysch and minor ophiolites. The geological features are very similar to collision settings (see below). Host rocks may differ, with higher proportions of flysch and basic and ultrabasic igneous rocks. Examples include the Outer Zone in SW Japan.

Back arc magmatic belts (Fig. 2.10b; e.g. Bolivia, western belt S.E. Asia)

These belts are usually convex to the continent, and are boarded on the continental side by oceanward-dipping thrusts. There is a genetic

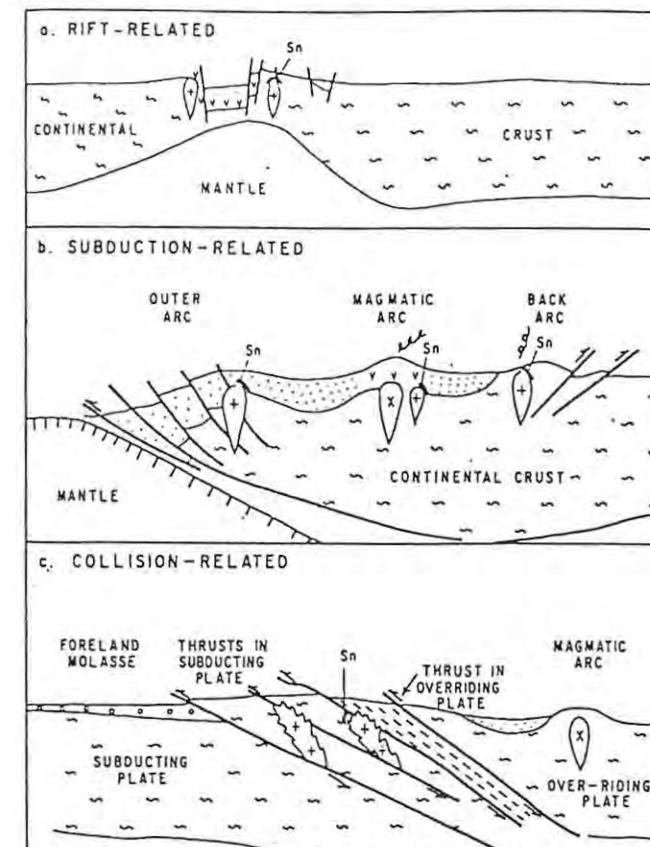
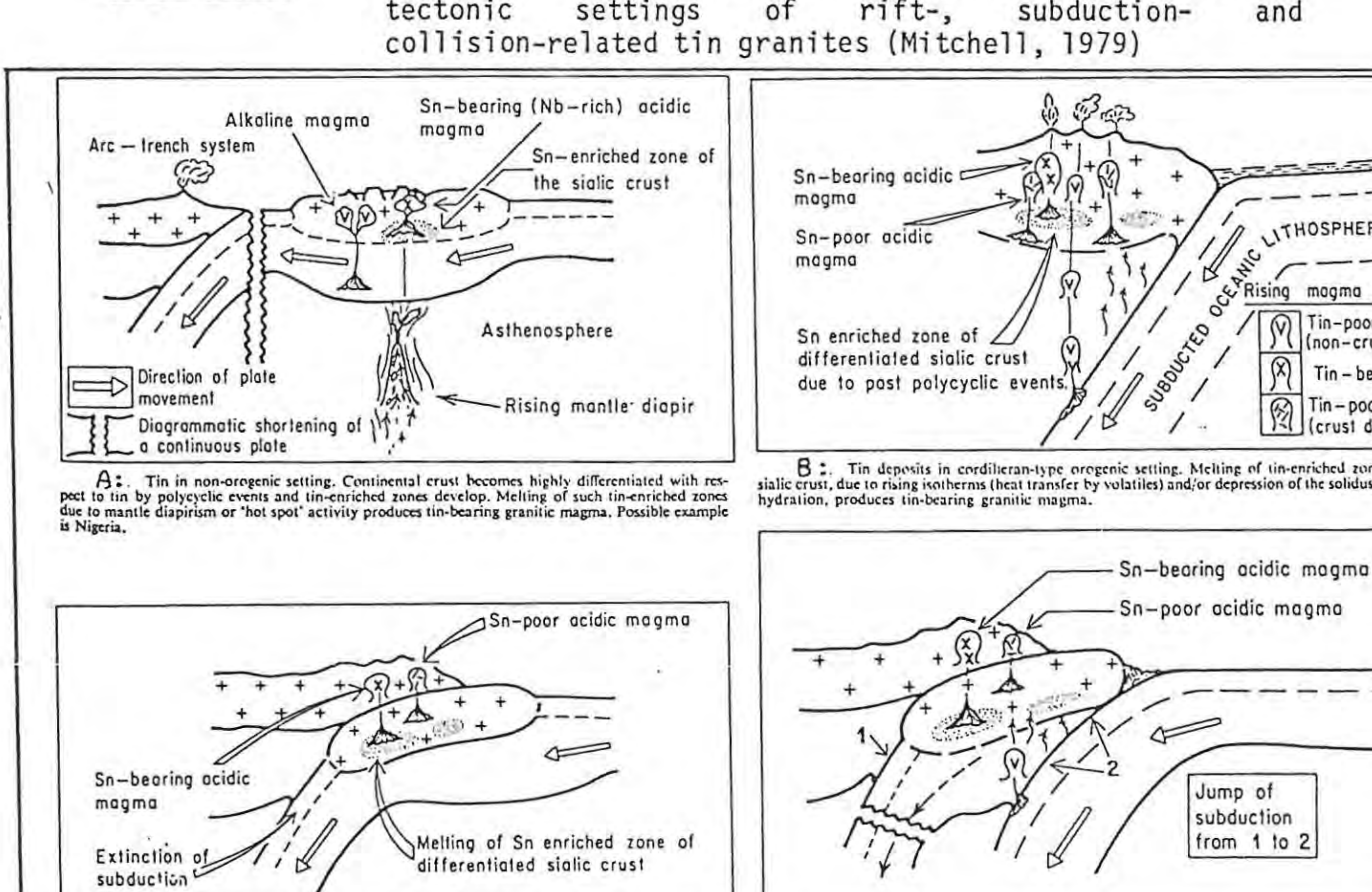


Figure 2.10: Schematic cross-sections showing main features of tectonic settings of rift-, subduction- and collision-related tin granites (Mitchell, 1979)



A: Tin in non-orogenic setting. Continental crust becomes highly differentiated with respect to tin by polycyclic events and tin-enriched zones develop. Melting of such tin-enriched zones due to mantle diapirism or 'hot spot' activity produces tin-bearing granitic magma. Possible example is Nigeria.

B: Tin deposits in cordilleran-type orogenic setting. Melting of tin-enriched zone in the sialic crust, due to rising isotherms (heat transfer by volatiles) and/or depression of the solidus through hydration, produces tin-bearing granitic magma.

C: Tin deposits in collision-type orogenic setting. Melting of the tin-enriched zone in the underthrust continental plate may ensue due to crustal thickening.

D: Tin deposits in complex orogenic setting (combined cordilleran and collision types). Melting of the continental crust may occur due to thickening and other subduction related processes. Temporal and spatial association of tin-bearing and tin-poor granites may result through melting of contiguous tin-enriched and tin-depleted zones in the crust.

Figure 2.11: Source models for tin mineralisation in various tectonic settings (Hutchison, 1979)

relationship between curvature, thrusting, granite emplacement and mineralisation. The interpreted scenario during subduction involves back-arc thrusting behind the granodioritic arc within a segment of the arc convex to the continent. Anatectic granites are generated relative to thickening of the continental crust and shear heating along the thrust at depth.

Collision related (Fig. 2.10c, 2.11c, d; e.g. Himalayas, Cornwall).

Sn/W mineralisation is associated with the upper parts of muscovite, 2-mica or tourmaline granites and leucogranites commonly with pegmatites and aplites. The granites are late- to post-tectonic, of S-type ilmenite series, massive or foliated and show sharp intrusive boundaries. The granites may be part of an inclined batholithic slab dipping towards the former overriding plate parallel to the regional major metamorphic foliation and thrust. Host rocks are typically tightly folded and often metamorphosed, and comprise shelf sediments of the foreland ±older foreland basement and flysch. Volcanic rocks of similar age to the granites are rarely preserved. Thrusts each side of the mineralised belt dip towards the suture, which invariably is characterised by a narrow zone of ophiolitic rocks. The overall structural vergence is towards the foreland. Greisen and vein-type deposits, as well as skarn type deposits due to the presence of limestones present in the foreland shelf succession, are common.

3.0 COMPARISON OF GEOTECTONIC SETTINGS

3.1 Variscan orogeny of Western Europe

The term Variscan is used in Anglo-Saxon scientific literature synonymously with Hercynian. The former is adopted here (after Rast, 1983), and is applicable to a general period of deformation and tectonism, over the period late Devonian to early Permian - a time cycle of about 100Ma. This period is characterised by variable sedimentation and volcanism patterns, and substantial granitoid emplacement, . A stratigraphic table of the European Palaeozoic, correlation of Carboniferous chronostratigraphy and the time of various tectonic events is shown in Tables 2-4.

The Variscan orogenic belt, after allowance for continental separation, is traceable from the gulf of Mexico to eastern Europe (Figs. 3.1A, B). In eastern North America it constitutes a late expression of the Appalachian fold belt, whereas in Europe it developed as a separate entity, distinct from the Caledonides to the north (Windley, 1984). In eastern Europe the Uralides emerged around the same time.

The vast amount of geological data accumulated for Western Europe is distributed in many journals, and written in many languages, making integration of existing knowledge difficult. In addition large areas of the Variscan belt are covered by younger rocks or have been reactivated during the Alpine tectonic cycle (Triassic to present). Many advances have been made over the last decade, although a totally acceptable synthesis of the Variscan orogen is still awaited. The following outline is largely from Rast (1983), Weber and Behr (1983), Giese et al. (1983); Windley (1984); Behr et al. (1984) and Lorenz and Nichols (1984).

3.1.1 Tectonic subdivisions

During the Variscan orogeny deformation and sedimentation were generally simultaneous, and due to a lack of isotopic constraints, stratigraphic chronology is the principal tool in tectonic subdivisions. The general relationships between major Carboniferous lithostratigraphic units are well known. Broadly, continental shelf deposits are recognised in Ireland, Britain, northern Belgium and Germany, while sediments of the Variscan orogenic belt lie to the southeast. The two regions are separated by a narrow marginal downwarp (Variscan foredeep of Krebs,

Table 2: Stratigraphic table for Palaeozoic Europe (Ager, 1980)

		System	Subsystem	Series	Stages	Alternative	Other	
PALAEOZOIC	UPPER	PERMIAN		U.	Tartarian, Kazanian	Thuringian	(Zechstein)	
				L.	Kungurian, Artinskian, Sakmarian, Asselian	Saxonian, Baigendzinian	(Rotliegendes)	
		280 million years						
		CARBONIFEROUS	U. (Silesian)		Stephanian	Uralian	Orenburgian, Gzhelian, Kasimovian	'U'
					Westphalian, Namurian	Moscovian, Bashkirian		'M'
			L. (Dinantian)		Visean, Tournaisian	Serpukhovian		
		345 million years						
		DEVONIAN		U.	Famennian, Frasnian	Strunian		
				M.	Givetian, Couvinian	Eifelian		
				L.	Emsian, Siegenian, Gedinnian	Lochkovian	Coblencian	
		395 million years						
	LOWER	SILURIAN (GOTHLANDIAN)				Přídolian		
				U.	Ludlow		Downtonian	
					Wenlock		(Middle Silurian)	
				L.	Llandovery	Valentian		
		435 million years						
		ORDOVICIAN		U.	Ashgill, Caradoc			
				L.	Llandeilo, Llanvirn, Arenig, Tremadocian	Canadian	(Middle Ordovician)	
		500 million years						
		CAMBRIAN		U.			Potsdamian	
				M.			Acadian	
				L.	Elankian, Botomian, Aldabanian, Tommotian		Georgian	
		570 million years						

Table 3: Correlation of Carboniferous chronostratigraphy (Rast, 1983)

Systems			Eastern USA	Mid-continent USA	British Isles		International Systems			Belgium	Donets basin USSR
Permian										Autunian	Asselian
Carboniferous	Pennsylvanian		Dunkard	Wolfcampian (Permian)	Stephanian	D	Carboniferous	Silesian	Westphalian		
			Monongahelan	Virgilian		C B A				Gap	Agzhelian, Kasimovian
			Conemaughian	Missourian		Cantabrian					Moscovian
			Alleghenian	Desmoinian	Westphalian	D C					
			Pottsvillian	Atokan		B A				B A	Bashkirian
			Pocahontasan	Morrowan	Namurian	Yeadonian, Marsdenian, Kinderscoutian, Alportian, Chokierian, Arnsbergian, Pendleian				Namurian	Serpukhovian
	Mississippian	Upper	Chesterian	Chesterian	Dinantian	Brigantian, Asbian, Holkerian, Arundian, Chadian	Dinantian			Viséan V 1, 2, 3	Viséan
		Lower	Valmeyeran	Meramecian, Osagean							
			Kinderhookian	Kinderhookian		Courceyan				Tournaisian Tr 1, 2, 3	Tournaisian
Devonian											

Table 4: Time scale of the Upper Palaeozoic and duration of magmatic activities, formation of intermontane troughs, uplift and subsidence of continental crust, and retreat and transgression of the sea in the area of Variscan Europe (Lorenz and Nicholls, 1984)

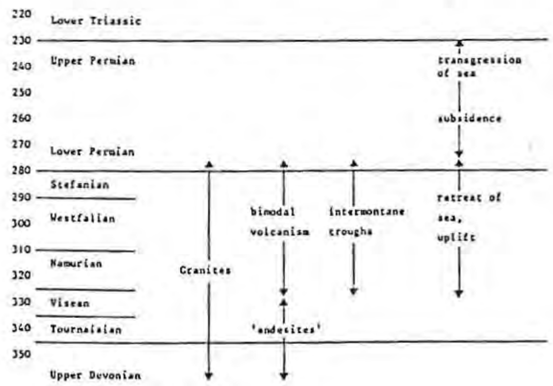


Table 4A: Variscan fold phases in Europe (Rast, 1983)

Phase	*Timing*
Pfälzic	Permo-Triassic
Saalic	Early Permian (Rotliegendes)
Asturian	Westphalian C to Stephanian
Erzgebirgian	Namurian–Westphalian boundary
Sudetic	Viséan to early Namurian
Bretonic	Late Devonian to Early Carboniferous

1976), which was eventually filled by Upper Carboniferous clastics (fore-arc accretionary prism?) The southern edge of the Variscan belt is imperfectly understood because in Europe it is overprinted by the Alpine orogenic belt (Rast, 1983).

Various interpretations have been made of the complex structures in southwest Europe (i.e. France and the Iberian Peninsula). In view of the fact that the Variscan massifs of central Europe are disconnected (Fig. 3.2), attempts have been made to extend faults from the better exposed Iberian Peninsula into France (Rast, 1983). Arthaud and Matte (1977) emphasised this feature, interpreting Late Palaeozoic wrench faulting in southern Europe and northern Africa as a right-lateral shear zone induced by the relative motion of two plates - a northern one including the Canadian Shield, Greenland and stable Europe, and a southern one (Gondwanaland) that includes the African Shield. Badham (1982) advocates small scale localised deformation of microplates, with oblique motion and development of strike-slip orogens capable of compressional and extensionsal tectonism. Further discussion involving some of these ideas is given in 3.1.3.

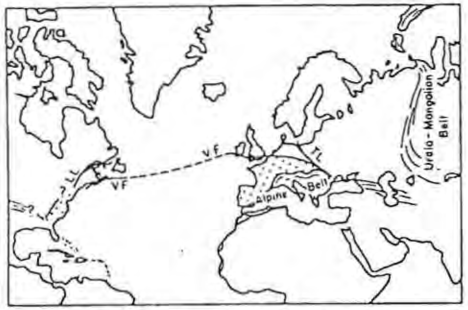
Most of the rocks in Europe deformed during the Variscan orogeny crop out in isolated massifs (Figs. 3.1B, 3.2), and are collectively termed the Variscan belt (Rast, 1983). Three main zones of contrasting stratigraphy and tectonic history are recognised

i) Rhenohercynian,
ii) Saxothuringian, and
iii) Moldanubian (Fig. 3.2C, Rask, 1983; Behr et al., 1984; Windley, 1984).

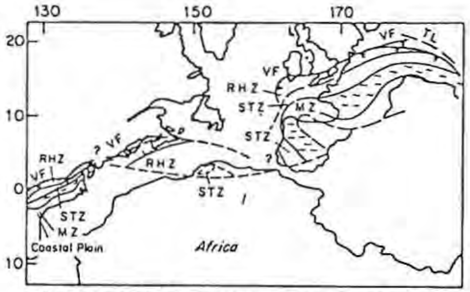
The stratotectonic subdivisions (after Kosmatt, 1927; Stille, 1951; in Rast, 1983) stem from the recognition of an association between the Variscan orogenic phases and the underlying Precambrian basement. Krebs (1976) further subdivided the Variscan Belt, distinguishing discrete unmetamorphosed sedimentary units ("fore deeps") peripheral to the main orogenic domains (Fig. 3.3D). These flysch-molasse troughs may tentatively represent wedges of fore-arc? sediments related to bipolar subduction processes during the Carboniferous (see later).

Rhenohercynian Zone (RH, Figs. 3.1C, 3.3A)

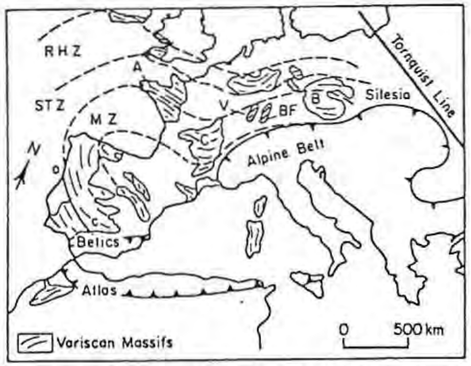
This zone is commonly taken to include most of the Sub-Variscan foredeep (see Fig. 3.3D). It comprises Devonian and Carboniferous sediments,



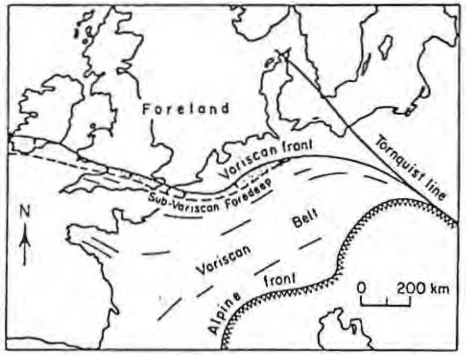
A : The Variscan orogenic belt in Europe and eastern North America. LL, Logan line; TL, Tornquist line; VF, Variscan front. The Variscan orogenic belt is coarsely stippled and possible African equivalents (Michard and Sougy 1977) are finely stippled.



B : Transatlantic zonal correlation after allowing for opening of the Atlantic. MZ, Moldanubian zone; RHZ, Rhenohercynian zone; STZ, Saxothuringian zone; VF, Variscan front; TL, Tornquist line. ———, major transform faults.



C : Hercynian (Variscan) massifs in Europe. A, Armorican Massif; B, Bohemian Massif; C, Massif Central; V, Vosges; BF, Black Forest; RHZ, Rhenohercynian zone; STZ, Saxothuringian zone; MZ, Moldanubian zone. ——— zonal boundary, O/C Oporto–Cordoba lineament.



D : Major boundaries of Variscan belt in eastern Central Europe and the position of the Sub-Variscan foredeep (stippled).

Figure 3.1: Major features of the Variscan fold belt in Europe (Rast, 1983)

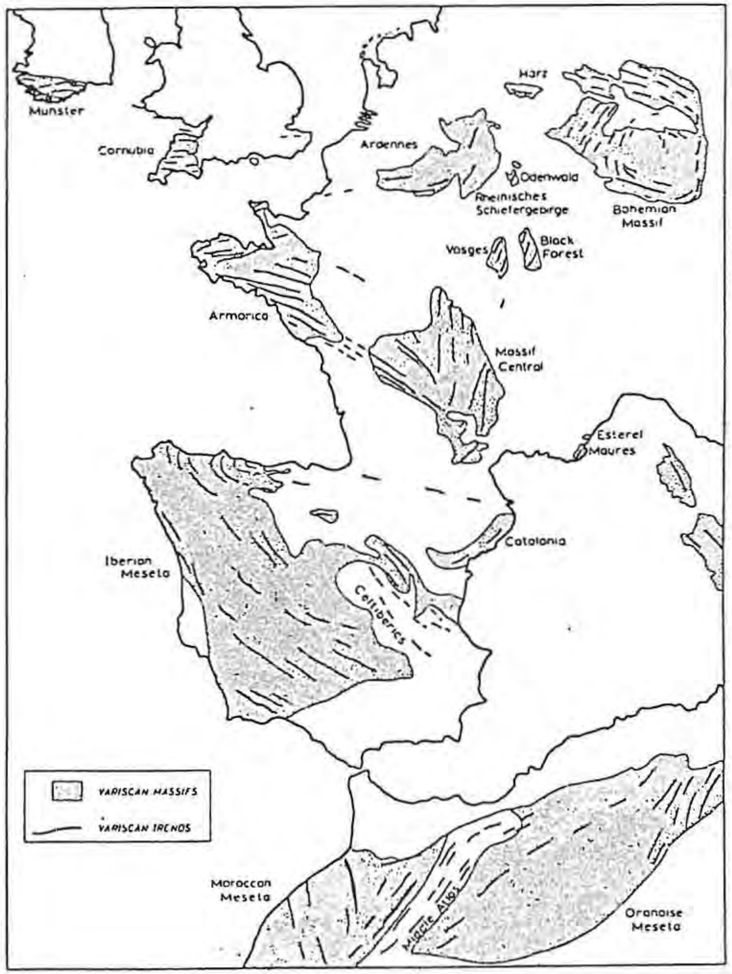


Figure 3.2: General trends of the Variscan massifs of western Europe (Ager, 1980)

Structural subdivision of the Variscan Belt of Central Europe, showing the distribution of several types of thrust complexes. Arrows—main transport directions; *RH*—Rhenohercynian Zone; *ST*—Saxothuringian Zone; *MO*—Moldanubian Region; *L.B.M.*—London/Brabant Massif (Caledonian); *D.M.*—Domnoneo-Mancellian (Cadomian) and Icartien (Lower Proterozoic); *T.P.*—Tepla/Barrandean (Cadomian); *M.S.*—Moravo-Silesian (Cadomian).

Figure 3.3A: Structural subdivision of the Variscan belt of Central Europe (Behr et al., 1984)

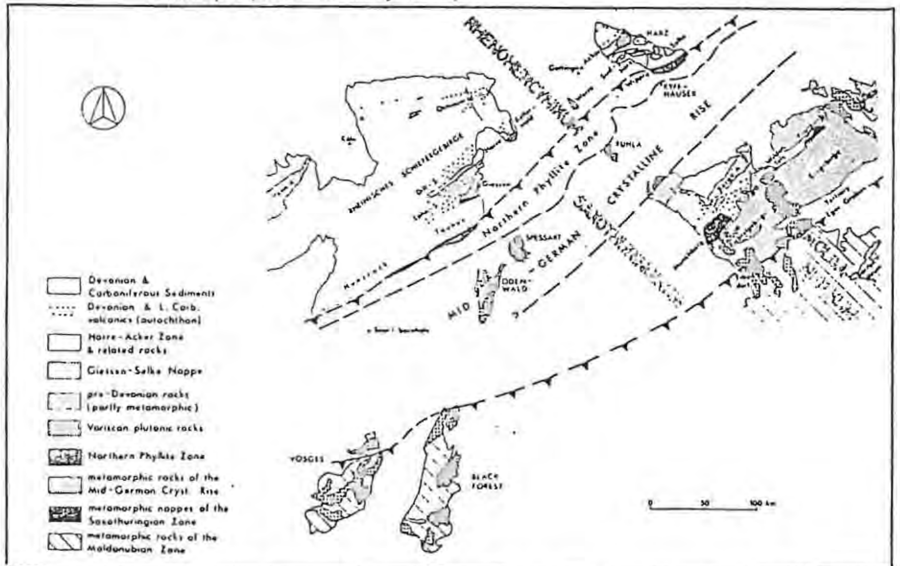
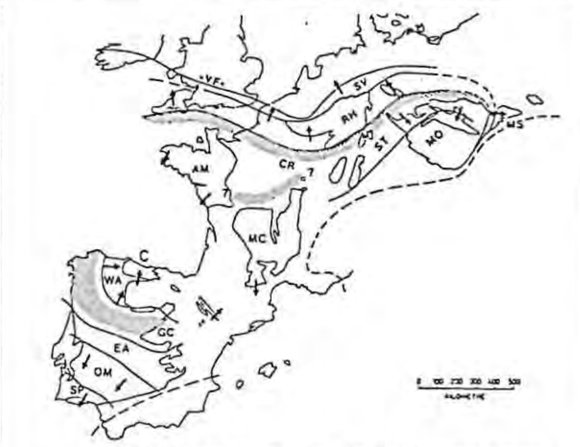


Figure 3.3B: Structural subdivision of the Variscan outcrops in Germany (Behr et al., 1984)



Variscan zones, crystalline median massifs (stippled) and direction of vergences (arrows) in *Meso-Europa*. 'VF': Variscan Front; SV: subvariscan zone; RH: Rhenohercynian zone; CR: Central German crystalline rise; ST: Saxothuringian zone; MO: Moldanubian zone; MS: Moravosilesian zone; AM: Armorican Massif; MC: Massif Central; C: Cantabrian zone; WA: West Asturian-Leonesian zone; GC: Galician-Castilian zone; EA: East Lusitanian-Alcudian zone; OM: Ossa-Morena zone; SP: South Portuguese zone

Figure 3.3C: Variscan zones, crystalline median massifs and direction of vergences in Meso Europe (Krebs, 1976)

The mobile flysch-molasse foredeeps in the external parts of the *Meso-European* Variscan fold belt. The arrows indicate the direction of migration. SP: South Portuguese zone; C: Cantabrian zone; SV: Subvariscan zone; MS: Moravosilesian zone

Figure 3.3D: Location of the Variscan flysch-molasse foredeeps (Krebs, 1976) consisting of shelf deposits to the north, and deeper water flysch-type deposits to the south (Rast, 1983). Basement to the north in Britain and probably Belgium is Avalonian (Late Precambrian to early Palaeozoic) volcanic and sedimentary rocks.

The RH strata are generally strongly deformed, although in northern parts the marginal metamorphism is mild (not exceeding greenschist facies). The main episode of tectonism is Middle to Upper Carboniferous (Westphalian C to Stephanian, Table 3.4; Rast, 1983). Structurally the RH zone differs from those of the continental belt in that it contains some large-scale recumbent folds. The predominant late Carboniferous deformation is N-verging thrusts and overfolds, and large horizontal nappe displacements (Windley, 1984).

The Variscan Front (Fig. 3.1D) is a term given to the most continuous line of overthrusting and/or overfolding which is associated with the junction of the RH zone and the Sub-Variscan foredeep/Caledonide Zone. In other places the front may comprise a zone of strong deformation or a sedimentary facies change. In the eastern Variscan the front changes in style from a thrust to transcurrent (transform ?) faults, marked by the so-called Tornquist line (Figs. 3.1C, D), which extends from northern Denmark to the Black Sea.

The sedimentary history of the RH zone is marked by a marine transgression in the Devonian, with erosion of continental facies and clastic detritus derived from a landmass to the NW (the uplifted Caledonian orogenic belt in W. Norway, NE Scotland, East Greenland, Windley, 1984), and deposition of greywackes, sandstones, orthoquartzites and shales. During the Mid-Devonian shallow marine facies evolved, with conspicuous reef limestones, and there was also widespread extrusion of sodic keratophyres and spilitic lavas and tuffs (Windley, 1984). The boundary between the RH and Saxothuringian Zone (ST) is interpreted as a site of subduction (Behr et al., 1984; Lorenz and Nicholls, 1984) or as a back arc marginal basin (Badham, 1982).

Mid German Crystalline Rise (MGCR)

The southern edge of the RH zone is marked by the Mid German crystalline rise, which includes massifs such as Odenwald, Spessart and parts of Thuringian Forest (Fig. 3.3B). These massifs comprise Cambro-Ordovician rocks, with cores of Precambrian gneisses (Rast, 1983). Variscan folding and metamorphism is recognisable.

Saxothuringian Zone (ST, Figs. 3.1C, 3.3)

The ST lies immediately south of the MGCR and continues westward into the Armorican Massif (see Fig. 3.1C). The Precambrian basement has been affected by late Precambrian (Cadomian) tectonism, with large rounded domes of high grade metamorphic rocks, with Variscan deformational overprints. The ST is characterised by Upper Devonian volcanics and intrusives, and highly variable sedimentary facies. Two periods of local deformation (late Devonian (Bretonic) and mid-Carboniferous (Sudetic)) are discernable. The southern border of the zone is marked by the occurrence of rocks indicative of suture zone blue schists, basic-ultrabasic granulites (in S. Brittany), eclogites, serpentinites, granulites, gabbros, tuffs and cherts (NW Spain), and glaucophane schists further to the west (Windley, 1984). An age of 450Ma is ascribed to this high pressure granulite-grade metamorphism. The suture is also marked by a prominent geophysical anomaly, inferring the presence of a major mafic body (possibly relic ocean crust).

Moldanubian Zone (MZ, Fig. 3.1C, MO, Fig. 3.3A)

The type locality of the MO is in the Bohemian Massif, from where the zone extends into the Black Forest, parts of France (Massif Central) and the Iberian Peninsula. Caledonian (middle Devonian) and Variscan age deformation is recognised (Rast, 1983), with the latter imposing a weak imprint (folding) on the main Caledonian event (400-365Ma). As such the MO involves Lower Palaeozoic rocks superimposed on a late Precambrian basement (Cadomian) which together are overlain by Upper Palaeozoic strata, with each of the major sequences being affected by early Carboniferous tectonism, metamorphism, widespread syn- and post-tectonic granite intrusion (mid-Carboniferous), and acid to intermediate explosive volcanism.

The Variscan orogeny was complete by the early Permian, except for localised post-orogenic intrusion of lamprophyres, and extrusion of acid and basic volcanics and local graben formation (Windley, 1984).

3.1.2 Variscan magmatism and metamorphism

Magmatism

Variscan magmatism occurred over various stages of the orogeny, and can be assessed in terms of pre-orogenic volcanism associated with basinal sedimentation, large-scale plutonism at the orogenic culmination, and post-orogenic volcanism after cratonic stabilisation (Floyd et al., 1983). In general terms Variscan magmatism is bimodal with basic and acid products characterising the RH and ST zones. Activity in the MO zone was more variable with a high proportion of calc-alkali andesites (Floyd, 1982).

Pre-orogenic volcanism was submarine in character with relatively thin, localised development of massive and pillowed lavas throughout the Devonian and Early Carboniferous. The abundance of the extrusive products generally increased from the early Devonian to a late Devonian maximum, before rapidly decreasing at the end of the Lower Carboniferous (Visean). The lavas were all basaltic, and show a number of chemical features which support an ensialic environment for the Variscan orogeny (Floyd, 1982l Floyd et al., 1983). These are

i) Tholeiitic basalts predominate, although alkali basalts are found at the northern (SW England) and southern (South Austria) margins of the belt. The tholeiites are generally characterised by incompatible element abundances typical of continental rather than oceanic environments.

ii) The distribution of the immobile elements, Ti, Zr, Y shows that nearly all the basalts and dolerites represent intra-plate volcanism, irrespective of location, age, magma-type or mode of emplacement (Exley et al., 1983).

A sustained, widespread sublithospheric thermal anomaly is required to generate basaltic volcanism over a wide orogenic belt, such as the Variscan, for a prolonged period of time. Tectonic interpretations include lithospheric stretching and rifting, above an enormous mantle plume, or a slowly subducted slab of lithosphere taken down into the mantle at a very low angle such that a wide area was affected above the subduction zone.

Late- to post-orogenic magmatism is marked by the emplacement of voluminous calc-alkali granitoid batholiths, with associated granite porphyry feeder dykes and rhyolitic lavas. Although poorly dated, an important period of granitoid emplacement appears between 330-280Ma (Lorenz and Nicholls, 1984; Derré, 1982), with attendant uranium, tin and tungsten mineralisation. The most voluminous intrusions are evident within the MO, where granites and migmatites are abundant (Windley, 1984). Syn-tectonic migmatites and post-orogenic granites also occur in the RH, the latter related to the important Cornubian Sn deposits.

An assessment of much the major and trace element geochemical data led Floyd et al., (1983) to conclude that the bulk of Variscan granitoids were derived from anatexis of lower crustal material - in particular widespread partial melting of a poorly hydrated garnet-bearing granulite, which produced large volumes of essentially dry granitic magma. Subsequently, by the combination of assimilation of wet country rock and anhydrous crystallisation, this magma became saturated and crystallised below the surface as granite batholiths. The range of compositions seen in individual plutons is explained by local, but minor, variations in source rock composition or degree of melting, together with later differentiation, contamination and metasomatism (Floyd et al., 1983). Some localities of Variscan granites have associated subaerial rhyolitic volcanism (e.g. Cornwall). This volcanism is attributed to smaller volumes of late-stage dry granitic magmas probably derived from a slightly different souce, and which travelled up fractures within the granite and surrounding country rocks. In general the granites have a restricted compositional range, high K/Na ratios (giving alkali feldspars), low Fe_2O_3/FeO ratios (inhibiting magnetite), a deficiency in hornblende and sphene, and intermediate to high initial Sr isotope ratios. In summary the published descriptions and analyses of most of the Variscan granites point to an S-type ilmenite series categorisation. In the Cornwall region collisional tectonics are invoked to produce the granite melts, (Mitchell and Garson, 1981) whilst elsewhere in western Europe a process of 'tectonic stacking' (with thrusting to the NW), subsequent to 'thin skinned' extensional tectonism, provided an environment capable of generating granites (see 3.1.3).

Metamorphism

Many of the regional metamorphic features have been mentioned in 3.1.1. The Variscan of western Europe is marked by the extensive development of high temperature - low pressure regional metamorphism, especially in the MO and ST zones (Zwart, 1979; Rast, 1983). The RH primarily characterised by low grade metamorphic facies and extreme polyphase deformation. Metamorphic events are distinguished during the late

Precambrian (Cadomian) and mid-Palaeozoic (Variscan), corresponding to peaks at 650,450, and 350-300Ma (Suk, 1977; in Rast, 1983). Limited high pressure assemblages have been described, and are spatially related to the suture zone along the southern border of the ST zone, marking a site of a former subduction to the northwest (see 3.1.3). Windley (1984) invokes the close association of blue schists and high-pressure granulites along the suture as analagous to those of the Himalayan Indus Suture in Pakistan, which is a well documented continental-collision case study (e.g. see Coward et al., 1982; Windley, 1984).

3.1.3 Geodynamic models

The Variscan fold belt has been subjected to many different interpretations, largely in terms of plate tectonics (Ager, 1975; Alderton et al., 1979; Windley, 1977, 1984; Behr et al., 1984; Lorenz and Nichols, 1984) and also in terms of lithospheric thermal disturbances (Zwart and Dornsiepen, 1978), and vertical and shear tectonics (Badham and Halls, 1975; Krebs, 1976; Badham, 1982).

Recent articles (Floyd, 1982; Giese et al., 1983; Weber and Behr, 1983; Behr et al., 1984; Lorenz and Nichols, 1984) point to a combination of ensialic (intracontinental) processes with limited subduction, and later continent-continent collision, with some strike-slip component. Problems in plate tectonic analyses/ interpretations are compounded by the general lack of magmatic plate margin characteristics such as typical ophiolite sequences, island arc or continental margin volcanic rocks (esp. andesites) and voluminous intermediate plutons. Recent studies (above) have elucidated the existence of basaltic rocks with mid-ocean ridge basalt affinities, high pressure metamorphic rocks, trench deposits, and large scale allochthonism, which may have effected the pre-Variscan basement, and locally has led to inversion of the metamorphic isograds (Behr et al., 1984).

Large scale thrusting is now recognised at <u>all</u> crustal levels, and substantiated by geophysical methods, which necessitate interpretation of the structures in terms of horizontal layering. Giese et al. (1983) envisage the Variscan orogen as forming from thin- and thick-skinned tectonics. Compressive movements during Late Ordovician through Carboniferous started from an attenuated crust which was stretched and thinned by rifting processes. During the subsequent orogeny sedimentary and basement complexes were detached from their underlying stratum, and horizontally displaced (up to 100km) and stacked - resulting in crustal thickening by a factor of 2-3 times (35-45km) by the Permian (Fig. 3.5). Thick crustal segments such as the Mid German Crystalline Rise underwent strong uplift and erosion (Giese et al., 1983).

Plate convergence forced closure of the RH and ST basins (the mid-European Ocean) and one or more Mediterranean basins to the south of the pre-Variscan blocks (Paleotethys sea, Fig. 3.4; Behr et al., 1984; Lorenz and Nicholls, 1984). These workers also postulate subduction from north and south under the dorsal core region (Fig. 3.6). Problems still arise with the timing and driving mechanism of the bilateral activity (Behr et al., 1984), however it gave rise to an extensive zone of hot, partially molten upper mantle above and between the subduction zones. Mafic magmas were generated from the diapiric uprise of portions of the melt, which also facilitated widespread partial melting in the lower and middle crust, high T, low P metamorphism in crustal rocks, and regional uplift and extension of the crust (Lorenz and Nicholls, 1984).

The Variscan belt, as presently known, originated from the collision of a Gondwana and Laurasia 'megaplates' (Perroud et al., 1984). This Carboniferous collision subsequently formed Pangaea and produced the Appalachian - Variscan orogeny.

Lorenz and Nicholls (op. cit.) envisage that diachronous collisions induced oroclinal bending of the Variscan, largely owing to the reduced rigidity of what is thought to be hot island-arc crust, and the irregular continental margins of the larger and thicker continental plates.

Further extensional tectonics over 20-30Myrs into the Lower Permian (back arc spreading of Badham, 1982; Floyd, 1982) continued after the collision processes,with concomitant upper mantle activity, and crustal uplift (Lorenz and Nicholls, op. cit). During Permian time upper mantle activity declined, with crustal subsidence and peneplanation, followed by Upper Permian marine transgression of the Zechstein sea (from the north) and the Tethys sea (from the south), marking the culmination of the Variscan geodynamic cycle.

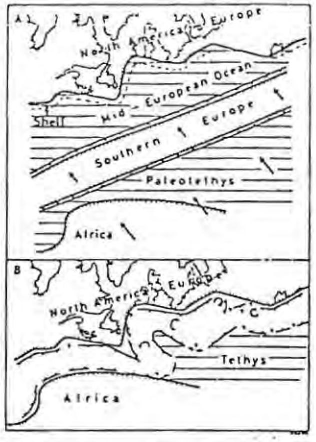
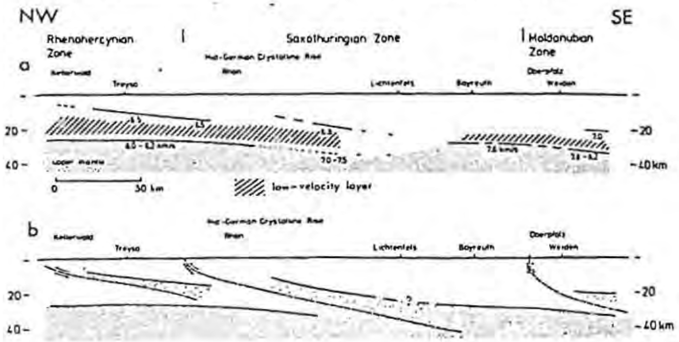


Figure 3.4: Pre- and post-collision tectonics in Southern Europe (Lorenz and Nicholls, 1984)

. A. Sketch map of the pre-collision Southern Europe plate (in ?Lower Carboniferous time) with its two subduction zones (in this stage probably not active simultaneously), the bordering oceanic areas, and the irregular continental margins of North America-Europe and Africa. Arrows indicate schematically direction of plate movements relative to North America-Europe (movements may have been more oblique).
B. Sketch map of the Southern Europe plate near the end of the continent-continent collisions. The Southern Europe plate adjusted to the irregular continental margins of the large neighbouring plates by intense internal plastic distortion, i.e. by oroclinal bending and thus formation of sub-plates. Curved arrows indicate sense of rotation of sub-plates. The area marked Tethys may have been underlain, in part, by Gondwana continental crust.

Figure 3.5A: Refraction seismic profile through the RH and ST Zones.
a. Simplified representation of the velocity/depth functions.
b. Tectonic interpretation of discontinuities as upthrust slices of lower crust and/or upper mantle material (Behr et al., 1984)



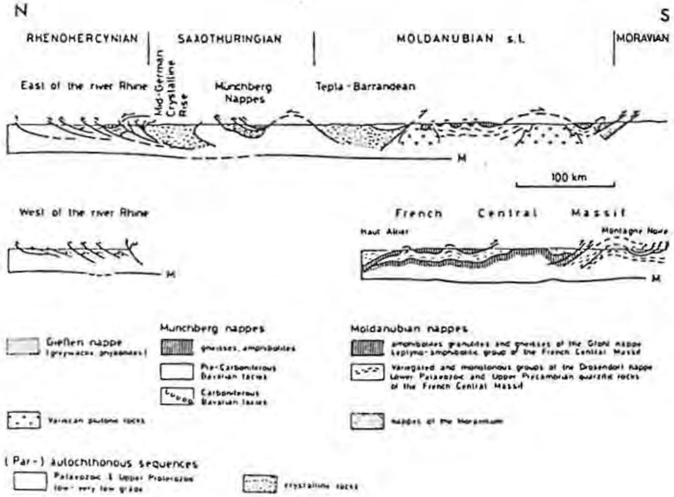
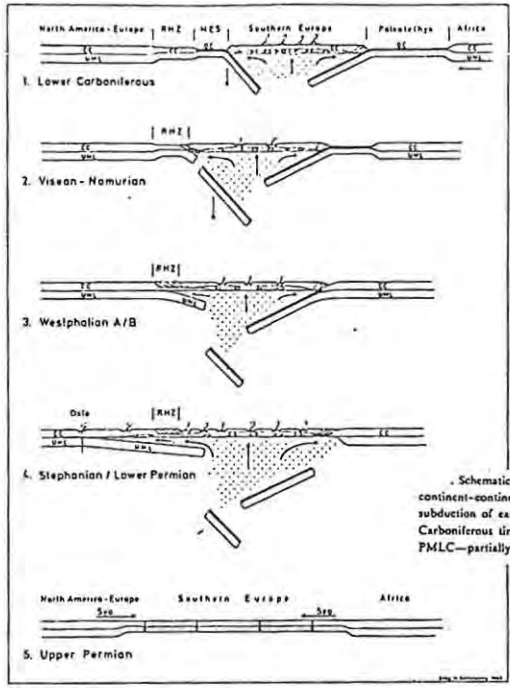


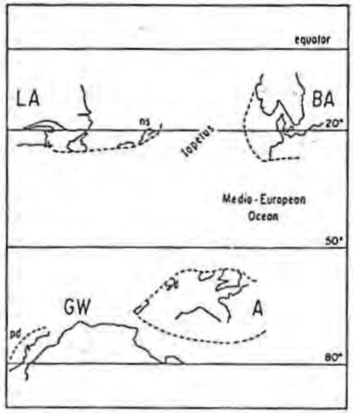
Figure 3.5B: Diagrammatic cross-section through the Variscan Belt in Central Europe. M = crust-mantle boundary (Behr et al., 1984)

Figure 3.6: A) Geodynamic development of Variscan Europe (Lorenz and Nicholls, 1984)



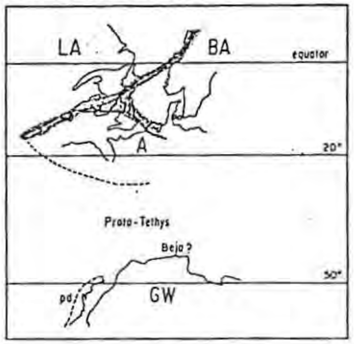
. Schematic diagrams showing potential geodynamic development of Hercynian Europe owing to continent-continent collisions with North America-Europe and Africa in Late Paleozoic time. Possible subduction of earlier oceanic crust of back-arc basins within Hercynian Europe in Devonian-Lower Carboniferous time omitted. *CC*—continental crust; *MES*—Mid-European sea; *OC*—oceanic crust; PMLC—partially molten lower crust; *RHZ*—Rheno-Hercynian zone; *UML*—upper mantle lithosphere.

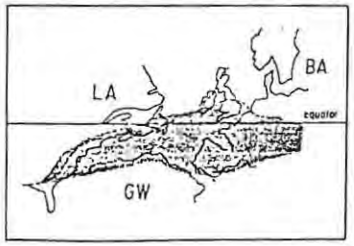
B) Paleogeographic reconstructions of the Variscan Belt from Ordovician to Early Permian based on paleomagnetic data (Perroud et al., 1984)



. Paleogeographic reconstruction for Ordovician time from paleomagnetic data. A, BA, GW, and LA denote Armorica, Baltica, Gondwana, and Laurentia, respectively. Armorica plate in this reconstruction includes most of high-inclination domain shown in Figure 1; pd = Delaware Piedmont, ns = northern Scotland.

. Devonian paleogeographic reconstruction after closure of both Iapetus and medio-European Ocean. Shaded zones are areas of collision between plates; they reveal Y shape of Caledonian (and Acadian) orogenies. Position of Gondwana is derived from paleomagnetism of Msissi norite (Hailwood, 1974) and Beja gabbro of Portugal (see text); pd = Delaware Piedmont province.





. Late Carboniferous-Early Permian reconstruction after Van der Voo (1982). Ruled area represents zone of Hercynian deformation due to collision of Gondwana and Laurasia in Carboniferous time. This zone is especially large in western Europe where Armorican domains have undergone ensialic deformation.

Constraints on this scenario include the uncertainty of the exact limits of the pre-Variscan blocks, as are the existence and width of oceanic areas (Perroud et al., 1984; Behr et al., 1984). In the ST zone and the south-facing parts of the MO zone there is good evidence of early Palaeozoic rifting and of late Ordovician through Carboniferous plate convergence. Only limited oceanic basins are inferred from palaeomagnetic climatic constraints in these zones, whereas a wider ocean is likely at the RH-ST boundary (Behr et al., 1984).

3.2. Malay Peninsula

South east Asia is a major Sn/W producing region of the world, accounting for about 45% world Sn production and, including China, approximately 30% world W. The major producing centres of the Thai-Malay peninsula lie within a belt some 3000km long and up to 800km wide, sometimes referred to as the Central Tin Belt of Southeast Asia (Mitchell, 1977; Mitchell and Garson, 1981). This section concentrates on the geotectonic setting of the Malay peninsula, where two major tin granitoid belts are separated by a central basin of sedimentary deposits, and effectively constitute a paired orogenic batholith belt (Hutchison, 1983).

3.2.1 Tectonic subdivision

Peninsula Malaysia can be divided into four geotectonic zones (Fig. 3.7; Hutchison, 1977):

1) A western stable shelf is characterised by Lower and Upper Palaeozoic miogeoclinal sedimentary formations (exposed on the islands of Langkawi, Perlis and Kedah), that are gently folded and generally unmetamorphosed. Granite is not abundant in this zone.

2) The Main Range Belt is characterised by the huge Triassic Main Range batholith, whose axis forms the mountain range from Malacca towards Thailand. These granites are emplaced into Lower Palaeozoic isoclinally folded phyllitic (and lesser marble) greenschist facies metasediments.

3) The Central Graben comprises gently folded Triassic and Mesozoic sediments underlain by more strongly deformed Permian rocks. The sedimentary sequences change from marine in the Permo-Triassic to continental facies by the Jurassic.

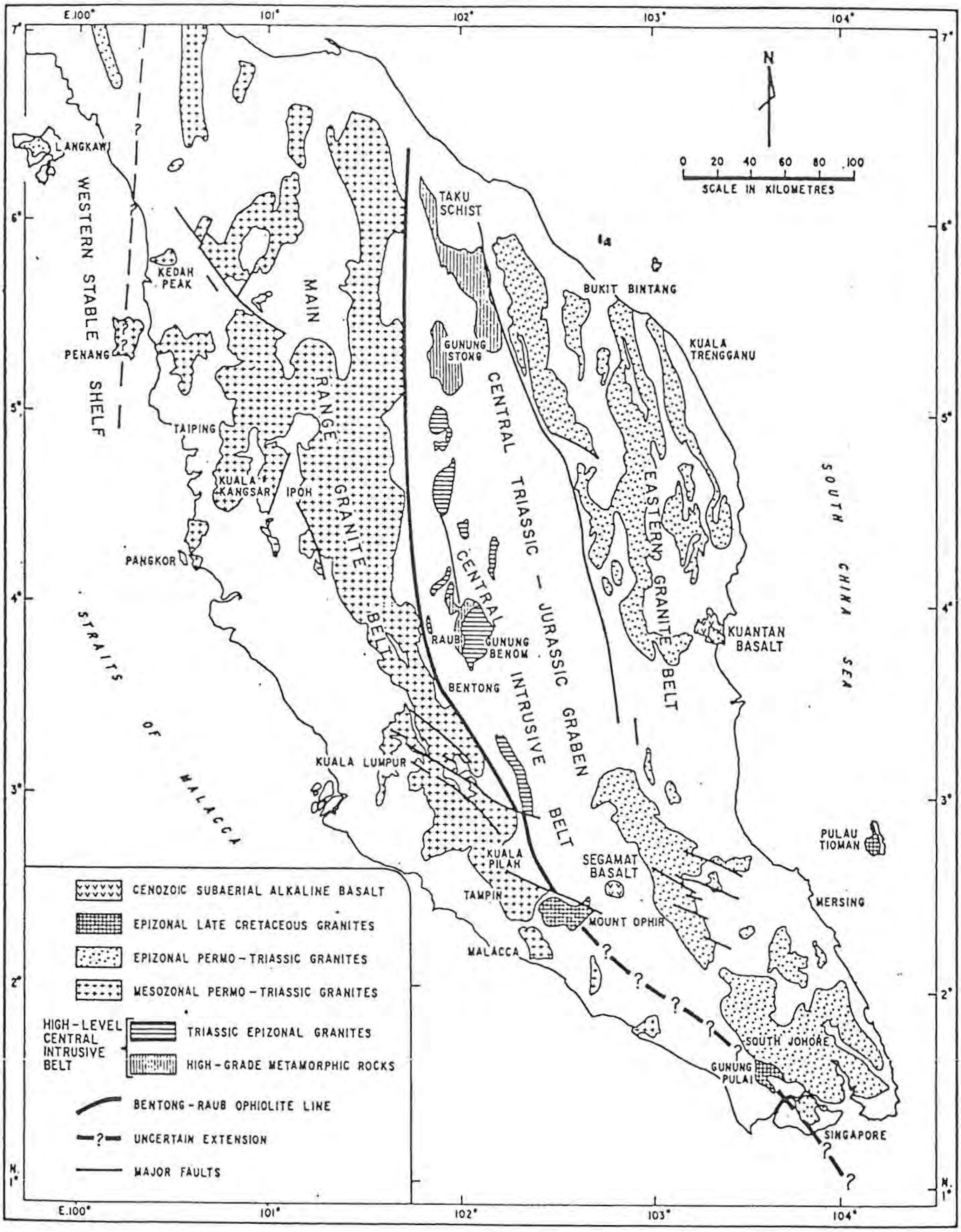


Figure 3.7: Outline geological map of the Malay Peninsula, showing the distribution of granites relative to the major tectonic features (Hutchison, 1977)

The western margin of the graben is bounded by the Bentong-Raub suture ophiolite, which separates it from the Main Range Belt. The ophiolite consists of a phyllitic melange in which blocks of serpentinite and metabasic rocks occur with radiolarian chert. Both Hutchison (1977, 1983) and Mitchell (1977) recognise the ophiolite as a suture marking a former subduction zone, but with different tectonic interpretations (see later). The Central Graben is separated from the Eastern Belt by the Lebir Fault.

4) The Eastern Belt features numerous elongate granitic plutons intruded during the Permo-Triassic through gently deformed, predominantly unmetamorphosed Permo-Carboniferous sedimentary sequences with associated pyroclastic and volcanic rocks of acid of intermediate composition.

   Minor, post-tectonic Late Cretaceous granites are known near Ophir and Gunung Pulai (Fig. 3.7).

3.2.2 Granitoid plutonism

The Main Range and Eastern Belt granites have been well studied in terms of their petrography, geochemistry and metallogeny (Hosking, 1973, 1977; Hutchison, 1977; Ishihara et al., 1979; Beckinsale, 1979; plus others). The two Belts are distinct in many facets, including their geodynamic evolution (see below), even though they were emplaced broadly over similar periods (Permian-Triassic). The main features and differences are outlined in Table 5. Variations in attendant mineralisation are discussed in Section 4.3.

The principal differences between the Main Range and Eastern Belts lies in their depth of emplacement, notably higher Rb/Sr ratio for the Main Range, porphyritic and muscovite-bearing nature of the former, as well as the conspicuous lack of a thermal aureole or attendant andalusite or cordierite. Ishihara et al., (1979) note the dominance of ilmenite-bearing granitoids in the Main Range Belt, with mixed magnetite- and ilmenite-bearing granites in the Eastern Belt. Central Graben and post-tectonic Cretaceous granites are distinctly magnetite-bearing. The implications of these variations regarding granitoid genesis are outlined below.

TABLE 5: Summary of properties of Malaysian Sn-belt granitoids (after Hutchison, 1977; Hosking, 1973; Hutchison and Taylor, 1978; Ishihara et al., 1979)

PROPERTY	MAIN RANGE BELT	EASTERN BELT
Emplacement	Mesozonal (6-10km)	Epizonal (1-6km)
Age Rb:Sr	Maxima at:- 200, 230, 280 Late Carboniferous, to Late Triassic	Maxima at:- 220, 250 Permian to Triassic
K:Ar	80-200Ma	220-250Ma
Initial $^{87}Sr/^{86}Sr$	Carboniferous 0.7111 Triassic 0.7098	Permian 0.7102 Triassic 0.7075
Cassiterite Pleochroism	Highly pleochroic (red-pale) Nb/Ta	Non - to weakly pleochroic
av. K_2O weight %	5.0	4.1
av. K_2O/Na_2O	1.7	1.3
av. Rb/Sr	10.0 increasing westwards	2.7
Femic minerals	Biotite Muscovite only in greisen	Biotite Hornblende is locally important
texture	coarse strongly porphyritic	equigranular to microporphyritic medium to coarse
intrusive bodies	extremely large batholith	large plutons
country rocks	Low grade phyllitic Palaeozoic rocks, regionally metamorphosed	Carbo-Permian gently deformed sediments (predominantly unmetamorphosed)
Contact metamorphism from granite	Slight local increase of dynamothermal grade. No andalusite or cordierite	Pronounced aureole containing andalusite and cordierite
Other features	No volcanic association	Associated volcanics, gabbro-tonalites
	Hot springs are common	Basaltic and Lamprophyric dykes are common
Classification	S-type, Ilmenite series	S- and I-type Ilmenite and Magnetite Series

3.2.3 Geodynamic models

Plate tectonic models accounting for the geodynamic evolution of the Thai-Malay peninsula have been proposed by Mitchell (1977); Hutchison (1977) and Beckinsale (1979) (Figs. 3.8, 3.9). The essential difference between Mitchell's and Hutchison's models lies in the sense of subduction proposed to generate the Eastern Belt granites. Otherwise the tectonic synthesis is similar. Beckinsale incorporates a further collision event in the Eocene (130-100Ma) to account for younger granites constituting the Western Belt in Thailand, and not discussed here.

The following synthesis is largely after Hutchison (1977, 1983). During the Early Palaeozoic the eastern Malay peninsula developed as an island arc above active subduction beneath a probable Gondwana craton (Fig. 3.8A). Back-arc rifting separated this region from the craton, forming a marginal basin (Fig. 3.8B, the Central Graben). During late Permian the spreading ceased, and there was a reversal in subduction polarity (Fig. 3.8C), which culminated in a late Triassic collision event between the island arc and its ancestral craton. Eastern belt granites, in the main consisting initially of S-type ilmenite series granitoids were generated after subduction to the west and during subduction to the east (Figs. 3.8B, C). Later I-type magnetite series granitoids evolved during the collision event when subduction had ceased (Fig. 3.8D). The evolution of the Main Range S-type ilmenite series granitoids is directly related to the collision event. In the collision model (Mitchell, 1977) the foreland of western Malaya approached the subduction zone and underthrust the magmatic arc. The Bentong-Raub suture ophiolite was evolved by westward thrusting of older oceanic crust and cherts, whilst sediments of the foreland were isoclinally folded, overthrust to the west and locally metamorphosed, and the late-to-post tectonic Main Range granites were generated along eastward-inclined intracontinental thrusts (Mitchell, 1977, Fig. 3.9C). The reader is referred to Table 5 for a summary of features and differences between the Main Range and Eastern Belt granites. Importantly the nature of mineralisation associated with these magma types is distinctive, with the S-type ilmenite series granitoids hosting abundant Sn/W mineralisation, whilst I-type magnetite series granitoids have often associated with porphyry Cu, Au, Fe ($\pm$W) deposits (e.g. Loei, Eastern Belt Thailand, Fig. 3.9).

Main Range granitoids were emplaced mesozonally (6-10km, Fig. 3.10; Hutchison, 1977, 1983), as evidenced by the lack of pronounced contact

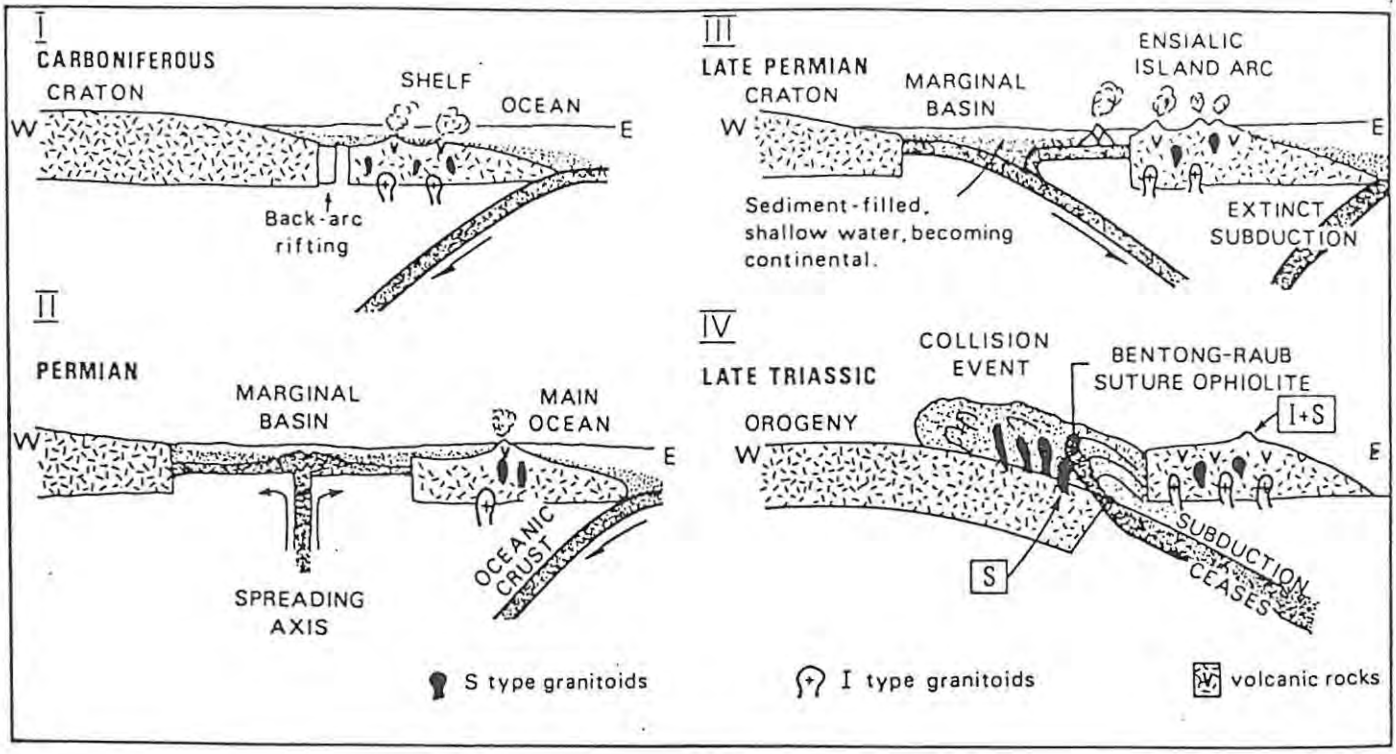
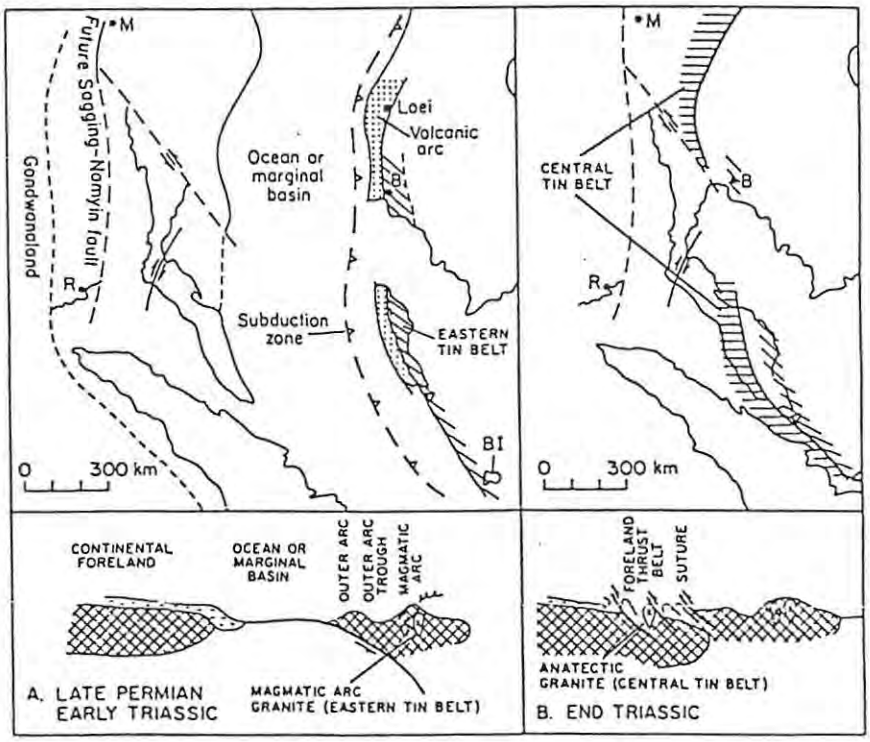


Figure 3.8: Evolution of the paired granitoid belts of the Malay Peninsula (Hutchison, 1977, 1983)



(A) Late Permian to Triassic eastward subduction of ocean floor beneath eastern Malaya and northeastern Thailand. Porphyry copper mineralization at Loei and tin mineralization in eastern Malaya and Billiton Island within magmatic arc. (B) Late Triassic collision of western Southeast Asia block with northeastern Thailand-eastern Malaya; tin mineralization associated with granites of foreland thrust belt in Main Range Malaysia and Thailand. B, Bangkok, BI, Billiton Island, M, Mandalay, R, Rangoon.

Figure 3.9: Schematic cross-sections of the geodynamic evolution of the Malay Peninsula (Mitchell, 1977, Mitchell and Garson, 1981)

aureoles. The actual depth of emplacement is dependent on the regional geothermal gradient and the water content of the magma (see 3.5 for further discussion). Chemical, isotope and geological data invoke generation by anatexis of continental sialic basement during the collision event. Since the Triassic the Main Range has been substantially uplifted (K:Ar dates infer early Jurassic to late Cretaceous uplift), with present outcrop levels indicating unroofing of the batholith as a whole.

Eastern Belt granites were emplaced at higher crustal levels (1-6km, Fig. 3.10) with concomitant loss of water into the unmetamorphosed country rocks, and pronounced contact metamorphism. Chemical, isotope and geological data such as the associated rhyolite and pyroclastic extrusives, and gabbroic precursors indicate much of the Eastern Belt likely to be derived from a more basic parent, possibly a mantle wedge and/or subducted oceanic lithosphere related to the earliest periods of subduction activity. The granites are still exposed at a high level, and the lack of erosion indicates that the region has been very stable since the Triassic (Hutchison, 1977).

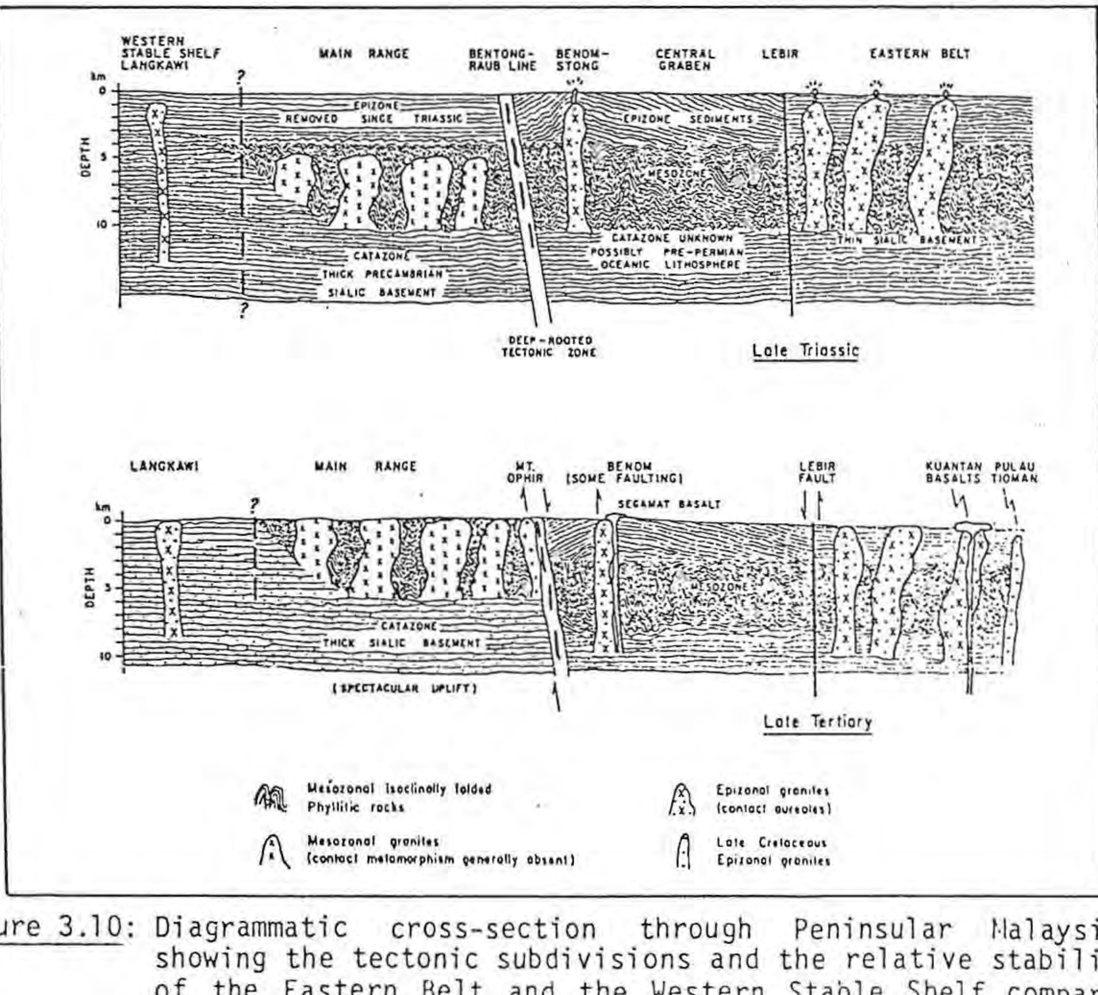


Figure 3.10: Diagrammatic cross-section through Peninsular Malaysia, showing the tectonic subdivisions and the relative stability of the Eastern Belt and the Western Stable Shelf compared with the Main Range Belt, which has been uplifted several kilometres since the Late Triassic (Hutchison, 1977)

3.3 Pan African tectonism in Southern Africa

Late Proterozoic tectonism, commonly termed the Pan African event, effected large areas of the African continent and adjoining areas of Brazil, Arabia, and Madagascar (Kröner, 1979; Martin and Porada, 1977; Porada, 1979). This event occurred over various stages, and effectively constituted the final episode of a lengthy period of Proterozoic instability. The term "Pan African" is applied principally to a thermo-tectonic episode around 450-550Ma, although in general terms it can be used pertaining to rift-bound troughs (aulacogens and continental margin geosynclines) affected by this event and by earlier tectonism (1000Ma-450Ma; Tankard et al., 1982).

In southern Africa the Late Proterozoic tectonism is related to partial continental splitting, rifting and spreading of the Kalahari, Congo and eastern South American cratons (Fig. 3.11), comprising a central portion of the larger Gondwana landmass. This resulted in the formation of a 3000km chain of rift-bound troughs following the western and southern coasts of the subcontinent. The chain is preserved as a series of remnants knowns as the coastal branch of the Damara Province, the Gariep Province and the Saldanian Province (Fig. 3.11; Tankard et al., 1982). An intracontinental branch of the Damara Province extends NE through Namibia towards south central Africa. This has been interpreted as an aulacogen or failed rift (Martin and Porada, 1977), related to a triple junction which eventually split forming the Ribeira (South America) and Damara Provinces, the northern branch and Gariep Province (to the south) constituting the preserved remnants of the triple point arms. The Saldanian Province formed at the same time, is less well exposed, but appears to have similar rift affinities although the sedimentation patterns reflect more a passive continental margin setting (see 3.3.3). The geotectonic synthesis of these Provinces has been complicated by subsequent (and conjectural) collisional/subduction processes in the Cambrian (Damara orogeny) and Permo-Triassic (Cape Fold Belt orogeny). The following outline concerns only the Saldanian and Damaran Provinces as they host Sn-W mineralisation, discussed in Section 4.3.

3.3.1 Saldanian Province

The term "Saldanian" is here used to encompass late Precambrian tectonism that affected a number of continental margin basins in the Cape region,

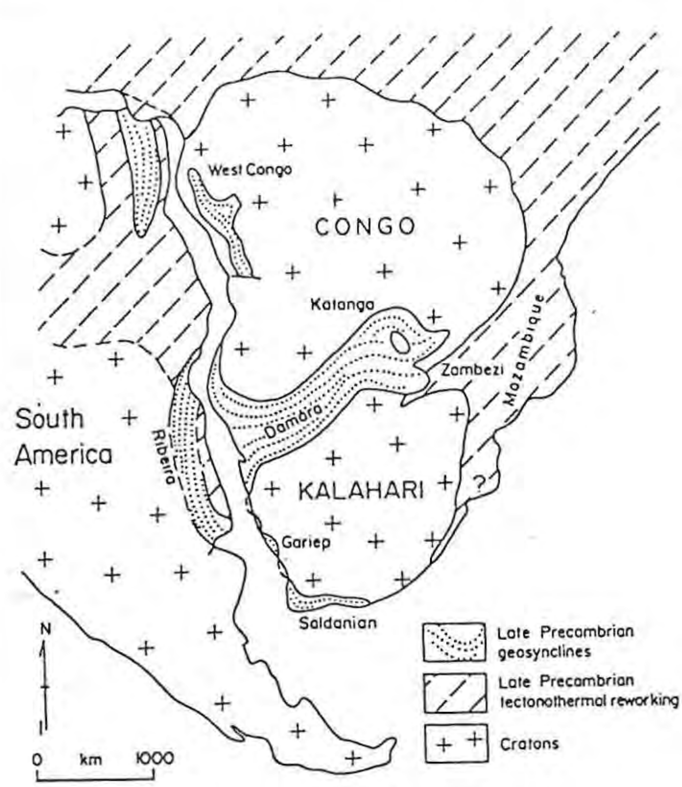


Figure 3.11A: Map showing the Late Precambrian structural framework of central and Southern Africa and eastern South America (Tankard et al., 1982)

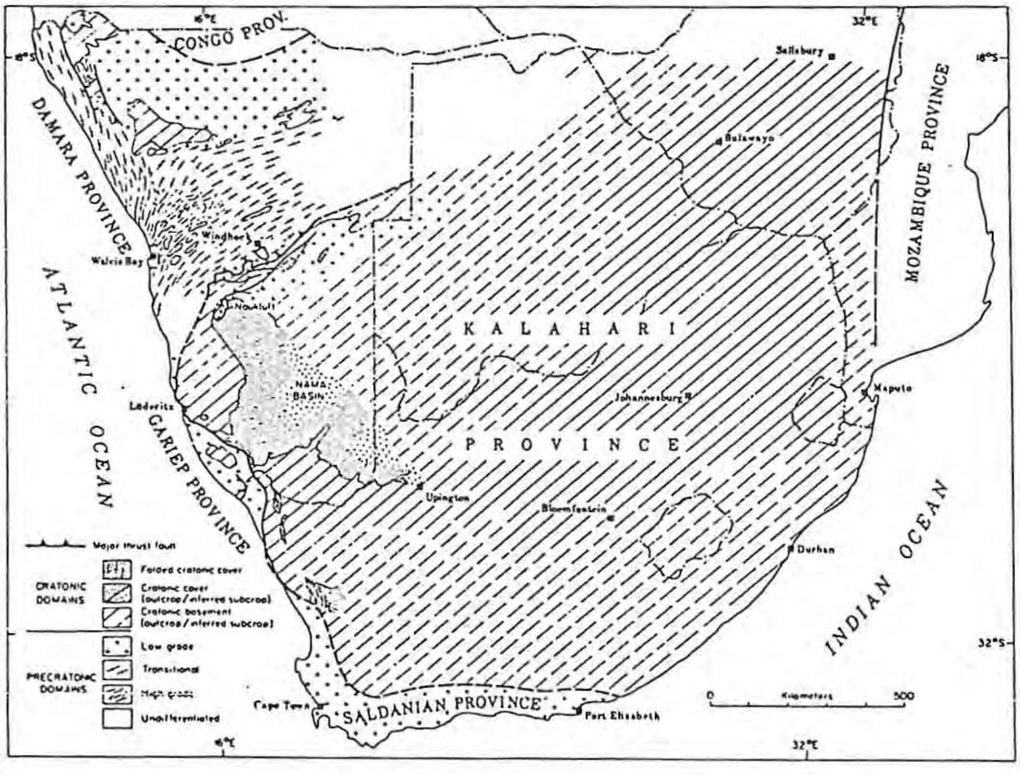


Figure 3.11B: Nonpalinspastic map showing the late Proterozoic rift basins and the ensialic Nama basin from 1000Ma to 500Ma (Tankard et al., 1982)

now preserved as scattered inliers of pre-Cape Fold Belt age (pre-Mesozoic, Fig. 3.12). A tectono-stratigraphic subdivision includes the Malmesbury Group in the southwestern Cape (Fig. 3.13), and three elongate inliers in the east comprising the Kango, Gamtoos, and Kaaimans Groups (Fig. 3.14). The stratigraphy of these groups is listed in Tables 6 and 7.

The poorly exposed pre-Cape inliers comprise a varied low-grade assemblage of sedimentary and metasedimentary rocks, which have been affected by tight upright folding along NW-trending axes, with considerable displacement by shearing and faulting, during the Pan African orogeny (Tankard et al., 1982). General structural complexity (especially overprinting by the later Cape Fold Belt orogeny) and poor exposure obscure the facies and stratigraphic relations and conceal the extent of lithostratigraphic units. The sedimentary succession has been intruded by syn- and post-tectonic granitoids (Cape granites).

Sedimentation

Initiation of sedimentation was around 950Ma, contemporaneous with rifting (the 'Proto Atlantic') and downwarping along N-S and NW-SE trends, and continued to about 600Ma.

In the west the Malmesbury Group has a eugeosynclinal character, whereas elsewhere the sediments characterise marine-shelf and alluvial environments. Two major zones of dislocation, the Saldanha-Franschhoek and Piketburg-Wellington fault zones, divide the Group into three discrete northwest trending tectonic prisms - southwestern, central, and northeastern domains (Fig. 3.13; Hartnady et al., 1974; Tankard et al., 1982).

The Tygerberg Formation, which comprises the southwestern domain, is dominated by turbiditic sediments, with minor limestone and conglomerate beds, and local exposures of altered volcanic rocks (tuffs and brecciated flows of intermediate composition). The predominance of argillites and fining upward sequences in the greywackes suggest slow hemipelagic deposition interrupted by occasional turbidite currents between abyssal plains and lower fans (Hartnady et al., 1974; Walker, 1978).

Central domain sedimentation is more variable, but appears to characterise more proximal deposition, with upwards-increasing grain size

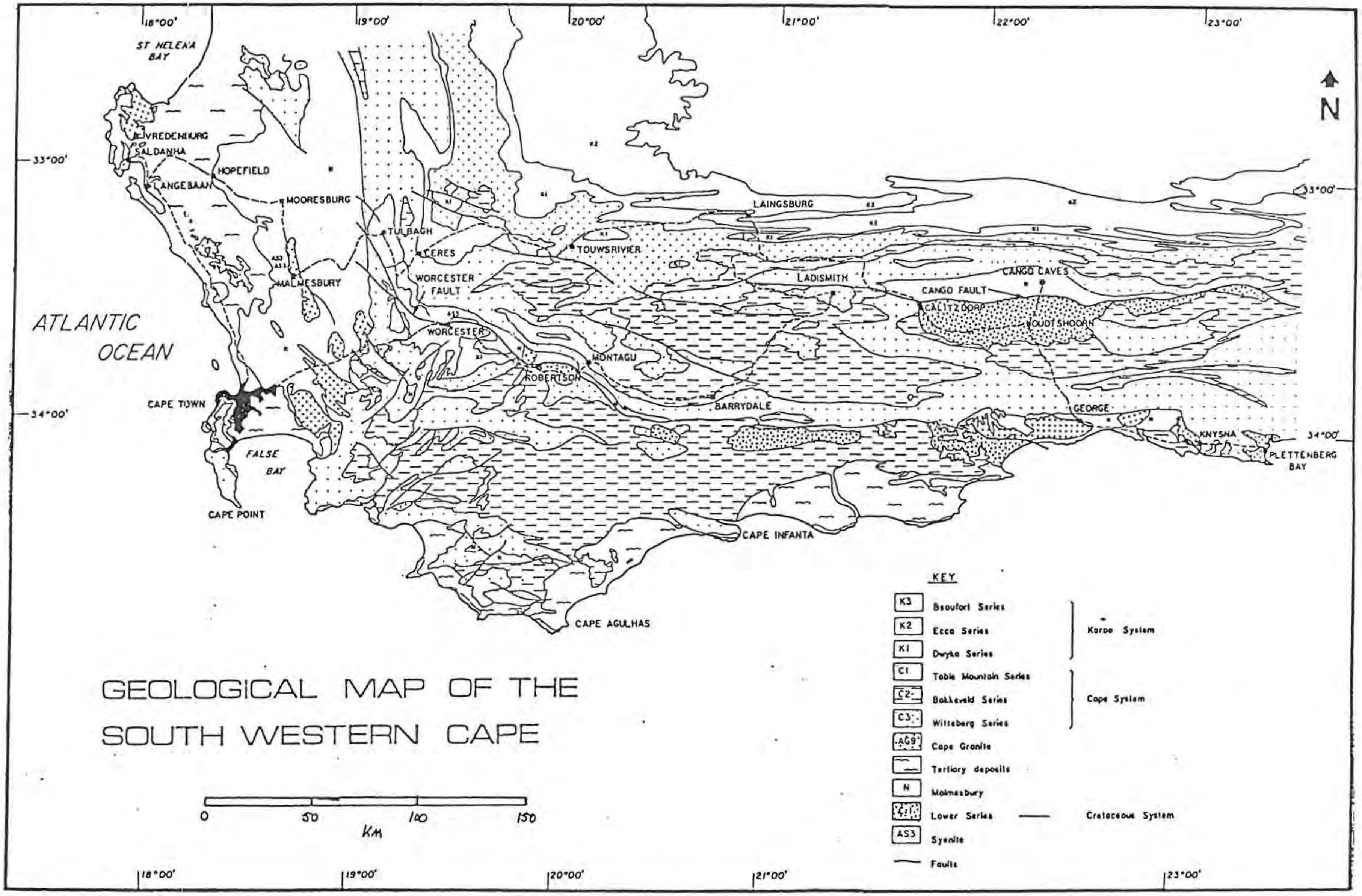
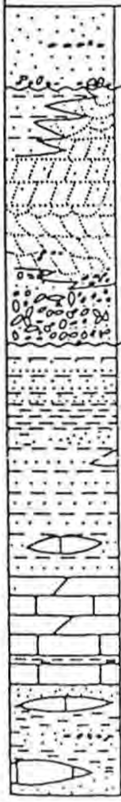


Figure 3.12: Geological map of the S.W. Cape (Hartnady et al., 1974)

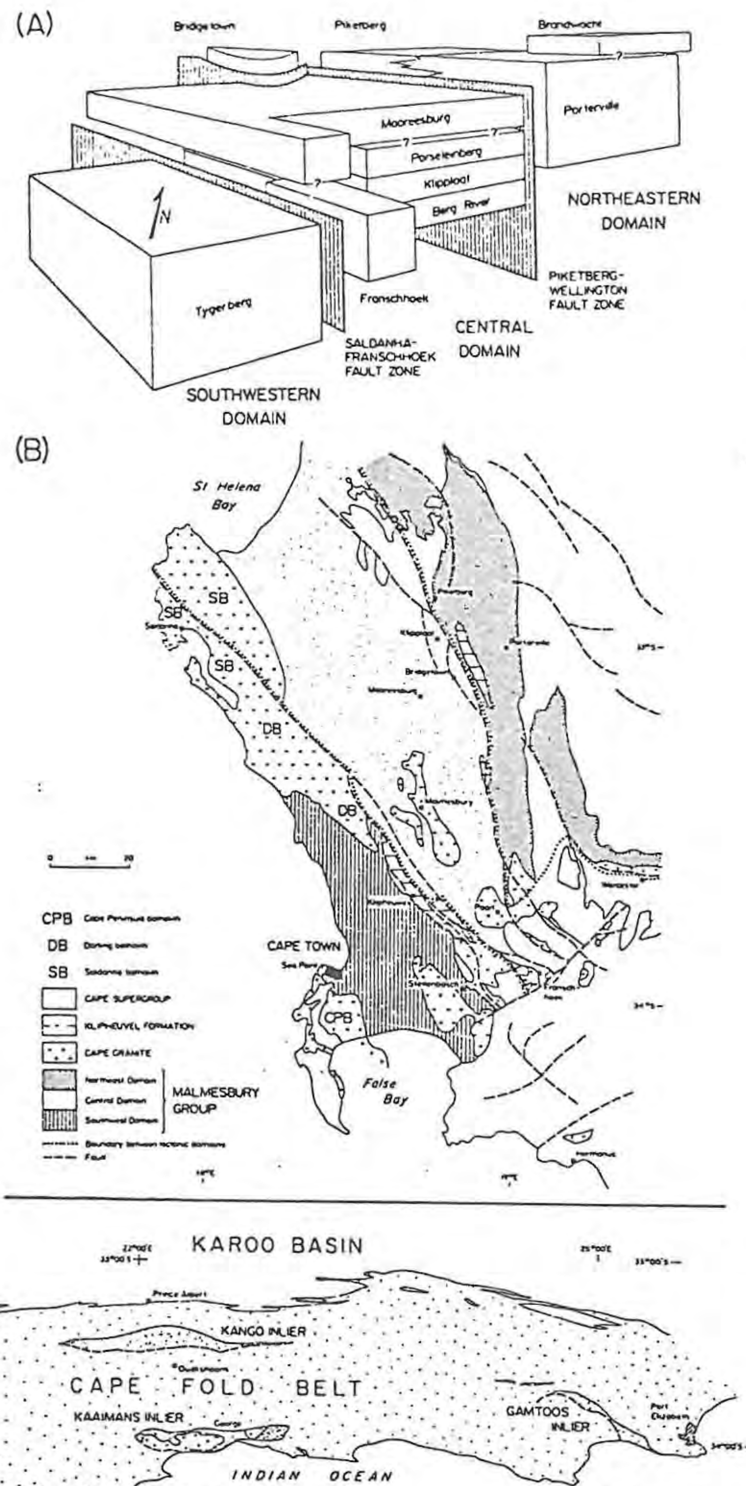
Table 6: Stratigraphy of the Malmesbury Group, Western Saldanian Province (Tankard et al., 1982)

Southwestern Tectonic Domain	Central Tectonic Domain	Northeastern Tectonic Domain
Tygerberg Formation: Alternating shale and fine-grained immature sandstone (regularly bedded; structures include graded bedding, cross-beds, load casts, slump structures); rare limestone and conglomerate. Intermediate volcanic rocks brecciated).	Franschhoek Formation: Pebble to boulder conglomerate, arkose, subgraywacke, quartzite, and slate (lenticular geometry). Bridgetown Formation: Basic rocks, chert, dolomite, limestone, shale, and phyllite. Swartland Subgroup Moorreesburg Formation: Irregular alternation of wacke and shale. Porseleinberg Formation: Well-laminated quartz–muscovite–sericite schist with interbedded phyllitic chlorite–muscovite schist. Klipplaat Formation: Quartz schist with mudclast breccia; mica schist, basally quartzitic and cherty; rare lenses of limestone. Berg River Formation: Alternating lenticular mica schist and quartz schist, immature wacke, thin limestone, quartzite.	Boland Subgroup Brandwacht Formation: Shale, graywacke with graded bedding, conglomerate; very conglomeratic basally; metabasites in upper part. Porterville Formation: Phyllitic shale and fine- to medium-grained graywacke in irregular units, conglomerate up to 90 m thick, limestone basally, quartzite, chert. Piketberg Formation: Coarse-grained feldspathic sandstone, graywacke, conglomerate, interbedded phyllite especially in south, rare feldspathic limestone.

Table 7: Pre-Cape Saldanian Basins of the southern Cape coast (Tankard et al., 1982)

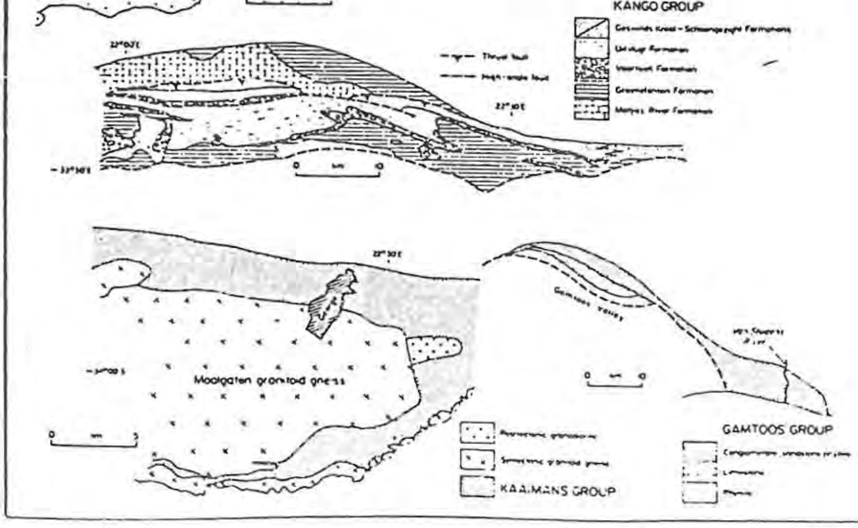


Kango Group	Gamtoos Group	Kaaimans Group
Unit 7: Schoemans Poort Formation, 600 m Subarkosic sandstone with bands and lenses of conglomerate		Unit 7: Homtini Formation Alternating slate, phyllite, graywacke; lenticular interbeds of coarse-grained sandstone and conglomerate
Unit 6: Gezwinds Kraal Fm (in W), 1500 m Schoongezight Fm (in E), 1100 m Pebbly sandstone grading westward into alternating sandstone and shale		
Unit 5: Uitvlugt Formation, 1000 m Subarkosic sandstone, cross-bedded, lenses of conglomerate and mudstone	Unit 5: Arkosic sandstone, medium to very coarse grained, cross-bedded, mudstone and limestone interbeds; discontinuous basal conglomerate; phyllitic upper parts	Unit 6: Victoria Bay Formation, 85 m Coarse-grained arkosic sandstone, lenticular conglomerate interbeds, occasional thin beds of marble
Unit 4: Vaartwell Formation, > 1000 m Conglomerate with sandstone interbeds		
Unit 3: Huis Rivier Fm, (in W), 1500 m Shale grading upward into graded quartz wacke with sole structures	Unit 4: (Kaan) Dark limestone (sparite), massive beds separated by thin beds of phyllite, quartzite, and calcareous sandstone	Unit 5: Soetkraal Formation, 300 m Phyllite and schist, locally rhythmically alternating; some hornfels and sandstone interbeds
Unit 2: Groenefontein Formation, 2400 m Alternating sandstone and shale, parallel bedded, sporadic limestone lenses	Unit 3: Phyllite and shale, and coarse-grained arkosic sandstone, thinly bedded; lenses of very coarse-grained granular sandstone and conglomerate	Unit 4: Skaapkop Formation, 260 m Quartzite, medium to coarse grained, phyllite medially. Phyllite (6m) basally Unit 3: Sandkraal Formation, 300 m Quartz-sericite schist, arenaceous, parallel bedded
Unit 1: Matjies River Formation, 2300 m 1300 m dolomitic limestone (oolitic) with shale, siltstone, and sandstone interbeds in lower part	Unit 2: (Kleinfontein) Dark dolomitic limestone (sparite), thin interbeds of calcareous and pyritic carbonaceous shale	Unit 2: Saasveld Formation, 600 m Andalusite-bearing biotite schist, hornfels, argillaceous mica schist
1000 m shale and graywacke, subordinate conglomerate lenses and limestone interbeds	Unit 1: Phyllite, with lenses of pyritic carbonaceous shale	Unit 1: Silver River Formation, 1300 m Chlorite-rich phyllonitic psammite and semipelite, some thinly bedded graywacke, rare limestone
(base unknown)	(base unknown)	(base unknown)



<u>Figure 3.13</u>:

The Malmesburg Group, western Saldanian Province:
A) Block diagram to show stratigraphic relations between the three tectonic domains.
B) Geologic map of the Malmesbury Group and younger Cape granites (Tankard et al., 1982)



<u>Figure 3.14</u>:

Geologic maps of the miogeosynclinal Kango and Gamtoos Groups and the eugeosynclinal Kaaimans Group in the eastern Saldanian Province (Tankard et al., 1982) (conglomerates) and less maturity. Lower successions have thin limestones, whilst stratigraphically higher alternating wacke and shale sequences have associated intrusive and extrusive basic volcanics. Deposition of the uppermost heterogeneous Franschhoek Formation appears strongly controlled by the prevailing tectonic activity, which is likely to have involved basin subsidence along numerous listric faults. A continental margin setting, with gradual subsidence due to extensional tectonism, can be invoked to account for the salient features.

The northeast domain is dominated by variable shale and greywacke units, with conglomerates (up to 90m thick), limestone, and chert. There appears evidence for both marine (shelf) and alluvial environments of deposition (Stump, 1976; in Tankard et al., 1982), with the former dominant and attesting submarine fan sedimentation.

In the eastern Saldanian Province (southern Cape region) the orogeny was also accompanied by deposition of thick sequences of sediments in marginal and rifted basins, now preserved as three elongate inliers in the Cape Fold Belt containing the Kango, Gamtoos and Kaaimans Groups (Fig. 3.14). The Kango and Gamtoos sedimentary successions have 'miogeosynclinal' characteristics, whereas the Kaaimans Group is 'eugeosynclinal' (Tankard et al., 1982). The sedimentary suites include terrigenous clastic and carbonate rock types that have been subjected to low grade regional metamorphism. They are tentatively correlated with the Malmesbury Group sediments, and were deposited in a long, relatively narrow embayment (a rifted east-west basin) subject to variable marine transgression/regressions (Le Roux and Gresse, 1983; Le Roux, 1983).

Metamorphism and granite plutonism

The regional metamorphic grade of the Saldanian rocks varies from nonmetamorphosed rocks that have undergone only diagenesis, through very low grade rocks that have undergone burial metamorphism to low-grade rocks metamorphosed to greenschist facies (Hartnady et al., 1978). The associated tectonic fabrics indicate that the metamorphism occurred during the pre-Cape Saldanian orogeny (Tankard et al., 1982).

The orogeny is marked by the emplacement in the southwestern Cape of at least 25 syn- to post-tectonic high level granitoid plutons, correlative to the post-tectonic Kuboos-Bremen intrusives in the Gariep and

Richtersveld Provinces. The distribution of the more important plutons is illustrated in Fig. 3.15. There is widespread evidence of multiphase intrusion, supported by a range of Cambrian radiometric dates from 600Ma to 500M (Allsop and Kolbe, 1965; Burger and Coertzee, 1973; Schoch et al., 1975; Schoch and Burger, 1976). Pb-Pb isochrons, which may be reasonable indicators of primary ages of intrusion, give ages of 522 $\pm$ 12ma for the Saldanha rhyolite porphyry (Schoch and Burger, 1976), 530 $\pm$15Ma for the adjacent Bredenberg granite, and 600 $\pm$20Ma for the Hoedjiespunt (part of the Saldanha Batholith), Darling and Cape Peninsula granites (Schoch et al., 1975).

The plutons are exposed as upfaulted or upfolded inliers in the Cape Fold Belt, and comprise two distinct geographic groups

i) an eastern group, comprising a multiphase batholith at George (Fig. 3.14, Krynauw, 1983), with east-west elongation and penetrative (largely Cape age) tectonic fabrics, and

ii) a southwestern group which are elongated NW with variable deformation (Fig. 3.12).

The exposure pattern of the basement granites is largely controlled by the fold trend in the Cape cover (see Fig. 3.12), although these trends are also manifested within the Malmesbury country rocks, and reflect the syn- to post-tectonic emplacement of the plutons. The timing of emplacement relative to the metasediments is also substantiated by shear belts which transect and border some of the intrusions, although there is good evidence that there has been repeated reactivation of these fault zones (e.g. the Saldanha-Franschhoek and Piketberg-Wellington fault zones).

The Cape granites have been divided into two groups on the basis of their major-element geochemistry (Schoch et al., 1977). An older group displays magmatic differentiation trends (hornblende-biotite to biotite granite) and coarse-grained and megacrystic to fine-grained textures, with partial foliation. Contamination due to assimilation of metasedimentary rocks is evident, with the presence of cordierite and muscovite, and well as an antipathetic correlation between feldspar and biotite (Schoch et al., 1977). Low- to medium-grade contact metamorphic aureoles are discernable. A far less voluminous (5-10% of the suite), younger group of granites have leucogranite compositions, devoid of differentiation trends, and invariably intrude the older granites.

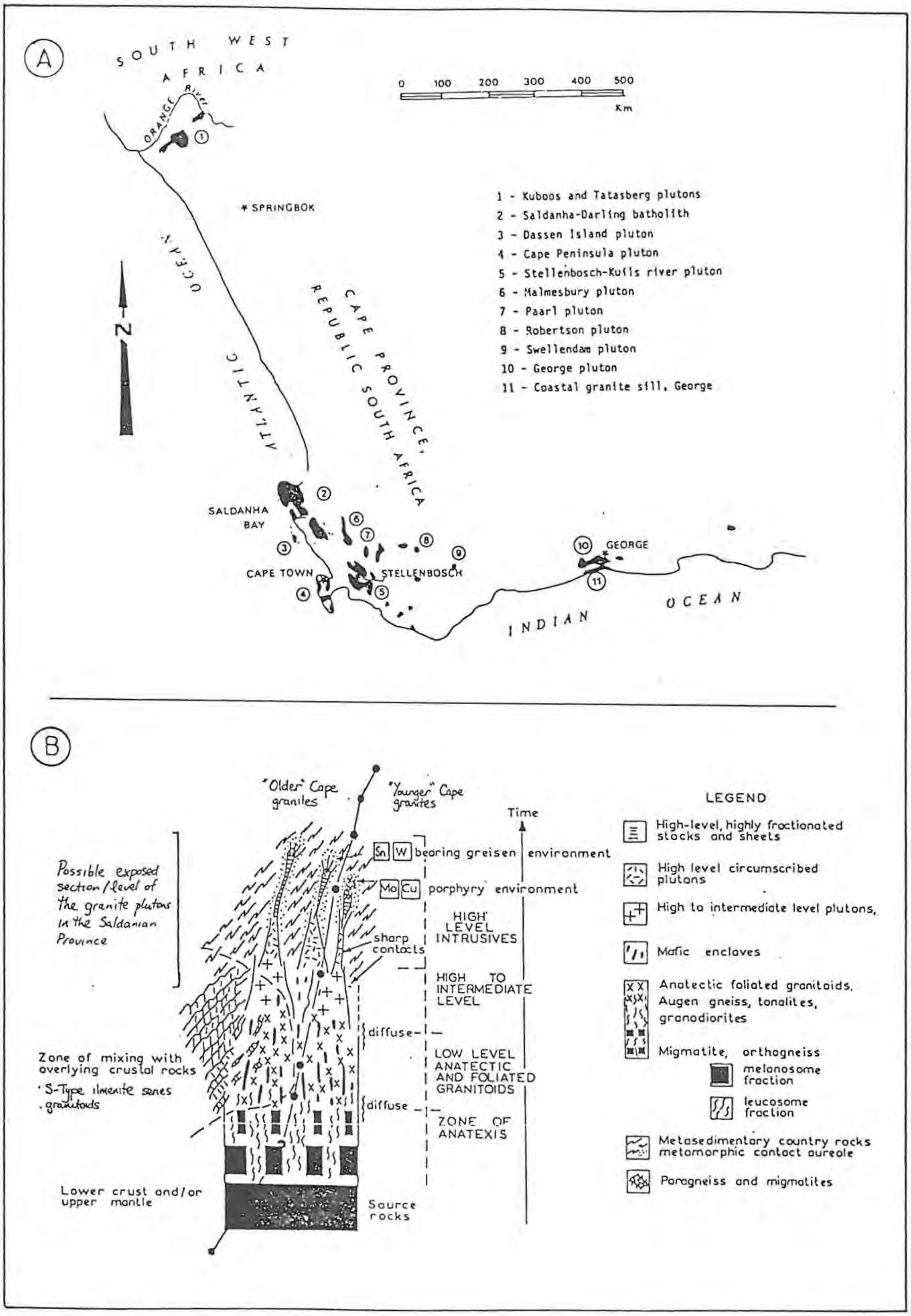


Figure 3.15A) Distribution of Cape granites (Schoch et al., 1977)

B) Schematic diagram showing anatectic processes, granitoid genesis, and related mineralisation (Pirajno, 1983)

Siegfried et al. (1984), reporting on the Paarl pluton, distinguished six granites on textural and mineralogical grounds:- porphyritic granite, medium-grained granite, fine-grained granite, quartz porphyry, aplogranite and a hybrid granite. The latter is interpreted as restitic material indicative of a more basic parental magma. Petrographic and major and trace element geochemistry and modelling invoke a restite-mixing origin for the porphyritic and medium-grained granites, whilst fractional crystallisation is proposed for the fine-grained granite. Associated quartz veins, quartz porphyries, pegmatites and aplites are all late stage intrusions.

A complete analysis of published geochemical data and petrographic descriptions of the Cape granites is out of the scope of the present work, however perusal of whole rock data presented by Kolbe (1966), Schoch et al. (1977), and Siegfried et al. (1984) indicates the likely presence of magmas with affinities to both S-type ilmenite series (the "older" granites) and I-type magnetite series (the "younger" granites). Field relationships and radiometric dating suggest that the granites are largely formed by anatectic processes (Fig. 3.15B). This would account for the clear evidence of granite/metasediment mixing, as well as the younger intrusives that have been emplaced as discrete plutons into the larger earlier batholithic granites. It is speculated here that different styles of mineralisation could be associated with the two phases of granites:-

i) Sn(W) in greisen environments related to the early ("S-type ilmenite series) granites, and
ii) porphyry Cu-Mo-(W, Au) with the later (I-type magnetite series) granites. The writer is unsure if in fact the Saldania Province has been fully assessed in this respect (e.g. Sn-W, Cu-Mo-W metallogenic provinces).

Of interest here is the close simularity of the above scenario to that documented in New Zealand (Pirajno, 1983; in press, Pirajno and Bentley, in press), where such metallogenic provinces have been distinguished. Interestingly, so-called I-type granites have evolved through anatectic processes, but appear in some cases to have gained S-type chemistry through emplacement at high structural levels where interaction with H_2O bearing country rocks (e.g. metasediments analagous to the Malmesbury Group) evolves high silica types, leading to post-magmatic alteration of the greisen type, producing Sn-W deposits (Pirajno, 1983).

One can further speculate that the granites may have formed from "anorogenic asthenospheric diapirism" type magmatism related to early crustal extension and thinning, (giving the S-type granites) followed by closure/compression resulting in limited subduction (and I-type granites). The overall scenario is further complicated by Cape age thrusting, which has generally modified the tectonic emplacement of the various plutons.

3.3.2 Damara Province

The Damara Province forms a part of the Pan African mobile belt system which largely fringes, but also transects the African continent (Fig. 3.11). Recent overviews on the tectonic evolution of the Damara Province (Martin, 1983a, b; Tankard et al., 1982) point to a model initially comprising a triple-junction (see Burke and Dewey, 1973; Burke, 1977, 1978), with two arms forming the present NW trending west coast of Namibia and eastern coast of Brazil, and the third an aulacogen projecting NE into the cratonic interior. This intracontinental branch is well exposed across its 300-400km width, and thus is the better studied part of the Damara Province. Several structural and lithostratigraphic domains are distinguished (Fig. 3.16) and metamorphism ranges from low-grade at the southern and northern margins to high-grade in central areas, where voluminous granitic plutonism with attendant U, Sn and W mineralisation is preserved. The following discussion outlines aspects of the structural subdivision, sedimentary, magmatic and metamorphic history of the Province.

Structural subdivision

The Damara Province displays a strong structural and metamorphic asymmetry. Four tectonic zones, all with contrasting geotectonic settings, are recognised in the intracontinental branch (Figs. 3.16, 3.17; Martin and Porada, 1977; Porada, 1979; Tankard et al., 1982; Martin, 1983): the Northern Platform zone (approximately 80km wide), a Northern zone ("transitional zone" of Martin and Porada, 1977; approximately 40km wide), the Central zone (approximately 160km wide), and the Southern marginal zone (approximately 150km wide).

The Northern Platform zone, which incorporates the Otavi Mountainland, overlies the rigid and stabilised Congo craton, which shielded it from orogeny. Tectonism was restricted to upright folding with slight

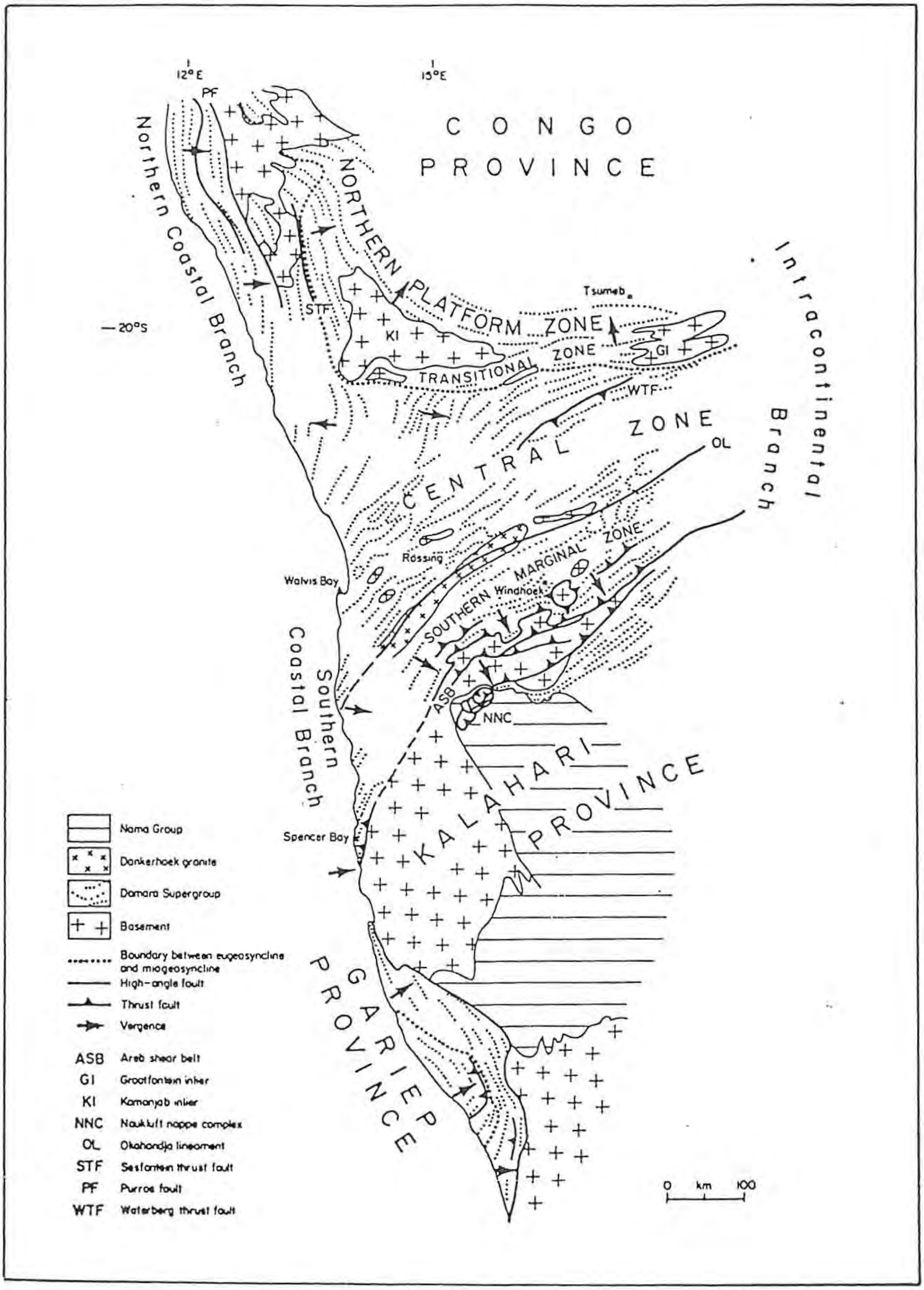


Figure 3.16: Tectonic map of the Damara and Gariep Province, showing tectonic zones, eugeosyncline-miogeosyncline boundary, pre-Damara basement inliers, and southern marginal thrust zone (Tankard et al., 1982).

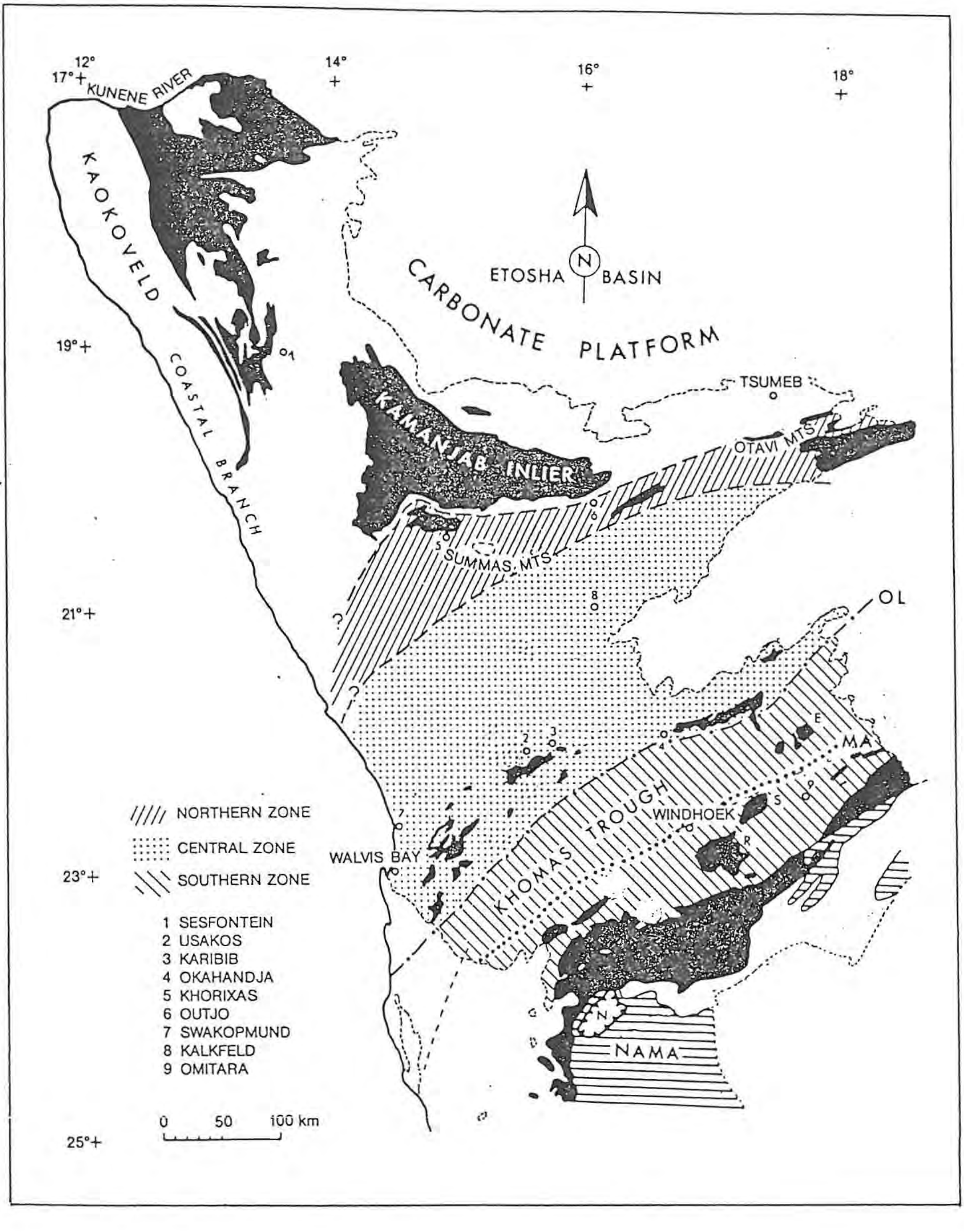


Figure 3.17: Main structural zones of the intracontinental arm of the Damara Province (structural zones of the coastal branch in the Kaokoveld are not shown) - Black = pre-Damara basement; OL = Okahandja Lineament; MA = Matchless Amphibolite; N = Naukluft Nappe Complex (Martin, 1983a)

northward vergence and weak metamorphism. The Northern zone, incorporating the Kamanjab and Grootfontein basement inliers, appears to have acted as a kind of hinge zone between the carbonate platform in the noeth and the subsiding trough of the Central zone to the south (Fig. 3.18). The Northern zone has undergone greenschist facies metamorphism and polyphase deformation, resulting in tight folds of variable axial trend (Martin and Porada, 1977).

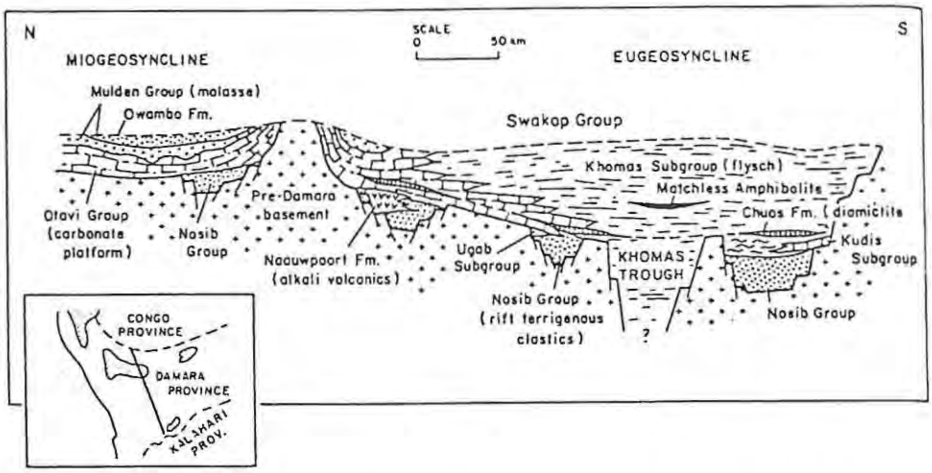
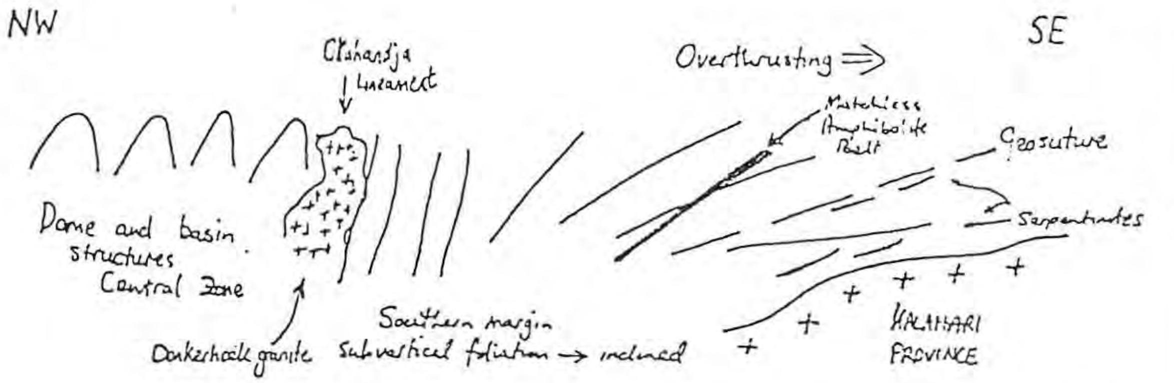


Figure 3.18 Schematic stratigraphic cross section across the Damara province showing the control of the stratigraphic elements by extensions (rift) faulting (Tankard et al., 1982)

The structure of the Central Zone is characterised by domal antiforms, elongated along the main SW-NE trend, and by synforms containing Kuiseb schists (see below) and syn- to late-tectonic granite (Salem-type). The structures are largely SW verging folds produced during an M_2 phase of folding, and interpreted to form large flattened tubes or sheath folds, commonly over 10km in wavelength (Coward, in press). The intense deformation is considered by Coward (op. cit.) to be due to a large low-angle shear zone, converse to other geodynamic models (see subsection 3.3.3). The domal structures are attributed to diapiric deformation due to the uprise of late stage granites and pegmatite concentrations.

The Okahandja Lineament (OL, Fig. 3.17) forms a fundamental tectonic boundary which divides the Central and Southern Marginal zones. The monoclinal flexure has been invaded, and is largely obscured, by the post-tectonic Donkerhoek granite (520-500Ma; Martin, 1983a). A distinct transition occurs from "dome and basin" structures of the Central zone to steep foliations in the Southern zone (Fig. 3.19). Progressing SE this is overprinted by a younger foliation, which gradually becomes inclined to a low angle ($\pm 30°$) NW. The tectonic style is characterised by intense strain fabrics, south- to southeast verging overturned to recumbent early folds, and imbricate interslicing of Damara and basement related to overthrusting (Tankard et al., 1982).



Sketch section NW-SE of the fundamental structural elements within the southern marginal zone. Note that the Southern zone is marked by steep foliation, whereas to the south it is overprinted by a younger foliation, gradually becoming inclined ≃ 30° to the NW. Major thrusting, tectonic slicing and crunching of crustal material has been interpreted in this area.

Figure 3.19: Sketch of the fundamental structural elements within the Southern marginal zone

The southern margin of the Damara Province is marked by a southeast verging thrust zone some 400km long (including the Areb shear belt, ASB Fig. 3.16), which represents the culmination of horizontal tectonics in this area (Coward, in press). Allochthonous Damaran thrust nappes (Naukluft Nappes), comprising mostly Swakop Group metasediments with subordinate Nama deposits were emplaced as intensely deformed rocks on older undeformed granitoids and subsequently infolded with the basement (Tankard et al., 1982; Hartnady, 1979). Recent investigations (Martin et al., 1983) indicate that the root zone of the Naukluft Nappe Complex was in the thrust belt forming the southern margin of the Damara Province, probably originating as a mobilised mylonitic sludge with a high content of hot saline fluid. The saline fluid is thought to have been derived from evaporite beds in the lower Damaran Duruchaus Formation, which fills a deep basin approximately 90km to the north-east.

Sedimentation

Illustrations of the stratigraphic nomenclature and proposed correlations of stratigraphic subdivisions, intrusions, and mineral ages, metamorphism and deformation phases are given in Table 8 and Figs. 3.20, 3.21.

Table 8: Lithostratigraphy of the Damara Sequence (Martin, 1983a)

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	HUTTENBERG	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	ROSSING	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)

The earliest sediments generally show shallow water or terrestrial sedimentary affinities, and were laid on pre-Damara basement (to about 2.0Ga, Jacob et al., 1978) preserved as large inliers scattered throughout the Province. Sedimentation was initiated 1000-900Ma ago with infilling of fault-bounded troughs in rifted continental crust. These troughs (Nosib Grabens) are about 50-70km wide and 200km long, and have been recognised trending NE within the intracontinental branch as well as in the northern coastal branch (Fig. 3.22, Porada, 1983). Upper zones of the Central and Southern rifts locally host synsedimentary Cu, Pb, and Ag mineralisation (e.g. Namib Pb, Oamites Cu-Ag, Klein Aub Cu, and Witvlei Cu).

Around 830Ma further rifting and extensional tectonics widened the area of deposition, resulting in the coalescing of these troughs, and sedimentation overstepped the basement highs. The 'mio-eugeosyncline' diachotomy began, with the Otavi Group platform succession accumulating on the gently subsiding foreland of the Congo Craton, and flysch-type clastics of the Swakop Group deposited in a deepening trough (Khomas) to the south (Table 8, Rift stage Figs. 3.18, 3.20). In the Northern zone the Otavi Group unconformably overlies the Nosib Group and basement (Table 8), and consists of a thick (to 7km) platform succession of stromatolitic dolomitic limestone with subordinate marl near the top.

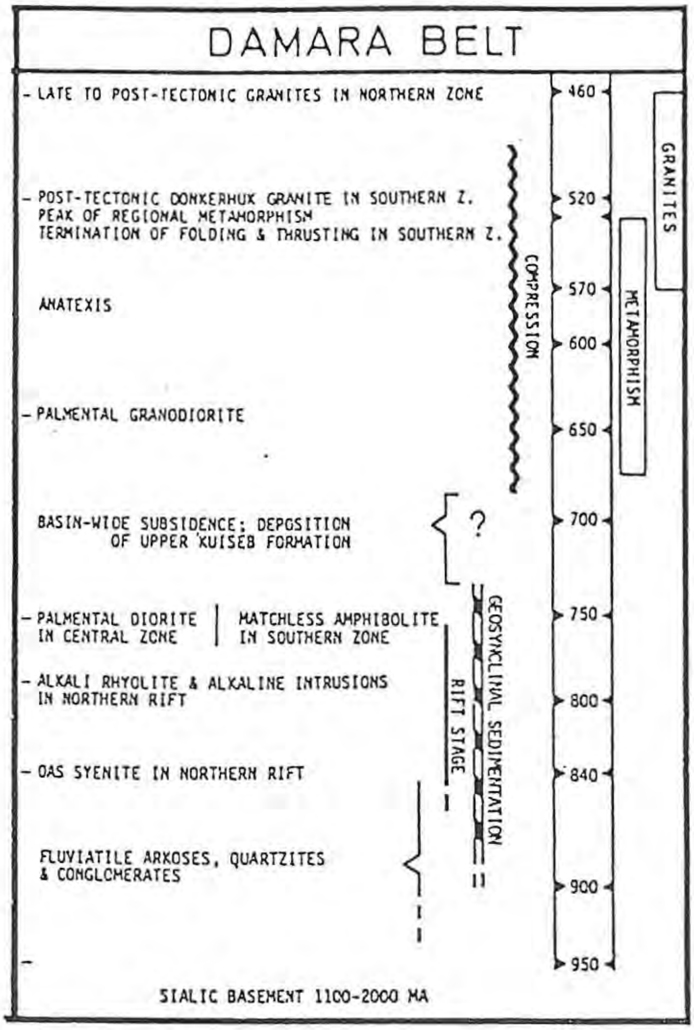


Figure 3.20: Generalised diagram of the evolution of the Damara Belt (Martin and Eder, 1983)

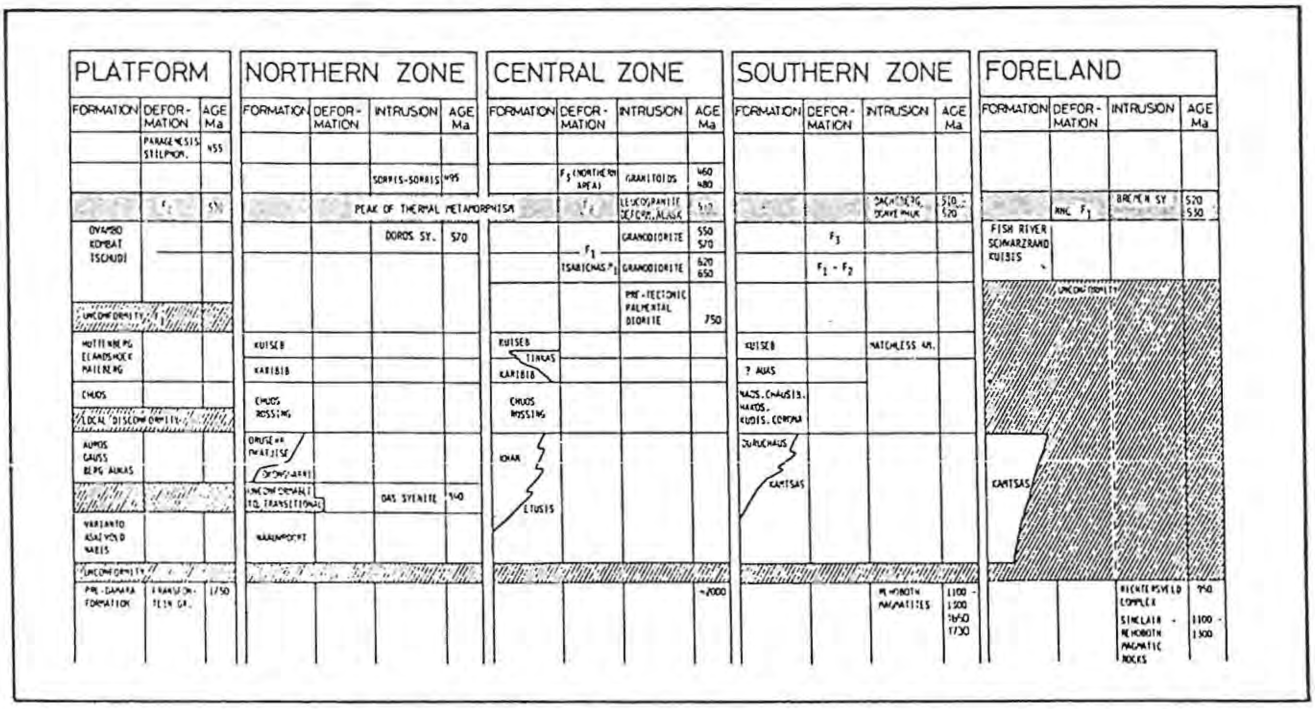


Figure 3.21: Tentative correlation of stratigraphic subdivisions, intrusions and mineral ages, metamorphism and deformation phases (Haack and Martin, 1983)

These sediments accumulated in a stable and largely shallow marine environment 830-760Ma, and host the Tsumeb-Kombat Cu-Pb-Zn mineralisation. The carbonate platform underwent gentle folding and erosion, evidenced by an angular unconformity, and was later covered by molasse-type sediments of the Mulden Group around 550Ma.

In the Central region of the Province the oldest sediments are coarse probably fluviatile clastics (Etusis Fm) overlain by quartzites, schists and impure carbonate now preserved as calc-silicates (Khan Fm), which together comprise the Nosib Group. They are thought to be broadly coeval with a suit of potassic lavas (Naauwpoort Volcanics) and associated syenites and carbonatites exposed along the northern margin (Martin, 1983a; Hawkesworth and Marlow, in press).

There is a slight discordance before passing up into a variable sequence of carbonates, quartzites, conglomerates, pelites and locally graphitic schists (Table 8). This is the start of a rift-trough (or geosynclinal sequence) which reaches its greatest present-day thickness (12-14km) along the margins of the Province (Martin, 1983a). Carbonate rocks are usually present, but tend to be dominant in the north (above) and subordinate to clastic in the south, where the bulk of sediments are turbiditic biotite schists (Kuiseb Fm) within the Khomas trough. The deepest parts of the Khomas trough are preserved in the Southern Marginal zone, where great thicknesses of the Kuiseb metapelites are tectonically thickened by recumbent folding and imbricate thrusting (Tankard et al., 1982).

Amphibolite metavolcanics (the Matchless Belt and Schlesian Line) occur stratigraphically high in the schists. These intensely deformed rocks apparently have oceanic (MORB) chemical affinities, and have been interpreted by some authors (Hartnady, 1979; Burke et al., 1977) to represent obducted oceanic crust, constituting a suture line (i.e. resultant from subduction and collision). There is still conjecture pertaining to this hypothesis, interestingly constrained by paleomagnetic evidence, (McWilliams and Kröner, 1981,) and also the fact that the Matchless Amphibolites have associated Besshi-type Cu-Zn mineralisation (not Cyprus-type, of ophiolitic affinity).

The final sedimentation cycle of the Damara orogen involved the deposition of continental sediments (molasse-type) derived from pre-existing Damara rocks. These comprise the Nama and Mulden Groups in the south and north respectively, which unconformably overlie the Kuiseb Fm, and are now only weakly metamorphosed.

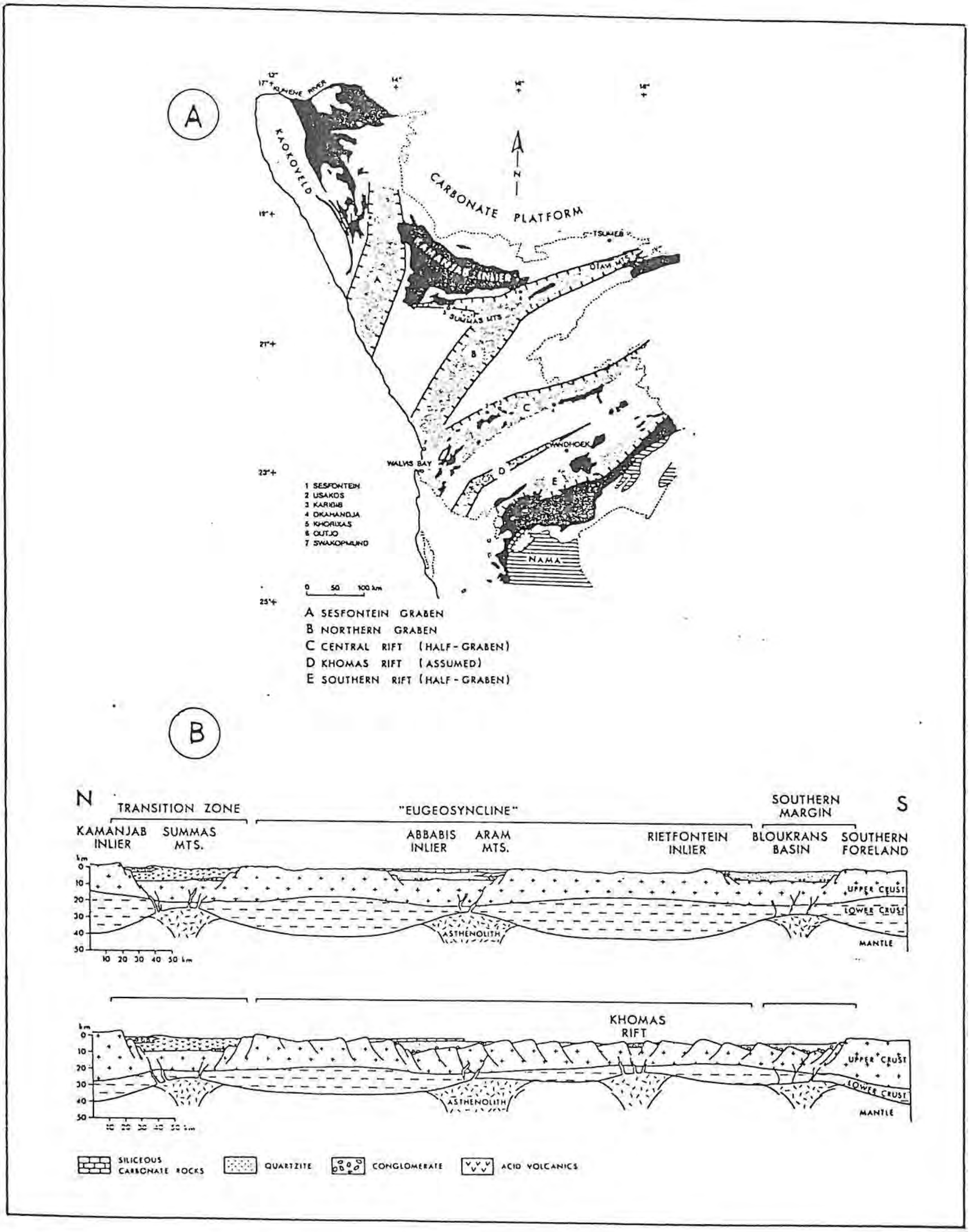


Figure 3.22: A) Supposed positions of rift systems during the early geosynclinal development of the Damara Province.

B) Schematic sections across the Damara Orogen showing alternative models of the crustal structure during the early rifting stage. Top, after Martin and Porada (1977). Bottom, model proposed by Porada (1983) (Porada, 1983)

Magmatism

Sedimentation in the "geosynclinal" stage has already been discussed, and is related to extensional tectonism and aulacogen formation (Martin and Porada, 1977; Martin 1983a; Porada, 1983). Significantly the magmatic activity that occurred during this stage was non-granitic, essentially consisting of syenites and diorites. Probably as early as 680Ma (Fig. 3.20) compressional tectonics became predominant, forming NE-SW trending fold structures, and leading to crustal shortening and thickening, with concurrent heating and metamorphism. Magmatism after the change from extensional to compressional tectonism is solely granitic (Haack et al., 1983; Haack and Martin, 1983a).

Generally granites in the Damara Belt are not known petrographically or chemically in great detail. The Central zone of the Damara Province is dominated by granitic rocks varying from diorites to granites and highly potassic alaskites. Four main types are generally recognised: Salem granites, Red granites, leucogranites and alaskites. Recognition of these types is largely based on macroscopic characteristics, and often little geochemical or isotopic data is known. Thus, there are many granites which, on the basis of their description in the literature, don't fit readily into any of these four categories. Fig. 3.23 illustrates the distribution of the granitic rocks relative to Pre-Damara basement.

Pre-Damara granitic basement is exposed at the northern and southern margins and in numerous inliers within the intra-continental branch of the Damara Province. Radiometric ages (U/Pb) indicate that those in the north and centre of the belt were formed 2Ga ago (Burger et al., 1976; Jacob, 1978), whereas those to the south are approximately 1.1Ga old. Mafic, acid volcanics, and granitic magmas were produced then. In the Central zone the basement consists of augen gneisses, metasediments and some metavolcanics.

In general terms the Salem and Red-granites exhibit a range of ages - the Salem type granites are now known to have intruded in a number of pulses from about 550Ma to 460Ma (Puhan, 1983; Martin, 1983a). Melts of this type were emplaced over a time span of about 100Ma, outlasting the time of intrusion, 520Ma ago, of the post-tectonic Donkerhoek Granite plutons into the Okahandja Lineament and the adjoining Southern zone (Martin,

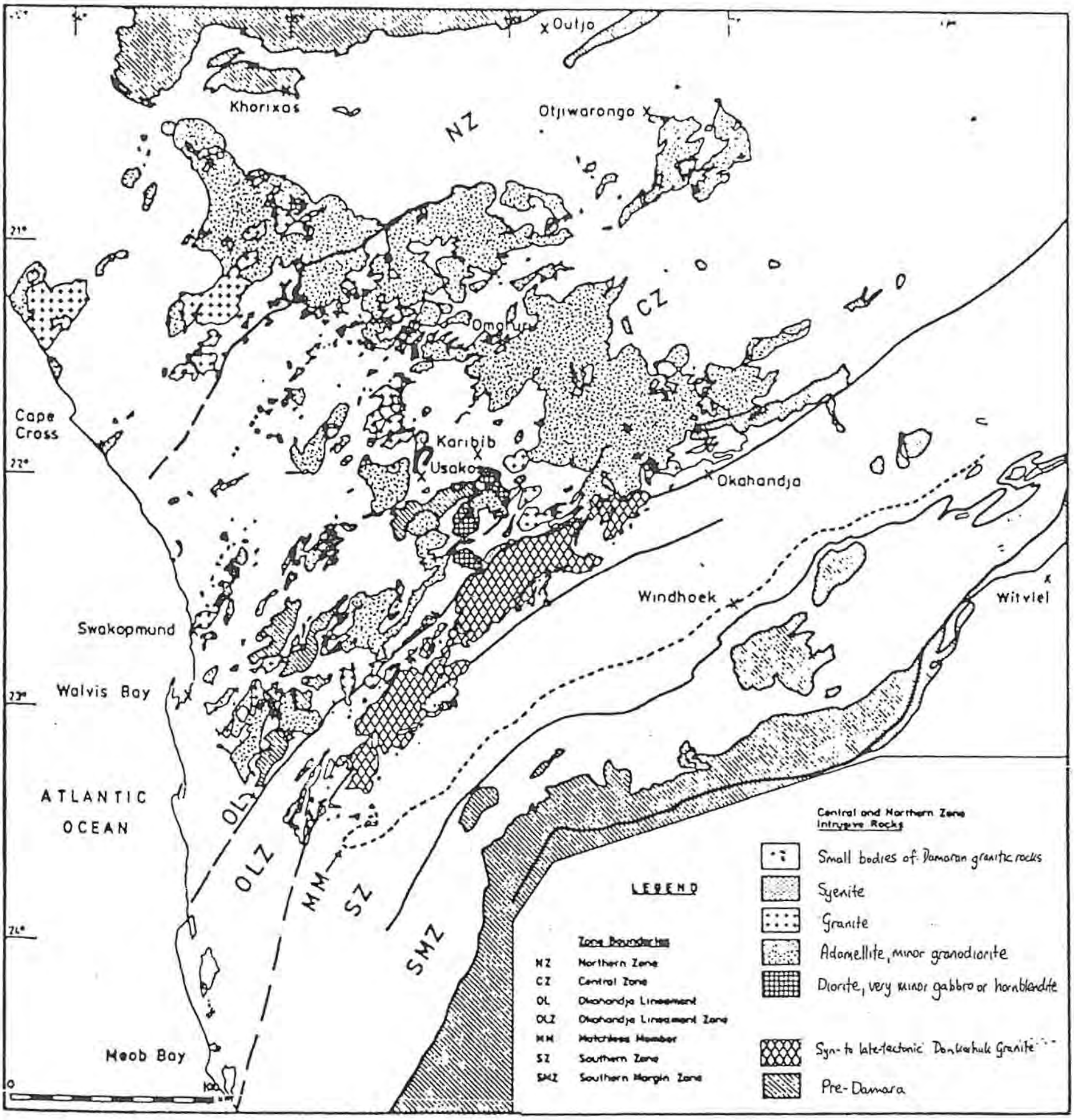


Figure 3.23: Distribution and types of Damara orogen plutonic rocks in the CZ and NZ of the Damara Province (after Miller, 1981)

1983a). Undeformed Red granites in the Northern, Central and Southern marginal zones give ages of 510 to 495Ma (Hawkesworth et al., 1981). The Salem and Red granites are both syn- and post-tectonic, where as the leucogranites and alaskites tend to be younger and post-date the last major regional deformational fabric (F_2, 510Ma, according to Haack & Martin, 1983; Fig. 3.21). Salem-type granites range in composition from diorite to adamellite (Miller, 1973), and often occur as discrete intrusive centres, or batholiths. Macroscopically they contain, in a biotite-rich matrix, large alkali feldspars which frequently enclose rims of biotite and porphyroblastic garnets (Haack et al., 1983). An origin by in situ granitisation or melting was formerly invoked for the Salem Suite, largely because in the area S and SE of Karibib these granites occur preferentially in synclinal structures above the level of the Chuos Formation and are underlain by the thick Karibib marble which usually dips towards the centre of the plutons (Jacob, 1978; Smith, 1965). Melts may have been generated by anatexis of the Khomas Subgroup metasediments at depth, with subtle intrusion virtually confining the granites to the stratigraphic level of the Khomas Subgroup (Jacob, 1978). Further studies to the north (Miller, 1973) have also elucidated a possible source from Damara metasediments. However, the pulse type intrusion of the Salem-type granites over a period of about 100Ma (above), and large differences in initial Sr-isotope ratios and ages provide evidence against a common rock source for the "suite" (Haack et al., 1983). They are now considered early syntectonic, and to have generated from within the crust (not in situ) even though feeders are not visible (R. Jacob, pers. comm., 1984). Leucogranites have been recognised as the youngest member of the suite, although they lack "Salem characteristics" (Jacob, 1978; Haack et al., 1983). The Salem-type granites appear to be the principal hosts to Sn-W mineralisation in the Central and northern Central zone, coupled with reactivated granitic intrusions of Karoo age (see below). At present there is no published data concerning the Sn-W contents of these granites, or even an assessment as to whether there is any metallic specialisation (e.g. as there is in U in the Central zone, see below).

In contrast to the Salem-type granites, the Red granites are characterised by a low percentage of ferromagnesian minerals and high K-feldspar (Plate 2A). They occur as domes or in dykes and lit-par-lit intrusions at the level of the Nosib Group and basement beneath the Karibib marbles. They are thought derived from a combination of partial melting of the Nosib Group and reactivation of the basement (R. Jacob pers. comm., 1984).

Significantly the basement gneisses are abnormally radioactive (up to about 50ppm U_3O_8). Outcrops of sheared and foliated basement augen gneiss were observed, with dykelike streaks and veins of Salem-type granite intruding shears, whilst blotches of red granite can be observed within and transgressing the foliation (Plate 2B). These features indicate the initiation of generation of the Salem- and Red granites. They are post-dated by alaskites, characterised by high alkali contents (5-6.5%K_2O, about 3.0%Na_2O), and an anastomosing and vein-like style of intrusion. The alaskites are significant for their economic uranium mineralisation (Rossing Mine, Valencia and Ida Prospects). The alaskites can be seen locally to be derived from in situ melting of both the pre-Damara basement gneisses and Nosib metasediments, and coupled with Sm-Nd systematics, this indicates that of all the Damara granites the alaskites were derived from high structural levels (Hawkesworth and Marlow, in press).

The emplacement of the alaskitic uraniferous melts is largely restricted to the western part of the Central zone (Khan-Swakop area) where high-grade metamorphism and anatexis prevailed. Dating of the emplacement of the uraniferous alaskites is problematical, as they are poor in Sr, are small bodies, are very rich in feldspar with only subordinate biotite, and U/Pb dating is complicated by 2° uranium mineralisation (Haack and Martin, 1983). Published Rb/Sr ages are around 460Ma for the Rossing Mine (Kröner and Hawkesworth, 1977), 542 $\pm$ 33Ma for Ida Prospect (Marlow, 1982), whilst alaskites at Goanikonites gave monazite and uraninite ages of 508 $\pm$ 2Ma (Brique et al., 1980, in Haack et al, 1983).

The Donkerhoek Granites (520Ma, Fig. 3.23) are two mica granites and are confined to a zone $\pm$250Km long and 30km wide confined to the Okahandja Lineament zone. These granites intruded not long after the peak of regional metamorphism (see Fig. 3.21) into a still very hot environment (Haack et al., 1983) and thus are essentially late to post-tectonic. The granites are only slightly deformed, as they were affected by only the very last movements of the Lineament. Initial Sr-isotopic ratios (0.707-0.712) indicate variable sources, possibly reflecting a differentiation sequence, and substantiated by the large number and great volume of associated pegmatites and aplites. Geochemical modelling of the Donkerhoek granite (Schmit in prep., in Martin, 1983a) indicates a source composed of both mantle derived and sedimentary rocks (e.g. a sequence of oceanic tholeiites and interbedded pelites in varying

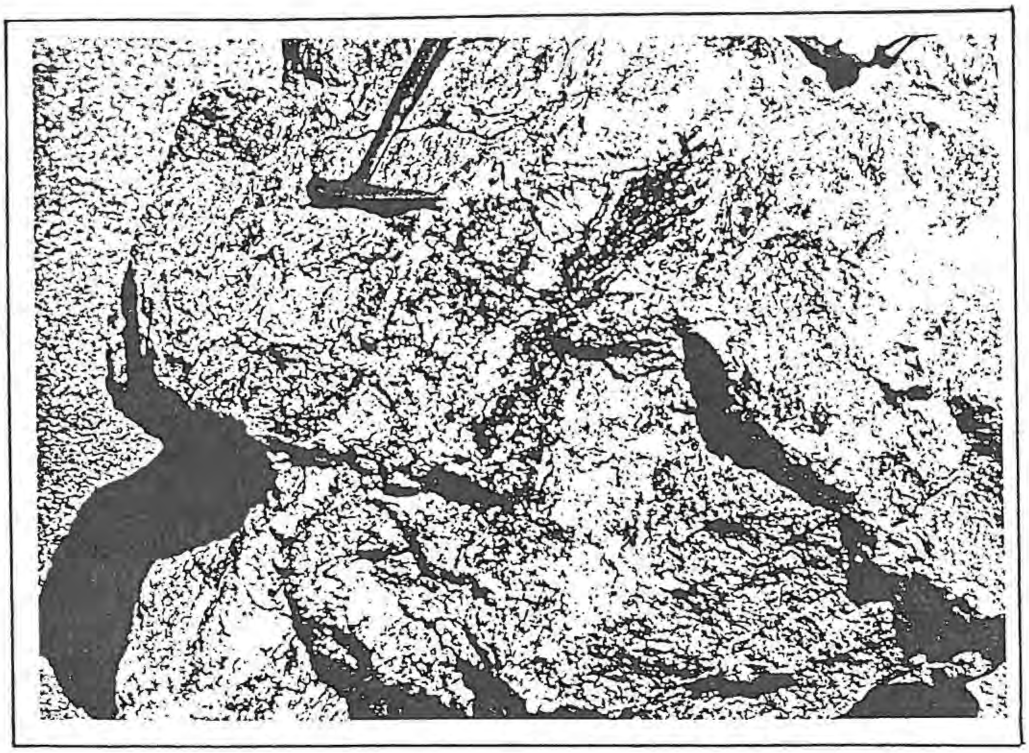


Plate 2A: Red gneissic granite, with alaskitic and reactivated basement injection

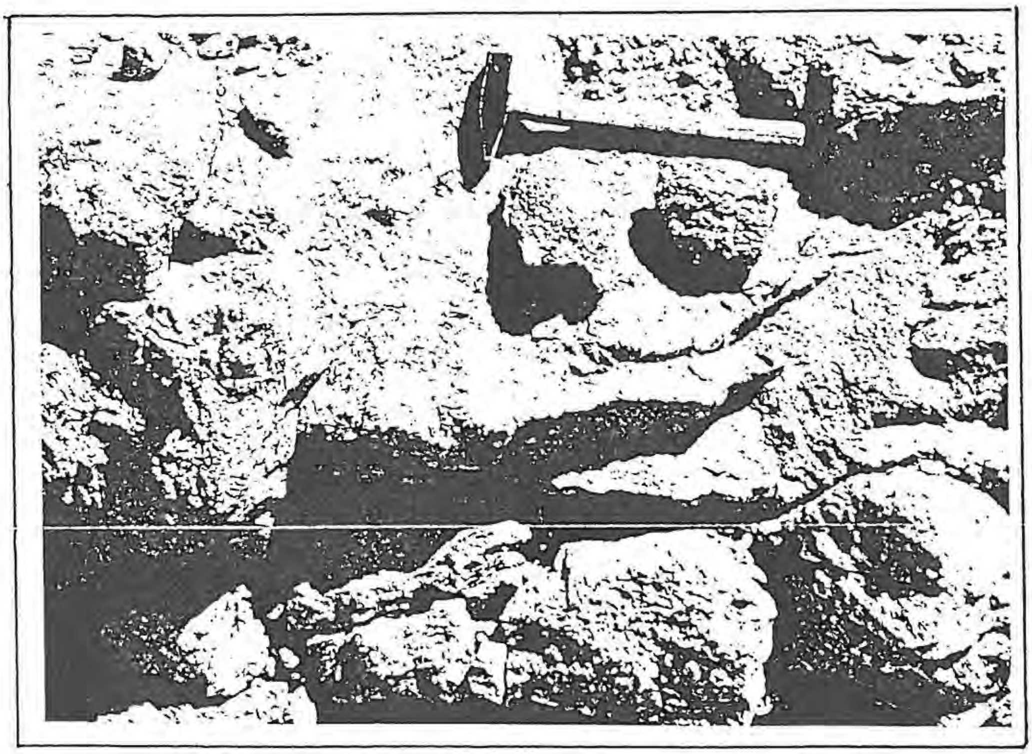


Plate 2B: Basement augen gneiss with streaks and veins of Salem-type granite, and blobs of Red granite transgressive and parallel to the foliation.

proportions). Partial melts formed at greater depths and higher temperature than the parent melts of the Salem granites. The Sorris-Sorris granite in the northern Central zone (Brandberg North area) is probably correlative to the Donkerhoek Suite. Other very late- to post-tectonic granitic stocks, sheets and anastomosing veins include the highly radioactive Bloedkoppie granite, from which the Langer-Heinrich sedimentary uranium deposit is probably derived.

Non-granitic intrusions within the Province include, in the Northern and Central zones, diorites (Palmental) and syenite, the latter associated with carbonatite in the Northern zone. These intrusions are older than or contemporaneous with the tectonic event which marked the initiation of compressional tectonism (about 550Ma), and formed the NE-fold trending structures (F_1). They are genetically different from the later granitoid intrusives, largely representing new additions to the crust whereas the granitoids are derived mainly from reworked older crustal material (Haack et al., 1983).

The next major magmatic event in Namibia involved hypabyssal intrusives and lava flows spanning the period 190-110Ma. This preceded and accompanied continental rifting in Western Gondwanaland which culminated in the separation of South America from Africa (Marsh, 1973; Marsh et al., 1978). The late Karoo magmas of the hypabyssal igneous suite overly Damara Sequence metasediments and Permo-Triassic Karoo sediments (Seidner and Mitchell, 1976). They comprise linear belts of ring complexes of highly differentiated tholeiitic and alkaline rock types which trend perpendicular to the coastline (Fig. 3.24; Marsh et al., 1978), and regional swarms of dolerite dykes. Extensive basaltic lava fields were also formed (e.g. the 900m thick Etendeka Plateau lavas, approximately 120Myr; and inland from the Cape Cross area (pre-132Myr). Of pertinence here are the late- to post-Karoo alkaline ring complexes (e.g. Brandberg, Erongo) which are spatially related to Sn and W mineralisation (see Section 4.3). At present it is not sure what the overall temporal relationship is between these intrusives and the mineralisation, but it is feasible that reworking of the Damara basement lithologies may have aided further concentration of elements such as Sn, W. The ring complexes are confined to linear structures interpreted as transform faults (Marsh, 1973) which are directly related to the rifting of South America and Africa 120-135Myr ago. The transforms are likely to have taken advantage of pre-existing weaknesses in the intracontinental branch dating from initial aulacogen propogation around 1000Ma.

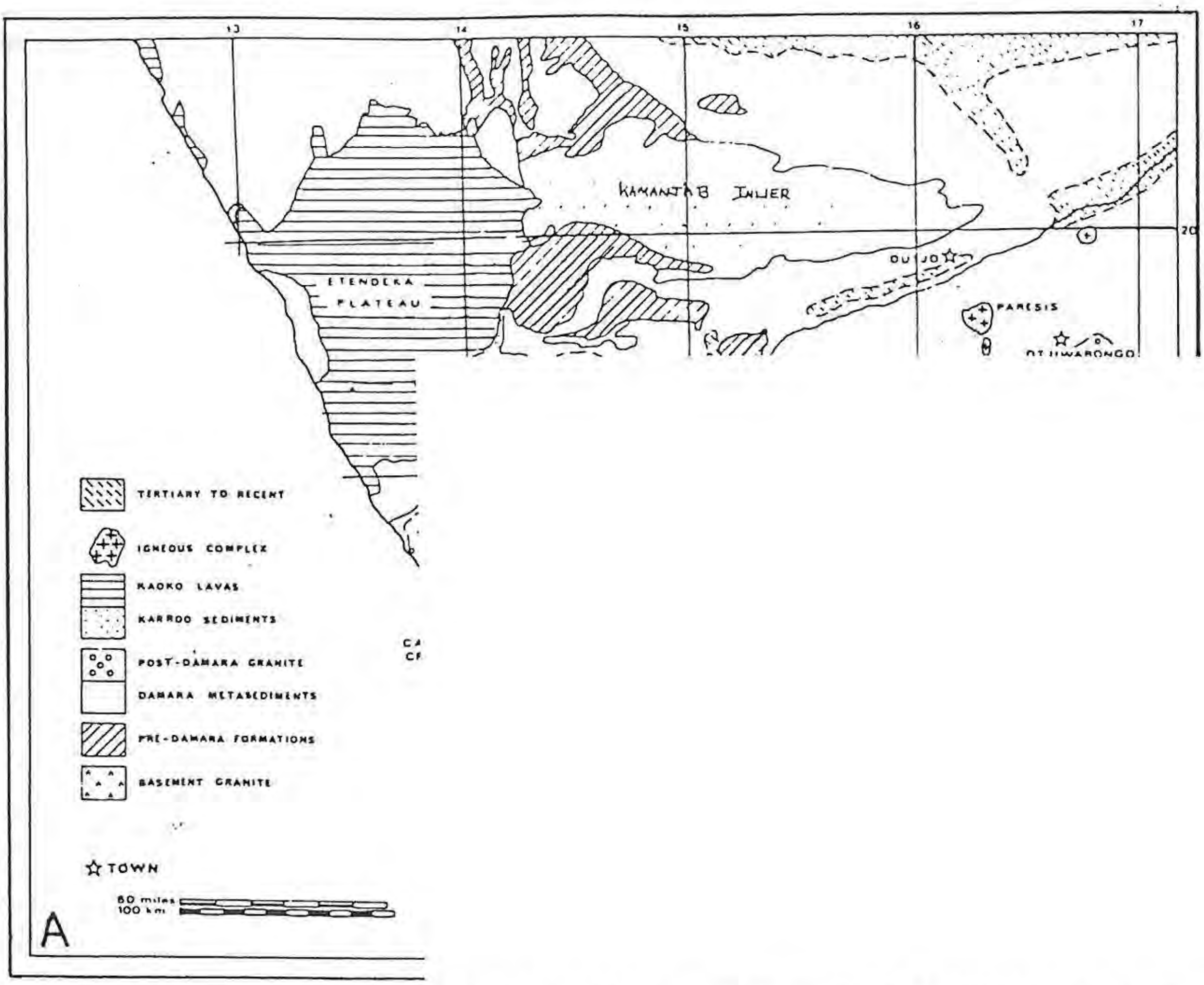


Figure 3.24A: Geological sketch map of the Damara Province showing distribution of Karoo lavas, sediments, and ring complexes (Seidner and Mitchell, 1976)

Figure 3.24B: The South Atlantic 140my ago, showing the distribution of alkaline complexes in Africa and South America. A - Angola Province; D - Damaraland Province; L - Lüderitz Province; U - Uruguay Province; B - Brazil Province, Damaraland Province - 1, Okorusu; 2, Paresis; 3, Etaneno; 4, Ondurakorume; 5, Kalkfeld; 6, Osongombo; 6, Otjihorongo; 8, Okonjeje; 9, Brandberg; 10, Doros; 11, Messum; 12, Cape Cross; 13, Klein Spitzkop; 14, Gross Spitzkop; 15, Erongo (after Marsh, 1973).

Metamorphism

It is now recognised that the first phase of development of the Damara Province involved an extended period of deposition on a rifting and stretching continental crust, lasting over 300Ma and ending before 550Ma ago, possibly at about 650Ma ago (Haack et al. 1983; Martin, 1983a, Fig. 3.20). The second phase of development is marked by a change from extensional to compressive tectonics. A relatively rapid evolution occurred between 550 and 510Ma, involving deformation, crustal thickening, uplift, metamorphic heating and voluminous granitic intrusions, terminating around 450Ma ago.

The grade of metamorphism increases from the margins of the Province towards the Central zone i.e. from lower greenschist to granulite facies (Fig. 3.25). Mineral growth phases are attributed to a gradual progressive increase in grade, (Höffer, 1977, in Miller, 1981) or to two separate peaks of metamorphism separated by a drop in temperature (Barnes and Sawyer, 1980; Kasch, 1979). Arguments for polymetamorphism are based on observations which show that structural fabrics, formed during the growth of certain prophyroblasts, were deformed (plicated or rotated) prior to the growth of new metamorphic mineral assemblages requiring changed P-T conditions (Martin, 1983a). However conjectural evidence exists, such as low grade assemblages (pre- or syntectonic cordierite-garnet-sillimanite porphyroblasts) in close proximinity to higher grade (post-tectonic cordierite K-feldspar-garnet porphyroblasts) in the southwest of the Central zone (Barnes and Sawyer, 1980). The change from low to higher grade may be accounted for by one prolonged episode of prograde metamorphism during which the rate and style of deformation changed (Martin, 1983a).

P-T conditions in the Central zone were low pressure (3-4Kb), high temperature (590-660°C; Puhan, 1983). In the Southern zone P-T conditions were around 6-10kb and 560°C. The actual thermal peak (560°C) would not coincide with the pressure peak, as during orogenesis metamorphic conditions tend to move through a P-T loop (e.g. see Limpopo Belt, Barton, 1983). The temperature peak was reached after the cessation of the last major petrofabric forming deformation (Puhan, 1983). The Central and Southern zones, with respect only to their P-T conditions, thus resemble a paired metamorphic belt. The highest temperatures (max. about 600°C, Fig. 3.25) were in the western part of the Central zone where partial melting of metasedimentary rocks was locally fairly extensive.

The peak of regional metamorphism was attained in the Okahandja Lineament Zone and in the adjoining parts of the Central and Southern zones. The peak occurred prior to the intrusion of the Donkerhoek plutons 520Ma ago, and after the termination of the main deformation events (F_1, F_2, F_3 - although different workers, recognise more in different areas). In the northern part of the Central zone the peak was reached at about the same time (514 ± 22Ma ago), but during or just after F_2 and prior to F_3, inferring diachronous structural events (Martin, 1983a). In general terms, right across the Province, the metamorphic peak was reached at around 530Ma ago (Fig. 3.21; Haack and Martin, 1983).

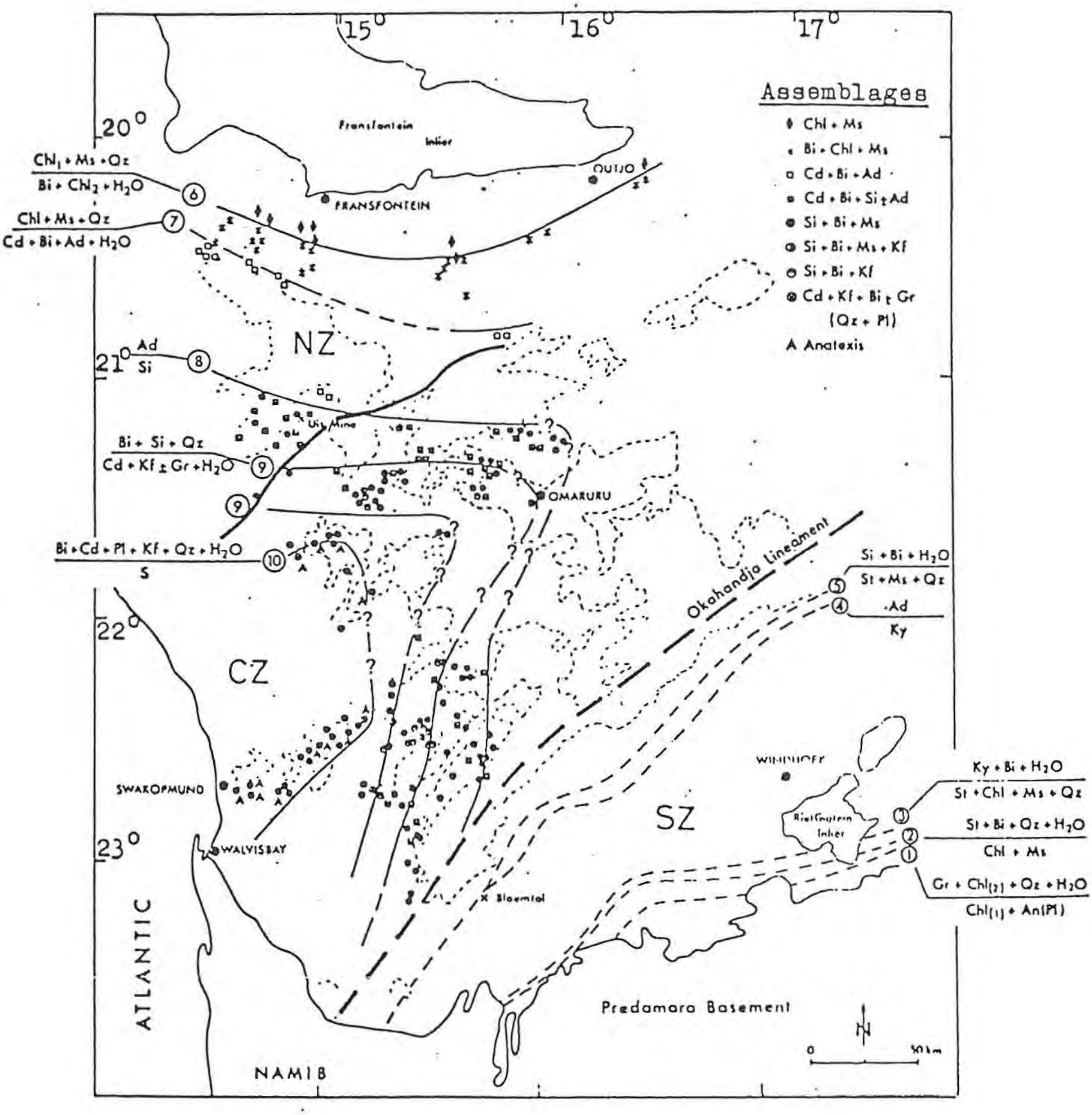


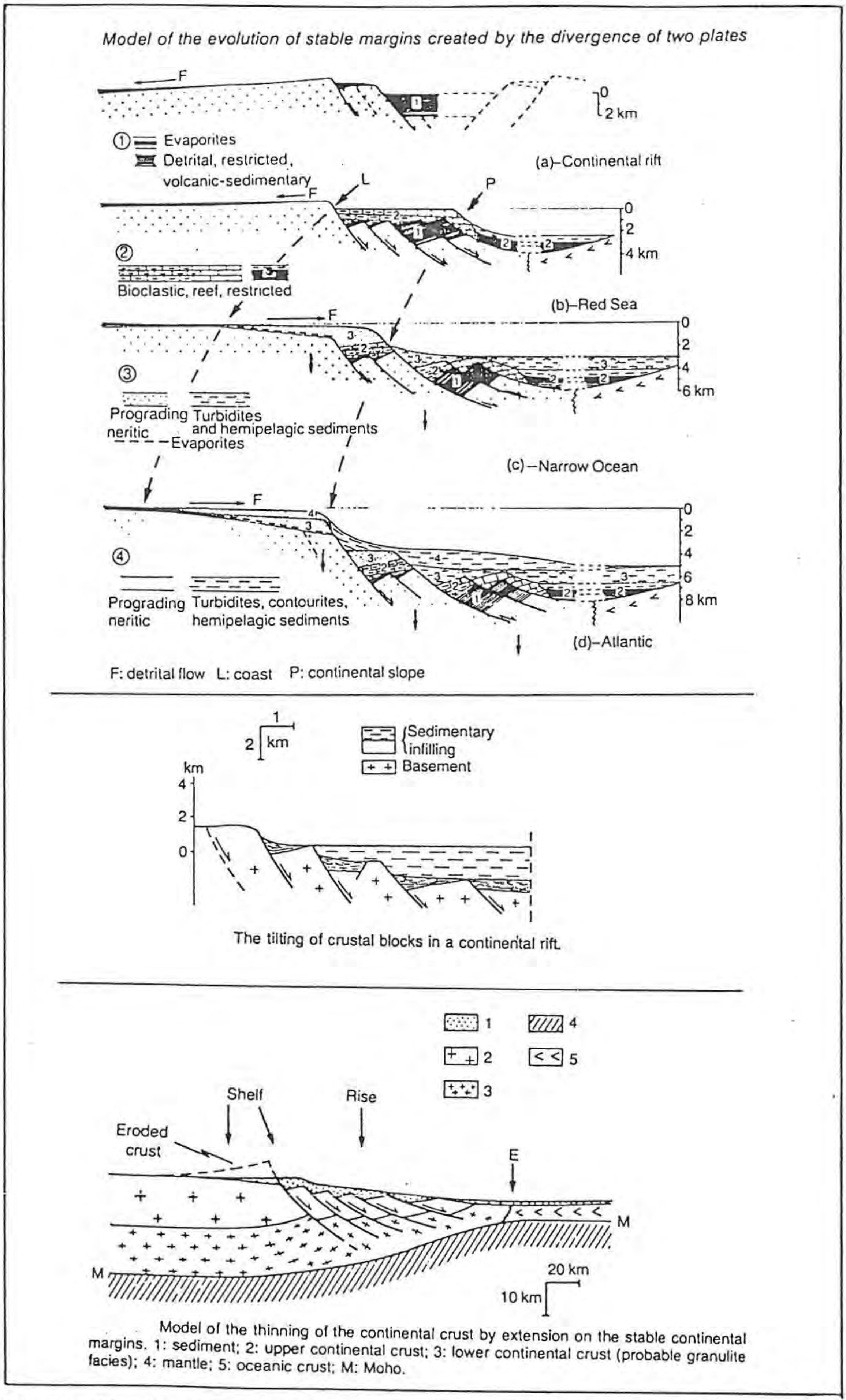
Figure 3.25: Isoreactiongrads in pelitic rocks in the Damara Province (Höffer, 1977, in Miller, 1981)

3.3.3 Geodynamic models

Saldania Province

Considerable research in the Cape Fold Belt has clarified certain aspects of the geodynamic evolution of the Belt, however no synthesis has yet been provided for the Saldanian Province. This is due to the poor preservation of the Province, coupled with a lack of exposure, and also the severe tectonic overprinting by the later Cape orogeny.

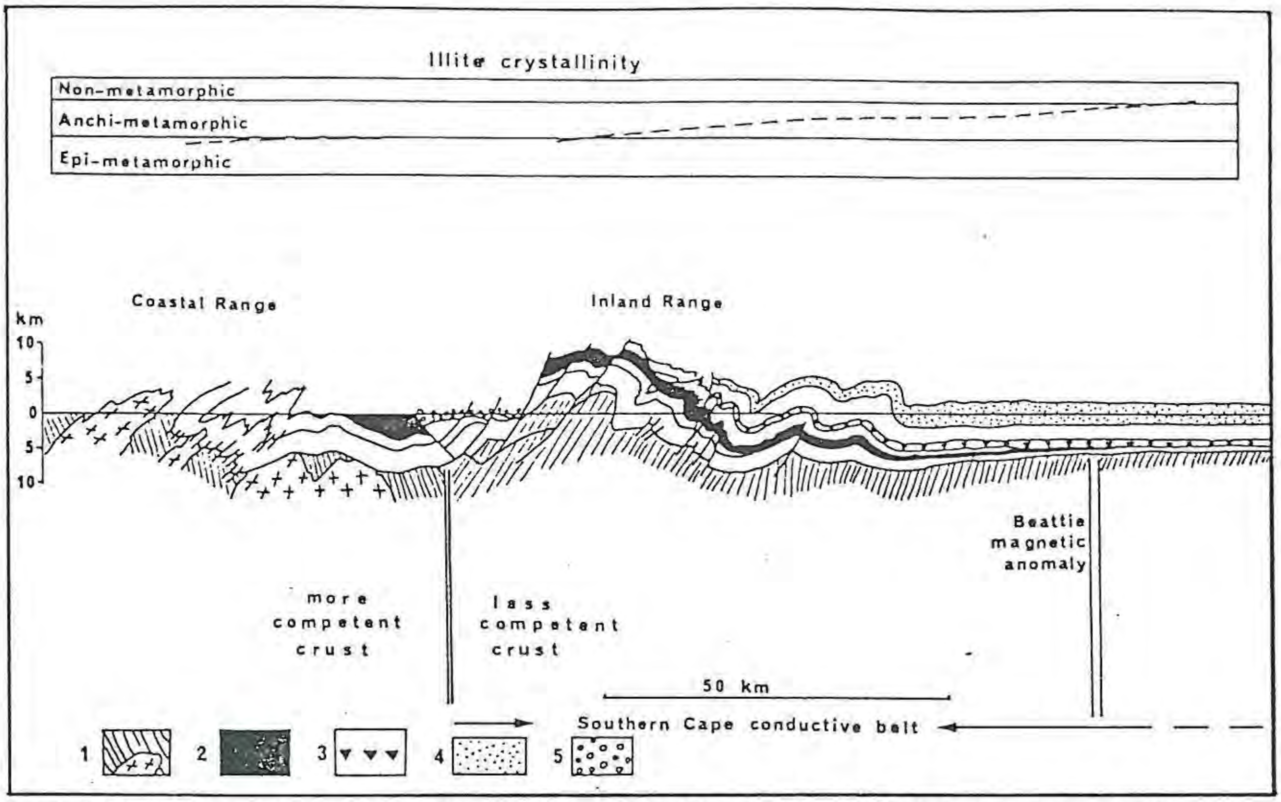
It appears that the Province formed as either an intracratonic basin (aulacogen) or continental margin peripheral to the Kaapvaal/Namaqualand craton, essentially in response to Pan African extensionism related to initial attempted splitting of Gondwanaland. A certain amount of divergence (rifting) is envisaged, which importantly occurred at a slow rate, enabling fairly lengthy periods of turbidite sedimentation, as well as reworking of more proximal neritic sediments during later stages of the orogeny. The minor volcaniclastic component of the Malmesbury Group was probably introduced along rift-fault margins. The lack of any substantial basaltic pile or ophiolitic assemblage points to probable opening of the rift to a narrow ocean stage (Fig. 3.26). It is suggested that the fragmentation of Gondwanaland largely removed axial portions of the rift sequence (turbidite/flysch sediments ±spilites?), and that what little is preserved now of the Saldania Province constitutes the rift shoulder environs. Tectonism during the Pan-African included a compressional phase (with low grade metamorphism) and syn- to late- and post-tectonic granitism. A cross-section of the Cape Fold Belt (Fig. 3.27) indicates the present exposure of the granites as upfaulted or upfolded inliers within the Cape Fold Belt. It is feasible that most of the granites were emplaced in rift margin areas adjacent to where maximum extension occurs (Fig. 3.26B), utilising the faults shears and planes of crustal weakness in this zone to facilitate diapiric uprise. Closure and minor collision within the Saldania Province around 700-600Ma may have also generated sufficient heat to cause generation of melts along the shear zones, (see section 2.0), which appear to have been reactivated numerous times, as well as creating suitable avenues for granitic intrusion. The Cape Fold Belt orogeny further modified granite emplacement, thrusting the granites N-NE. This may be analagous to granitoid emplacement along shears in the supposedly collision-related Cornubian batholith (Badham, 1982; Hutchison, 1983). However the actual evidence for substantial collision (e.g. ophiolite-suture zones, high



The tilting of crustal blocks in a continental rift.

Model of the thinning of the continental crust by extension on the stable continental margins. 1: sediment; 2: upper continental crust; 3: lower continental crust (probable granulite facies); 4: mantle; 5: oceanic crust; M: Moho.

Figure 3.26: Aspects of continental rifting and the evolution of a stable continental margin, involving thinning of the continental crust, which may be applicable to the Saldania Province (Boillot, 1981)



Stratigraphic and structural profile through the Cape Fold Belt between 22 and 23°E. 1 = pre-Cape sequences with granitoids, 2 = Bokkeveld Group, 3 = Dwyka Group, 4 = Beaufort Group, 5 = Jurassic-Cretaceous

Figure 3.27: Stratigraphic and structural profile through the Cape Fold Belt (Dingle et al., 1983)

pressure metamorphic assemblages) is not discernable. One can only speculate that the closure of the Saldanian Province occurred when the rift-trough was largely filled with a substantial sedimentary pile, subsidence of which may have led to granitoid generation at depth, although there is no unequivocal metamorphic evidence of sedimentary burial (Hartnady et al., 1974). The sediments and granites are likely to have been deformed and thrusted back onto the continental foreland along the same fracture zones along which initial subsidence occurred, which may explain the overall lack of or low grade metamorphism within Malmesbury Group lithologies.

Damara Province

The following discussion is given in some detail, as the Damara Province is one of the better studied Proterozoic and perhaps Palaeozoic orogenic belts that do not appear to fit the Wilson Cycle model. Resolution of the broad dichotomy between concepts of Wilson Cycle orogeny and those of ensialic orogeny, and the fundamental differences between the two are central to a fuller understanding of ancient orogenic belts, as well as the implications of the various metal deposits present or expected present in such belts.

The Pan African Damara orogen has, during the last decade, been the subject of multidisciplinary studies aimed at elucidating the geodynamic development of the well-exposed intracontinental branch of the fold belt. Most of the debate on the geodynamic evolution of the Damara Province revolves around the actual amount of extension or spreading, coupled with models to account for the deformation, crustal thickening, uplift, metamorphism and granitoid magmatism that are related to the compressional tectonic regime. All the proposed geodynamic models assume subduction processes of some kind, including continental subduction (Ampferer subduction) and ocean floor subduction (Benioff subduction). The former is represented by an "aulacogen" model (Martin and Porada, 1977) and a "delamination" model (Kröner, 1977, 1981, inpress). Models within the second category include two models assuming subduction of a wide ocean (Barnes and Sawyer, 1980; Kasch, 1979, 1980), one model proposing subduction of a narrow ocean arm (Miller, in press) and one advocating processes of strike-slip faulting coupled with oblique subduction of small oceanic pullapart basins (Downing and Coward, 1981; Coward, in press). Fig. 3.28 illustrates these concepts, although the reader is referred in particular to Martin (1983b) for a critical discussion, as well as Tankard et al. (1982) for an overview.

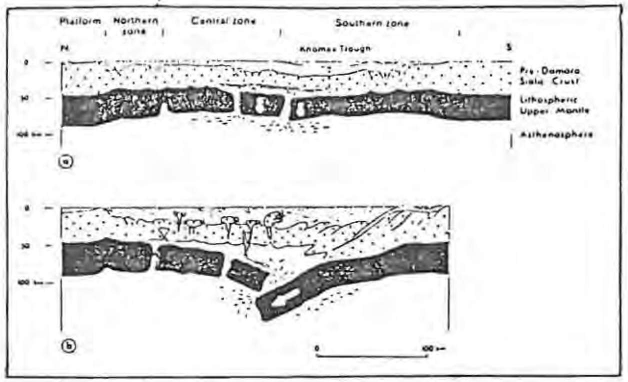
Only salient points of proposed models are given here, followed by an overview of some of the constraints recent investigations have elucidated.

Aulacogen model (Fig. 3.28A).

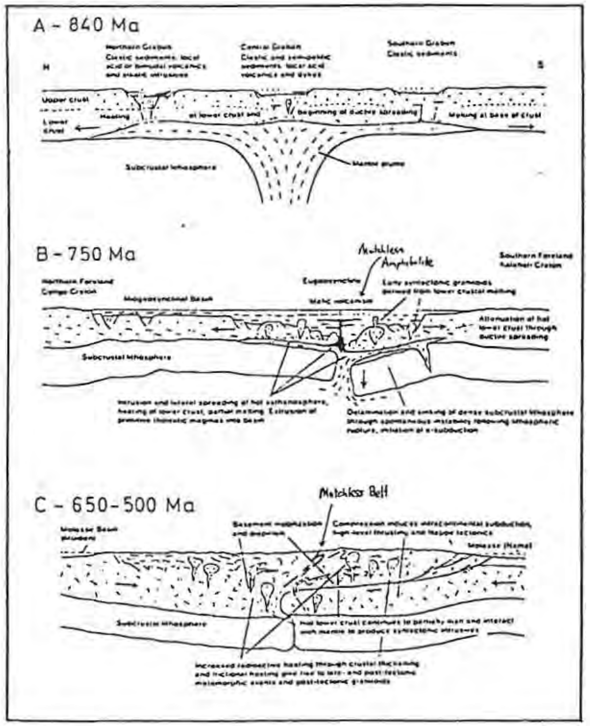
The geosynclinal pile was deposited in a broad rift basin (aulocogen) with considerable stretching and thinning of the crust. Subsequent gravitational instability of the upper mantle causes detachment and sinking of the dense lithospheric layers, enabling ascent of an asthenolith diapir. Further sinking of the dense slabs causes folding of the weakened crust. High heat flow from the growing asthenolithic diapir causes high-grade metamorphism, anatexis and granitic plutonism within the crust.

Recent work has elucidated some points and interpretations that apparently weaken the "aulacogen" model (Martin, 1983b):

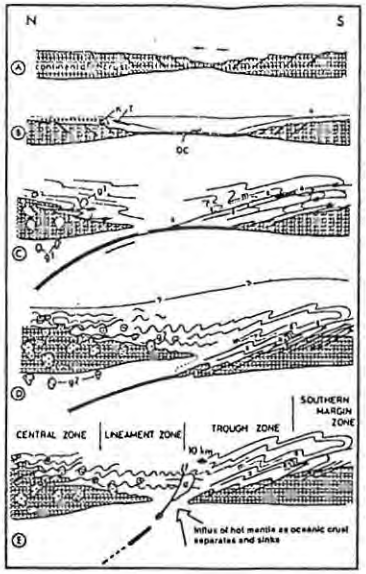
i) The supposed glaciogenic mixtites (Chuos Fm) have been reinterpreted as mass-flows and various types of gravity flows. This has eliminated any argument for a chronostratigraphic correlation of similar lithofacies units of the lower Damara



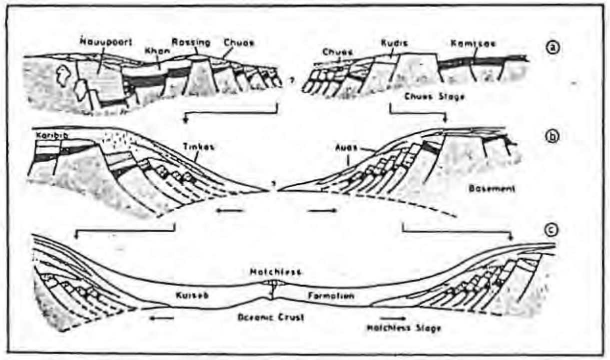
A: Schematic interpretation of the orogenic stage of the Damara fold belt (after Martin & Porada, 1977).
a: After deposition of the geosynclinal pile in a broad rift basin ("aulacogen") gravitational instability of the upper mantle causes detachment and sinking of dense lithospheric layer with concomitant ascent of an asthenolith diapir.
b: Further sinking of dense slabs causes folding of the weakened crust; heat of the growing asthenolithic diapir causes high-grade metamorphism, anatexis and granitic plutonism within the crust.



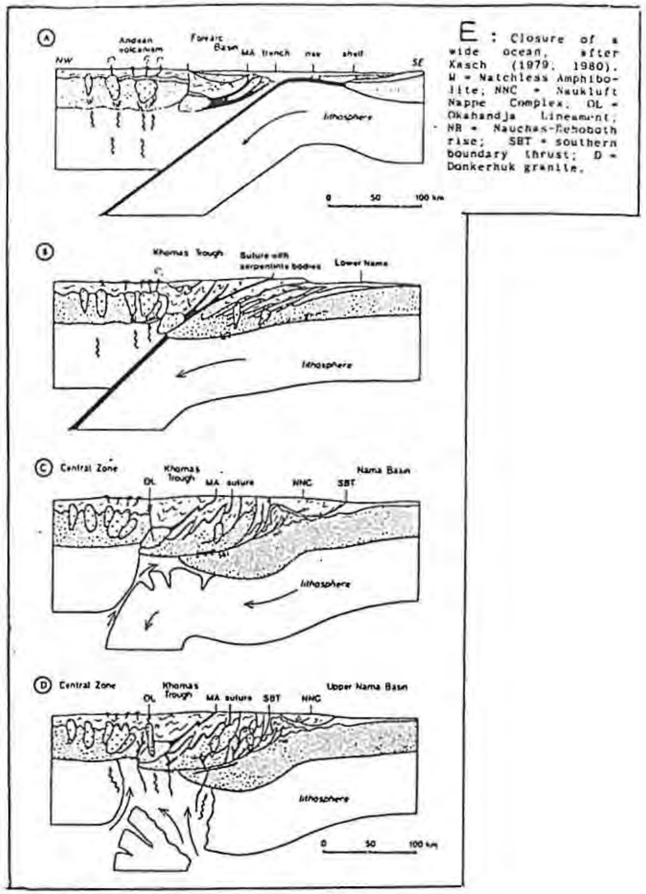
B: "Delamination model", after Kroner (in press) showing three stages in the suggested evolution of the Damara Belt.



C: Closure of a wide ocean, after Barnes & Sawyer (1980). OC = oceanic crust; S = serpentinite; M = Matchless Amphibolite Member; a = Auas Formation; K = Karibib Formation; T = Tinkas Member; g1,g2 = early and late kinematic Salem Granite emplacement; d = Donkerhuk Granite.



D: Opening of a narrow ocean, after Miller (in press). Progressive deepening of the depositional environment in the Southern Zone; concomitant burial of the spreading ridge by the Kuiseb Formation. a): Chuos mixtite, continental slope (?); b): Auas Formation, Tinkas turbidites, deep water fans; c): Kuiseb Formation, oceanic stage with deep-water flysch and mid-ocean ridge volcanics.



E: Closure of a wide ocean, after Kasch (1979, 1980). M = Matchless Amphibolite; NNC = Naukluft Nappe Complex; OL = Okahandja Lineament; NR = Nauchas-Rehoboth rise; SBT = southern boundary thrust; D = Donkerhuk granite.

Figure 3.28: Geodynamic models for the evolution of the Damara Province
A. Aulacogen model, Martin and Porada (1977)
B. Delamination model, Kroner (in press)
C. Closure of a wide ocean, Barnes and Sawyer (1980)
D. Opening of a narrow ocean, Miller (in press)
E. Closure of a wide ocean, Kasch (1979, 1980)
(Figures from Martin, 1983b)

Sequence exposed in the Northern and Central zones with those of the Southern Marginal zone. It is possible that the sediments of the Nosib Group and the Rössing, Chuos and Karibib Formations may be older than the Kamtsas/Duruchaus and Kudis successions (see Hartnady, 1979).

ii) The Matchless Amphibolite has geochemical characteristics of MORB, with the basic volcanic rocks depleted in light rare earth elements and having high $^{143}Nd/^{144}Nd$ ratios, indicating derivation from source regions that had been depleted in LREE's for a considerable period of time (Hawkesworth et al., 1981), and that may be analagous to magmatism at active spreading centres. Problematic to an interpretation of the Belt as a suture zone is the remarkable similarity of the chemistry and isotope composition of the Kuiseb schists on either side of the Matchless Belt, as well as the lack of evidence for ophiolitic rocks, volcano-sedimentary mélanges, sheeted dyke systems, and the extensive length and very narrow nature of the Belt.

iii) The number and size of basement thrust and nappes is more easily explained with ocean floor subduction c.f. 'continental' subduction. On the other hand the ocean floor models have problems reconciling the lack of island arc volcanism, and the lack of andesitic volcanism or its plutonic equivalents. The granitic rocks of the Central and Northern zones appear to have been derived by partial melting from older sialic basement or Damara metasediments.

Delamination model (Fig. 3.28B)

Kröner (in press) has proposed the above model, assuming that the ensialic geosynclinal basin which had been formed by crustal stretching over a mantle plume, was closed by a process of delamination, followed by continental subduction, crustal underthrusting, and interstacking. The general concept is similar to that of the aulacogen model (above). Kröner's model has numerous deficiencies (Martin, 1983b), including

i) in stage A, the formation of grabens relative to a single mantle plume below the central graben. This assumption conflicts with the fact that by far the most voluminous and varied igneous activity occurred in the northern graben;

ii) in stage B, plutonic intrusions into the southern part of the Khomas Trough are shown for which there is no evidence at present. Also the great thickness of the Kuiseb formation in the Northern and Central zone has not been incorporated;

iii) in Stage C, there is no indication of the large granitic plutons which have intruded the Northen zone, or a reason for the formation of these melts. Also no indication is given of the differential vertical movements related to the Okahandja Lineament (Miller, 1979), or the existence of the numerous sepentinite bodies tectonically emplaced into the Southern Zone, which Barnes and Sawyer (1980) interpret as evidence for the former existence of oceanic crust.

Subduction of a wide ocean (Fig. 3.28C)

Subduction-type models were first proposed by Hartnady (1974) and Watters (1974), and have received further support over recent years (Hartnady, 1978, 1979; Blaine, 1977, Sawyer, 1978; Barnes, 1979; Kasch, 1979; Barnes and Sawyer, 1980). The general conception is continental convergence facing northwest (Watters, 1974, originally proposed convergence to the SE, producing the 1300-900Ma Rehoboth Magmatic Arc), with the Kalahari plate being subduction beneath the leading Congo plate. The Khomas trough is interpreted as a fore-arc trench, and the Central zone granites as equivalent to a magmatic arc. The Matchless amphibolite was initially proposed as a suture zone, and comprising an obducted sliver of oceanic crust trapped between the converging plates.

(Note: Rehoboth magmatic ages appear to be more between 1200-1700Ma, Fig. 2.21, and the Rb-Sr 1300-900Ma ages may be a metamorphic overprint. They certainly are hard to reconcile with Damara ages).

Features in favour of the subduction models, following closure of a wide-ocean, include:

i) the mid-ocean ridge geochemistry of the Matchless meta-tholeiites;
ii) the existence of major thrust sheets and nappes in the southern part of the Southern Zone;
iii) the tectonic emplacement of numerous serpentinite bodies in the same area;

iv) the P-T conditions revealed by metamorphic grades, which indicate high pressure in the southern part of the Southern Zone, whereas the Central Zone is characterized by lower pressure and higher temperature conditions. As much as a paired metamorphic belt (e.g. Miyashiro, 1977) may be invoked, there is no evidence in the Damara Province of voluminous ophiolites, a conspicuous feature of such belts (Martin, 1983b).

Problematic at present within the proposed models, are

i) The occurrence of the Kuiseb Formation in similar facies and great thicknesses in the three main structural zones, which conflicts with the interpretation of the Khomas Trough as a forearc basin in which the Kuiseb Formation was deposited. Interpretation of the Northern zone as a backarc basin may aid this concept, however the continuity of the Khan, Karibib and Kuiseb Formations right across the Central zone detracts from the presence of an arc in the Central zone.

ii) The plutonic rocks of the Central zone have been interpreted as a calc-alkaline arc. Geochemical characteristics indicate derivation from sialic crustal sources, and are not indicative of Andean-type magmas. This is further substantiated by the lack of tonalites, mafic plutonic rocks (very rare and of small volume), and that granites by far dominate over granodiorites (Haack et al., 1983).

iii) Dating of granitoids indicates a spread of about 100Ma (even up to 200Ma) of emplacement. Subduction generated granite emplacement over that period of time, at present day subduction rates, require subduction of several thousand kilometres of oceanic crust. Available paleomagnetic data for the Kalahari Congo Provinces (McWilliams and Kröner, 1981) does not substantiate this.

Subduction of a narrow ocean arm (Fig. 3.28D)

The main difference from the other proposed subduction models, in this case (Miller, in press) is the assumption that in the area of the later Khomas Trough initial rifting and crustal thinning led to the formation of an ocean of limited width (approximately 100km), narrow enough to allow sedimentation to keep pace with spreading and to bury the spreading ridge (e.g. analogous to the Red Sea). The Matchless Member is interpreted as a late extrusion of ridge tholeiite through Kuiseb sediments which had already covered the ridge. Conjectural points include the comparison of the Central zone to an Andean-type plateau, erosion from which partly provided the Kuiseb beds of the Khomas Trough. The inference that these are younger than lithologically similar, and across the Okahandja Lineament, contiguous sediments in the Central and Northern zones, is not yet fully substantiated (Martin, 1983b). The great thickness of the Kuiseb Formation in the Northern zone remains unexplained.

Small ocean basins and strike-slip movements

Downing and Coward (1981) and Coward (in press) have proposed a model accounting for the Damara orogen in terms of three major movement phases. Earliest movement (M_1) was confined to the northern part of the coastal branch, producing the SE verging Sesfontein thrust and nappe zone. The second phase (M_2) caused only weak SW verging structures in this zone, but correlates with the main phase of deformation in the Central zone of the intracontinental branch. There the deformation is attributed to differential shearing along a sinistral, NW dipping, low-angle transform fault. Large-scale SW vergent sheath folds involving Damaran metasediments and pre-Damara basement resulted. Pull apart basins, as opposed to a wide ocean, partly floored by oceanic crust, may have formed along this fault, acting as traps for thick sediment accumulation. Damaran granites were generated partly from oblique subduction of oceanic material within the basins, and partly from shear-heating at local crustal delaminations and/or radioactive self-heating.

A third phase (M_3) was initiated by a change of the stress field, causing the main Damara plate (Congo plate) to over-ride its southern foreland in a SE direction, producing several fold and thrust phases in both the Damara and Kalahari Provinces. Coward's model has yet to be fully substantiated by field observations, in particular the interpretation of major structures of the Central Zone as large-scale sheath folds produced by pervasive shearing.

Discussion

None of the above models fully explains all the features presently known about the Damara Province. The models circle around one essential paradox: the area which exhibits the most spectacular uplift structures, the Central zone, also contains the rocks with the lowest pressure assemblages (Hawkesworth et al., in press). Possibly the uplift of the Central zone occurred while the rocks were still hot, so that prograde metamorphism and partial melting at high structural levels reduced the pressure. This hypothesis requires a 50-70km thick crust before 560Ma, and implies that the 550-560Ma regional metamorphism was at least partly due to crustal thickening (Hawkesworth et al., in press). Subsequent uplift caused the Naukluft gravity nappes to move south (the margins cooling at around 500Ma), while continued uplift of the deeper hotter material in the Central zone resulted in dome-like structures and prograde metamorphism until 460Ma.

Hawkesworth et al. (in press) have summarised some of the key topics which when unravelled will aid geodynamic interpretations of the Damara Province. These include the striking occurrence of dome-like structures formed during the late uplift of the belt. There is a problem in assessing what has been eroded and the extent to which the sediments that are preserved reflect the large scale processes which led up to, and were responsible for the Damara thermal event. Both Nd- and Sr-isotope analyses of Nosib metasediments and U/Pb results on detrital zircons within them suggest probable derivation from the underlying basement. The fact that 3-4km of Nosib metasediments are still present in the Central Zone and also that they remain in their likely source terrane is possible evidence contrary to large scale overthrusting related to plate collision.

The Matchless Amphibolite belt is cited as evidence for the existence of a small ocean and the widespread granites in the Central zone are attributed to Andean-type magmatism above a northerly dipping subduction zone (Kasch, 1979; Barnes and Sawyer, 1980; Miller, in press). Intraplate geochemistry, including depleted LREE's infer a mid-ocean ridge analogy. However, models have to accommodate the related Besshi-type Cu-Zn mineralisation, the lack of evidence for obducted ophiolite, and the similarity in chemistry and isotope composition of the enclosing Kuiseb schists on either side of the Matchless belt.

The few analyses of Damara granites conflict with the suggestion that they were generated above a subduction zone. Recent calc-alkaline rocks rarely have over 80ppm Ce, whereas both Salem-type and leucocratic granites from the Damara have 100-300ppm Ce (Hawkesworth et al., 1981) as well as far higher LREE abundances. Similarly low Nb contents (less than

15-20ppm) are usually a distinctive characteristic of subduction zone related magmas, however the Salem Suite has an average of 40ppm Nb (Miller, 1973) as well as initial $^{87}Sr/^{86}Sr$ ratios in the range 0.7048-0.7054. High LREE and Nb abundances are typical of more recent, incompatible enriched, usually alkaline, magmatic provinces in intraplate environments. The initial Sr-isotope ratios are significantly higher than Andean-type magmas, collaborating an intracratonic affinity and a crustal source for the granites (i.e. S-type ilmenite series).

Other possible objections to Andean, plate collision, or Himalayan or even marginal basin models include the lack of an unambiguous ophiolite suite and blueschist metamorphic assemblages although these may have been lost to erosion during late uplift. Paleomagnetic results tentatively suggest little relative movement between the Congo and Kalahari cratons since pre-Damara time (McWilliams and Kröner, 1981). The intra-continental branch of the Damara is far wider (450km) than many recent orogenic belts formed along plate margins and/or in collision zones. There is also no indication of diachronous magmatic provinces characteristic of active destructive plate margins (Hawkesworth et al., in press).

Geochronological investigations have shown that earliest events include alkaline magmatism at 800-750Ma, a cluster of granite rocks around 750Ma (Kröner, 1981), and that by 650-620Ma there was further granitic activity and an associated M_1 metamorphism (Kröner, 1981; Downing and Coward, 1981; Hawkesworth and Marlow, in press). Molasse-type sediments were deposited before 550-570Ma, indicating that some uplift had already taken place in the centre of the belt.

The next major event was widespread granitic activity and regional metamorphism at 560-550Ma. This episode post dates the major regional F_2 deformaton. Most of the granites are Salem-type, outcropping right across the Damara granite belt. Metamorphism is reflected in the Rb/Sr whole rock age of 548 $\pm$ 56Ma on the Kuiseb schists.

Problems arise correlating and assessing the regional significance of particular deformation events. Hawkesworth et al. (in press) in summary suggest that the most significant regional tectonic movements had finished by 560-550Ma (F_2), and that the subsequent thermal, magmatic, and even deformation events in the period 550-450Ma may be viewed as consequences of those early movements.

The period 550-450Ma is significant in respect of:-

- the last magmatic event in the central area is the emplacement of late alaskitic granites. Model Nd ages indicate derivation from crustal material at least 2.0Ga old, and some can be seen to have been generated by in situ melting of pre-Damara basement gneisses and Nosib metasediments. Partial melting at higher structural levels is indicated, especially as earlier granites are clearly intrusive at the present erosion level, substantiated by higher $^{87}Sr/^{86}Sr$ ratios for the younger granites (Hawkesworth and Marlow, in press).

- The generation and intrusion of alaskites (510-460Ma) is intimately associated with uplift and the formation of dome-like structures in the Central zone (Barnes and Downing, 1979). The Okahandja Lineament is important as early relative movements between the Central and Southern zones took place along it. The last movements are post-dated by the intrusion of the Donkerhoek Granite (520Ma). This period appears to mark a transition from uplift along discrete linements to regional updoming, and importantly may reflect greater ductility in response to increased geothermal gradients (Hawkesworth et al., in press).

- Transport from the north of the Naukluft Nappes is consistent with uplift in the central Damara between 530 and 485Ma.

- Metamorphic analyses generally reflect P/T regimes between 560-450Ma. Northern margin rocks are of greenschist facies. Regional metamorphism in the south attained around 620°C @ 6-8kb (Sawyer, 1980), and locally has been overprinted by a thermal effect by intrusion of the Donkerhoek Granite, yielding pressure estimates of approximately 4kb. In contrast the Central zone high grade metamorphism (650°C @ 3-4kb) has associated partial melting to form alaskite (emplaced 510-460Ma). The presence of such partial melts indicates that prograde metamorphism was taking place, and the absence of thermal aureoles indicates that this was the last thermal event (Hawkesworth et al., in press).

Hawkesworth et al., (in press) have advanced a qualitative model (Fig. 3.29) for the late stage evolution of the Damara Belt. It incorporates features discussed above.

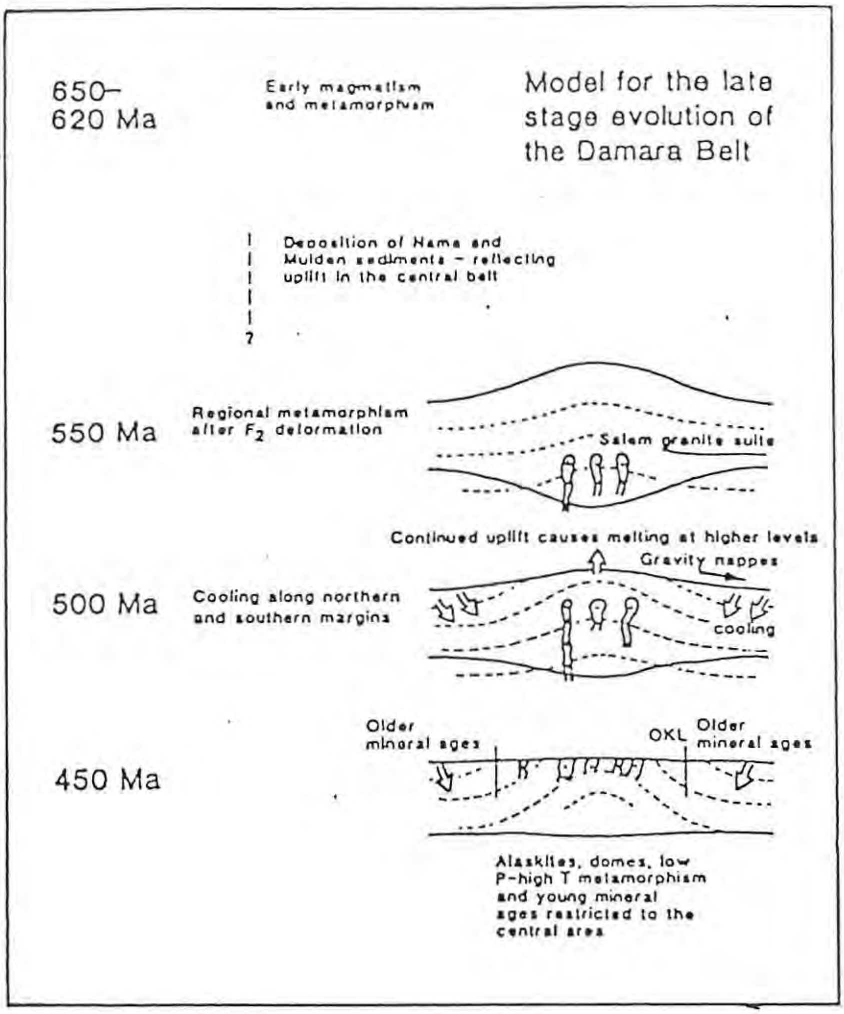


Figure 3.29: Model for the late stage evolution of the Damara Belt (Hawkesworth et al., in press).

Although isotopic studies have demonstrated the Damara event to have resulted in remobilisation of considerable quantities of old pre-existing crustal material, the picture is complicated by the enriched intra-plate characteristics of many of the igneous rocks (Hawkesworth et al., op. cit.), making difficult any assessment of how much new continental crust was generated. Furthermore, much of the available evidence appears to support two potentially conflicting schools of thought (ensialic versus ocean floor). The close relationship between the Nosib metasediments and underlying basement, the intraplate component in those granites analysed, and lack of detectable relative movements between the Kalahari and Congo emphasise the intracratonic (or intraplate) nature of the intracontinental branch. On the other hand variations in both metamorphic conditions and Rb-Sr and Ar cooling ages (Hawkesworth et al., in press), the development of spectacular dome-like structures in the Central Zone, plus partial melting at higher stratigraphic levels point to a prolonged period of uplift, presumably as a response to crustal thickening, and present day analogies of the latter occur along destructive plate margins or in Himalayan-type collision zones. It would appear that a closer truth may lie somewhere between these concepts, probably involving initial development of an aulacogen or a limited ocean arm, with limited flat-type subduction processes involved in the compressional phases of tectonism. The lack of sedimentary-volcanic associations typical of island arc environments, and the considerable similarity to the geosynclinal development of the North Sea Basin (see Porada, 1983), substantiate this concept

3.4 Anorogenic settings

Anorogenic settings, as the term implies, are distinct from those discussed above as there is no diastrophic tectonic activity as such (i.e. subduction, collision etc.). Only brief mention is made here to balance the picture, as significant tin deposits are associated with these settings, although the dissertation topic is constrained principally to some of those deposits related to orogenic systems. Examples cited of Sn/W mineralisation related to anorogenic settings include the Nigerian Sn-province, the Rondania province in western Brazil, and can also be extended to include large layered complexes and associated intrusives such as the Bushveld Complex, South Africa.

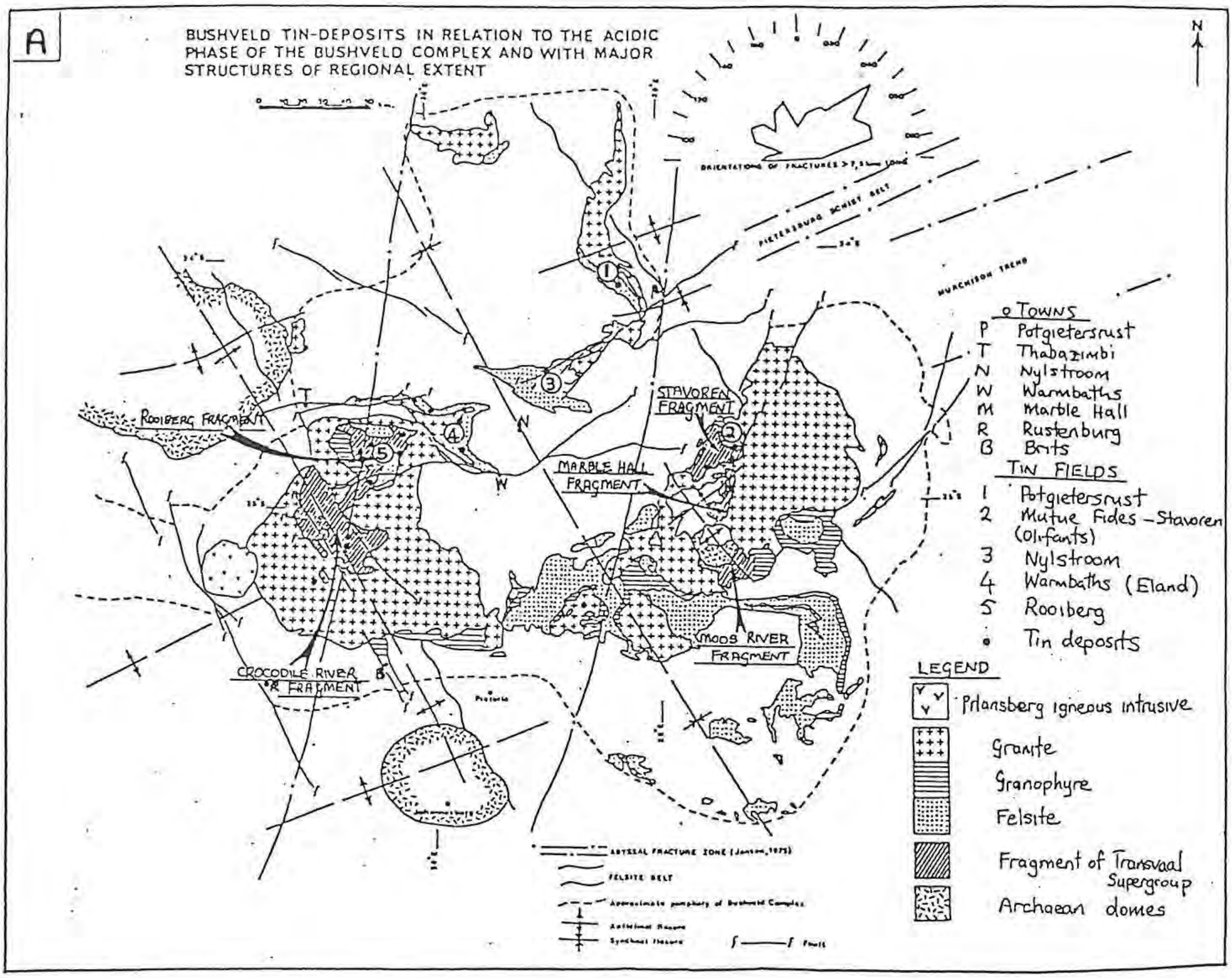
Magmatic activity in anorogenic settings is ascribed to thermal processes in the asthenosphere (mantle plumes) giving rise to hotspots, whereas intra- and inter-continental rifts and aulacogens are related to incipient plate boudaries. Mantle plumes are these days considered a fundamental aspect of lithosphere-asthenosphere interactions that in a broad sense comprise plate tectonics (Sawkins, 1984). In continental settings the manifestation of an underlying hotspot on the upper crust depends critically on the rates of relative motion between the asthenospheric source of the hotspot and the overlying continental lithosphere (Burke, 1977). Thus lines of basaltic volcanoes may mark a potent hotspot as a continent drifts across it. If the relative motions of hotspots and overlying continental crust are negligible or small, mantle hotspots impinge more substantially on overlying continental areas and appear capable of generating not only various igneous rocks, but also vast volumes of magma (e.g. Bushveld Complex, Stillwater Complex and other layered mafic intrusions). In general hotspots are characterised by alkaline magmatic activity with well developed ring-structures. Eruptive rocks are predominantly rhyolitic with minor trachyte and in some cases basalt; intrusive rocks include carbonatite, per-alkaline granites and undersaturated alkaline rocks together with peraluminous granites. The initial strontium ratios are highly variable, with basalts and some plutonic rocks showing low ratios indicative of a mantle source, while granites mostly have high ratios attributed to crustal anatexis (Mitchell and Garson, 1981).

Mineralisation associated with continental hot spots is almost entirely associated with alkaline intrusives, and lavas flows and peraluminous granites. Tin mineralisation often has accompanying W, U, and Nb, F and base metal sulphides, and may occur in wide variety of environments. For instance, tin associated with the acid phase of the Bushveld Complex occurs in endogranitic pipes and as primary disseminations (Zaaiplaats) or as exogranitic fissure veins, fault breccias, or replacement bodies (Union and Rooiberg Mines; Wilson, 1979; Crocker and Callaghan, 1979; see 4.3.3). The deposits (including fluorite) are genetically related to end member differentiates of the Lebowa Granite Suite (usually volatile-rich biotite granites). The granites are considered to have resulted from anatexis of basement rocks of the Bushveld Complex. The high fluorine content of the rocks is significant, as it is considered to have aided fluxing of the magmas, increasing the volatility, and enabling the fluids to become enriched in tin and other elements by scavenging processes (Wilson, 1979). There is a coincidence of late-stage tin-bearing granites and fluorine enriched tectonic zones. The long lived nature and periodic reactivation of these zones of crustal weakness is critical to the emplacement and localisation of the deposits. Intersections of such lineaments which are held significant in terms of the emplacement of the Bushveld Complex itself (Fig. 3.29; Hunter, 1976; Stear, 1977; Wilson, 1979).

Probably the most significant aspect of Sn mineralisation related to anorogenic settings is the derivation of economically viable concentrations of the metal through anatexis of continental crust, which is also a key facet in elucidating and evaluating Sn-W deposits within orogenic settings.

3.5 Discussion

The outline above of the various tectonic settings incorporates many aspects of Phanerozoic plate tectonics, which are increasingly being applied to Proterozoic mobile belts. In particular there is an often complex interplay of initial extensional (rift) tectonics, with accompanying 'geosynclinal' sedimentation followed by compression,



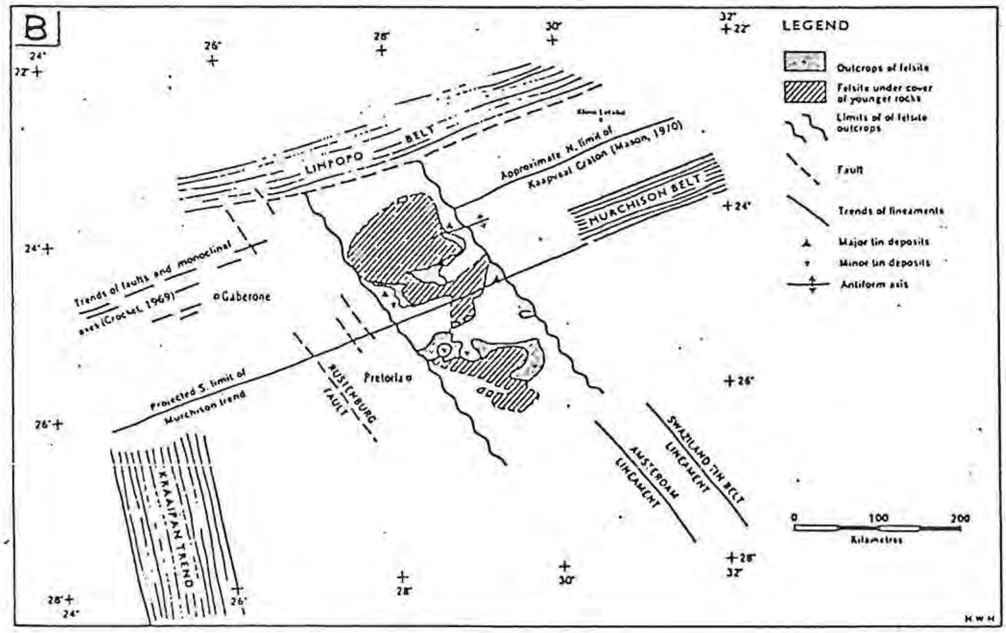


Figure 3.29A: Bushveld tin deposits in relation to the acidic phase of the Complex and with major structures of regional extent (Stear, 1977)

B) Distribution of tin mineralisation relative to lineaments (Hunter, 1976) metamorphism and granitism. Of note is the fact that Sn/W mineralisation is invariably associated with the late stages of granitoid emplacement, and, as may be the case in Namibia (and Nigeria), may also be reconcentrated by later Jurassic-Cretaceous magmatism related to the incipient break up of the South American and African continents. Anorogenic settings are characterised by a lack of deformation/ tectogenesis, although substantial volumes of alkaline magmas are generated through anatexis of the continental (cratonic) crust.

In orogenic settings there are fundamental similarities and differences, and each orogeny effectively must be assessed as a distinct entity. The Variscan Belt and southern African provinces are initially discussed as they do not readily fit into Wilson cycle models, whereas the Malay Peninsula has been conclusively shown to have evolved through various phases of subduction and later collision. A comparison of the evolution of the Variscan and Damaran Belts is given in Fig. 3.30.

The Variscan Belt is of substantial width (1000km in western Europe), and is distinct in its lack of basement inliers. The Belt has a bipolar structure, with a dorsal zone structural parting (the Mid German crystalline rise) from which the vectors of tectonic and sedimentary transport diverge north and south, coupled with an outward migration of regional metamorphism, tectonic deformation, and flysch sedimentation. Of note is that the migration of the tectonic fronts led to a "recycling" of metamorphic and plutonic rocks formed during early Variscan orogeny, resulting for instance in gneisses and anatexites with early Devonian metamorphic ages now being located at the top of nappe complexes, which were stacked during the later Devonian and lower Carboniferous (Martin and Behr, 1983).

The Variscan Belt contains considerably more mafic and ultramafic rocks than the Damara or Saldanian Belts, and the early rift volcanism in the former seems to have been pronouncedly more bimodal and voluminous. This greater magmatic activity in the Variscan Belt is regarded as an indication of greater mantle activity, related to a thinner crust in comparison to the Damara or Saldanian Provinces. The high mantle heat flow also outlasted the Variscan Orogeny, as evidenced by voluminous rocks in intermontane molasse troughs, and further substantiated by the persistence of a compressional regime after the peak of regional metamorphism had passed. In southern Africa heat production may well have been due to radioactive heat production in a tectonically thickened

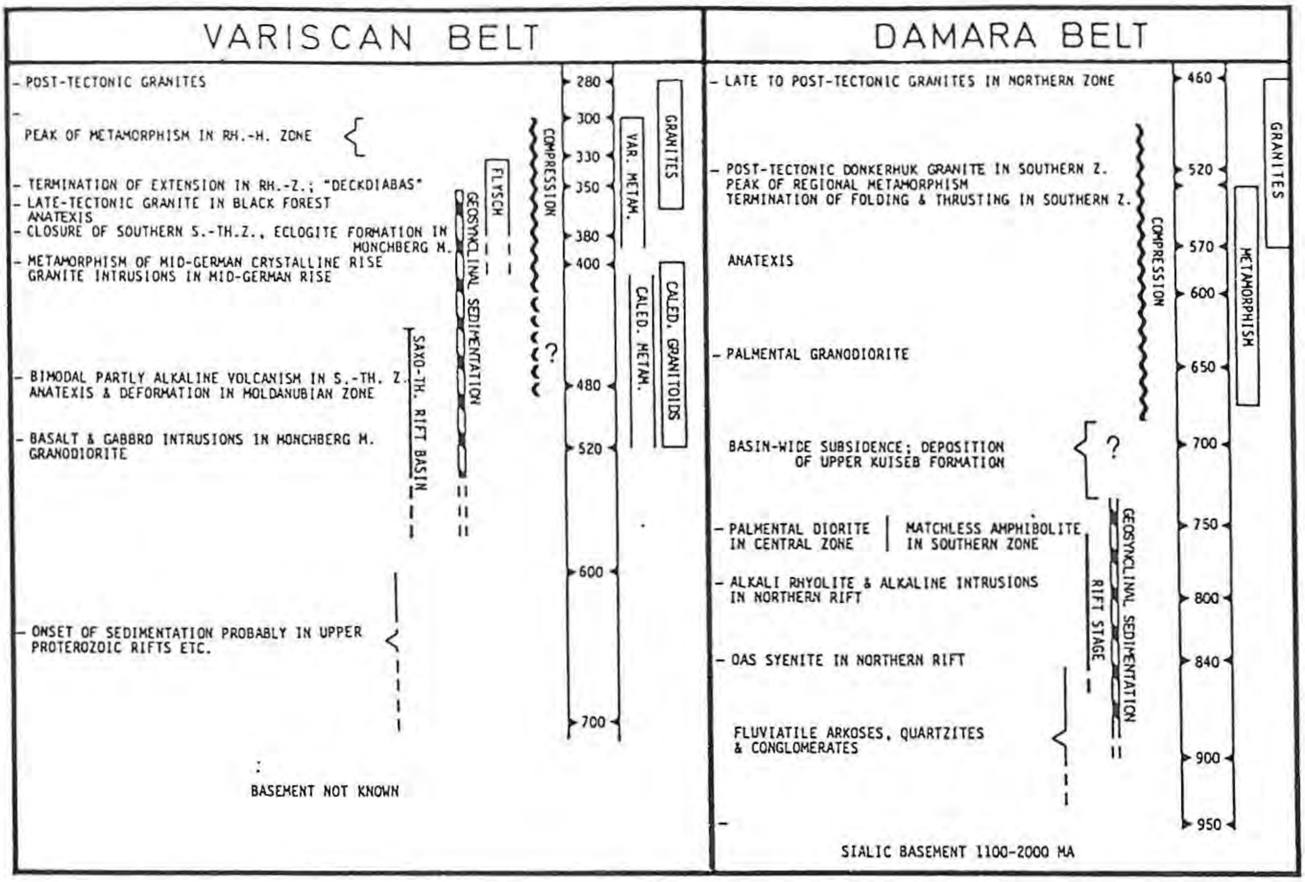


Figure 3.30: Generalised diagrammatic comparison of the evolution of the Variscan Belt with the Damara Belt - RH. - Z. = Rhenohercynian zone; S. - TH.Z. = Saxothuringian Zone; Münchberg M. = Münchberg Gneiss Massif; Var. metam. = Variscan metamorphism; Caled. Metam. = Caledonian metamorphism (Martin and Eder, 1983)

crustal segment with an above average content of radioactive elements. Cratonisation in the Damara and Saldanian Provinces appears complete after the intrusion of post-tectonic granites.

In comparing the Variscan and southern African Belts there are numerous general similarities. Geosynclinal development was initiated by rifting of continental crust, accompanied by the ascent of hot asthenolithic mantle material up to the level of the sialic crust. In the Saldania Province there is a likelihood of slower, prolonged extensionism, and thinning of the crust was probably isostatically compensated by concomittant sedimentation. In the Variscan and Damara Belts rifting was followed by a subsidence stage that lasted 100 to 200Ma. Both regions are characterised by zones of high T/low-medium P metamorphism, and it is conjectural as to whether this is due to "rift metamorphism" or, in the case of the Variscan, from late Caledonian orogeny to the north. In the Damara the lowest pressure zone (Central zone) may be due to uplift while rocks were still semi-molten.

The Variscan, Damaran and Saldanian Belts apparently contain no ophiolites that could be regarded as obducted slices of oceanic crust, as well as volcano-sedimentary mélanges, or plutonic suites which may indicate an island arc or Andean continental margin environment. Both the Variscan and Damara Belts, however, contain a zone in which metatholeiites, characterised by MORB geochemistry, indicate a high mantle temperature underneath a very thin crust. This feature may be the result of either considerable ensialic stretching, or incipient oceanic spreading. There is good evidence for compressional tectonism (high pressure metamorphic rocks in the V.B.), trench deposits, and large scale thrusting, and allochthonous nappe emplacement. Large scale subduction is not envisaged, but features such as major facies gaps in the Variscan Belt (adjacent to the MGCR) indicate telescoping and/or subduction of probably at least 100 to 200km (Martin and Behr, 1983). In the Damara Belt telescoping in the southern part of the Khomas Trough and the southern thrust belt may have attained a similar magnitude. There is no evidence to hand concerning the poorly exposed and studied Saldanian Province.

The Variscan Belt, therefore, is characterised by broad polycyclic orogeny, which is likely to have been dictated by intermittent mantle convection, or the expansion of a convection cell, possibly related to bipolar subduction (after Lorenz and Nicholls, 1984). On the other hand the Damara Belt has developed along a far narrower zone (400km wide vs. V.B. 1000km), probably guided by parallel Irumide (1300Ma) structures, with the development of an ocean arm propagating into the continent from a trilete junction which was eventually fundamental to South America-Africa continental divergence.

The Saldanian Province appears to have formed in similar fashion (tectonically and temporally) to the Damara Province, although this is largely speculative as only part of the rift margin is preserved.

Of pertinence to this dissertation is that termination of all the orogenies (including the Malay Peninsula, see below) is equated within the intrusion of post-tectonic granites. In comparing the Variscan and Damara Belts, Martin and Behr (op. cit.) note that the late syn- to post-tectonic granitic melts are largely generated by anatexis of sialic crustral rocks, and were emplaced over a time span of about 100Ma. The granitoids show no spatial arrangement in terms of their chemical characteristics. Their emplacement was preceded by the initiation of the main compressional phase, which in turn was preceded by a very long geosynclinal (rift-trough) development believed to be associated with concomittant pronounced crustal thinning. Initial rifting of continental crust in both Belts is manifested by bimodal anorogenic volcanic and plutonic rocks.

The Malay Peninsula has more definitive evolutionary traits, which Mitchell and Garson (1981) and Hutchison (1977, 1983) have adequately equated with subduction and collision settings. Hutchison (1977) delineates a paired orogenic batholith belt. The Eastern Belt granites are equated with subduction to the west, whilst the Main Range granites are considered to be related to continental collision in the Late Triassic. The Bentong-Raub ophiolite is interpreted as the remnant suture zone. The setting, including the collision origin of the granites, is analagous to the Himalayas.

The major geological events accompanying collision and of relevance here are underthrusting of the continental foreland on the subducting plate beneath the over-riding plate, and development of new intracontinental thrusts or continental subduction zones, usually within the foreland of the underthrusting plate. This is accompanied by tectonic thickening of crust, regional metamorphism and emplacement of granitic rocks related to shear heating along thrusts (Mitchell, 1977). In the Malay Peninsula there is a conspicuous absence of associated major metamorphic nappes and of inclined granitic sheets. However it is feasible that continued uplift of the Main Range Belt, which has unroofed the granite plutons, may have eroded much of this evidence, and that what is exposed at present is a fairly deep section of the overriding foreland.

When one assesses the occurrence of Sn/W mineralisation in the various tectonic settings outlined above, one important parameter is common. Granite generation, by various processes, is always related to tectonism where there has been significant crustal thickening and general anatexis of continental crust. Geothermal gradients in all settings are variable, but commonly enhanced, be it from asthenospheric diapirism beneath incipient rifting, melting of subducted oceanic crust, or fusion during continental collision. Associated mineralisation appears to always be located in regions where there has been repeated reworking of crustal lithosphere by magmatic processes.

In the Variscan Belt Derré (1982) invokes geochemical anisotropies in rocks older than the poorly differentiated subautochthonous Late Devonian granites which apparently were already anomalous in Sn. Since the early Cambrian Sn-W mineralisation has been concentrated through repeated fractional crystallisation and remelting, and appear to have effectively reconcentrated Sn to a slightly greater degree than W. Aspects of the sources of Sn have been touched on in Section 2.0, (although out of scope of this dissertation) and the reader is referred to Hosking (1979) and Hutchison and Chakraborty (1979) for extended discussion. The present writer follows the latter preferred case, where the continental crust is advocated as an immediate source of tin, and that the element is progressively concentrated by polycyclic events, including variable magmatic and melting processes (see 2.0). This one factor can speculatively be assigned as a common denominator to the tectonic and mineralised settings discussed in the following section.

4.0 Sn-W MINERALISED ENVIRONMENTS

Tin and tungsten mineralisation occurring in the geotectonic environments discussed (above) are reviewed, with comment on some of the more important provinces and/or deposits.

4.1 Variscan granitoids, Western Europe

Important tin and tungsten deposits are associated with Variscan granites in southwest England, the Erzegebirge-Krusne Hory area, Germany and Czechoslovakia, and the western Iberian Peninsula (Figs. 4.1, 4.2, Stemprok, 1981; Derré, 1982). The mineralisation is closely tied to the youngest of two groups of granitic intrusives (260-300Ma), which are predominantly alumina-rich alaskites and lithium albite granites (Sawkins, 1984).

The spatial distribution of Sn-W deposits emphasises an alternating W- and Sn-rich banded pattern (Fig. 4.2, Derré, 1982). These zones have a fairly constant trend (WNW-ESE in the French Massif Central, E-W in NW Iberia) and are generally independent of the Variscan orogenic domains. Derré (op. cit.) explains the distribution in terms of geochemical anisotropies in poorly differentiated pre-Late Devonian granites, already enriched in Sn. Repeated fractional crystallisation and remelting since Cambrian times has reconcentrated the Sn and W. Two differentiated granite suites have attendant Sn-W mineralisation

i) Namurian-Westphalian (Lower to Middle Carboniferous) in the French Massif Central
ii) Westphalian-Autunian (Middle Carboniferous-Permian) in the NW Iberian Peninsula

The Namurian granites and mineralisation have been emplaced along WNW-ESE structures, whereas later stage more differentiated mineralised cupolas in the roof zones of the granites follow NNW-SSE and NNE-SSW trends (Derré, 1982). Fig. 4.3 summarises the different metallogenic and structural alignment of granitoids, whilst Fig. 4.4 illustrates a N-S section of the Massif Central and the intrusion of the two mineralising phases envisaged by Derré (op. cit.).

The types and characteristics of the major European Sn-W deposits are shown in Table 9. The most important deposits are spatially associated

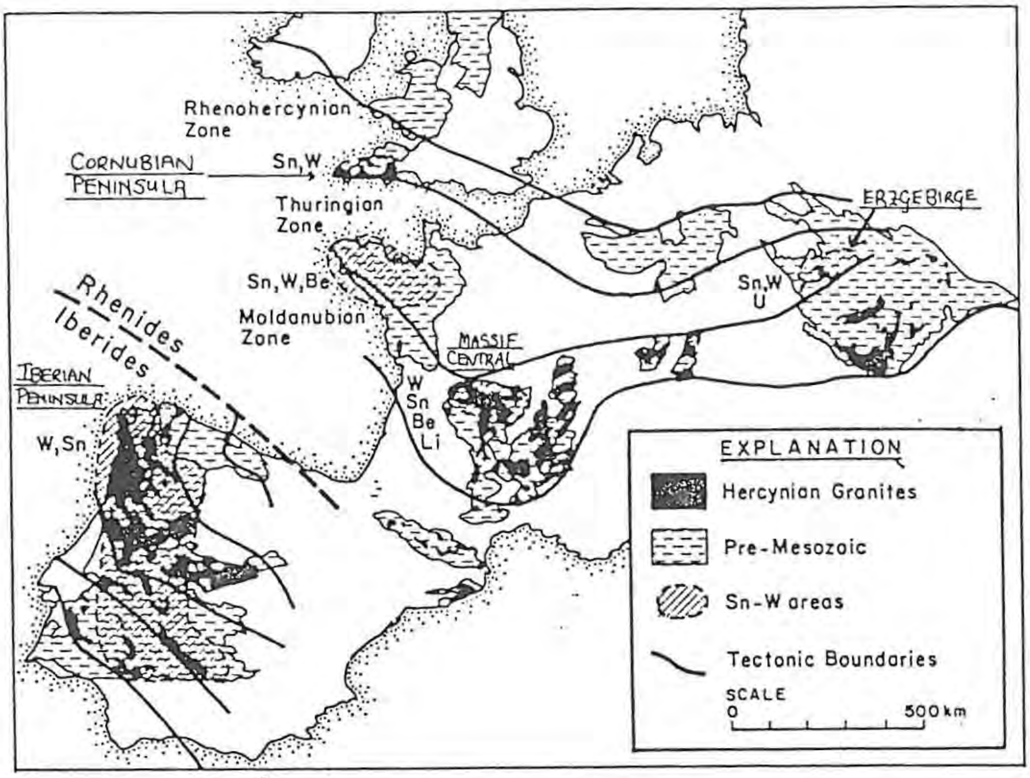


Figure 4.1: Pre-Mesozoic and Variscan (Hercynian) granitoids in western and central Europe (in black) and the distribution of Sn-W provinces (Stemprok, 1981; Sawkins, 1984)

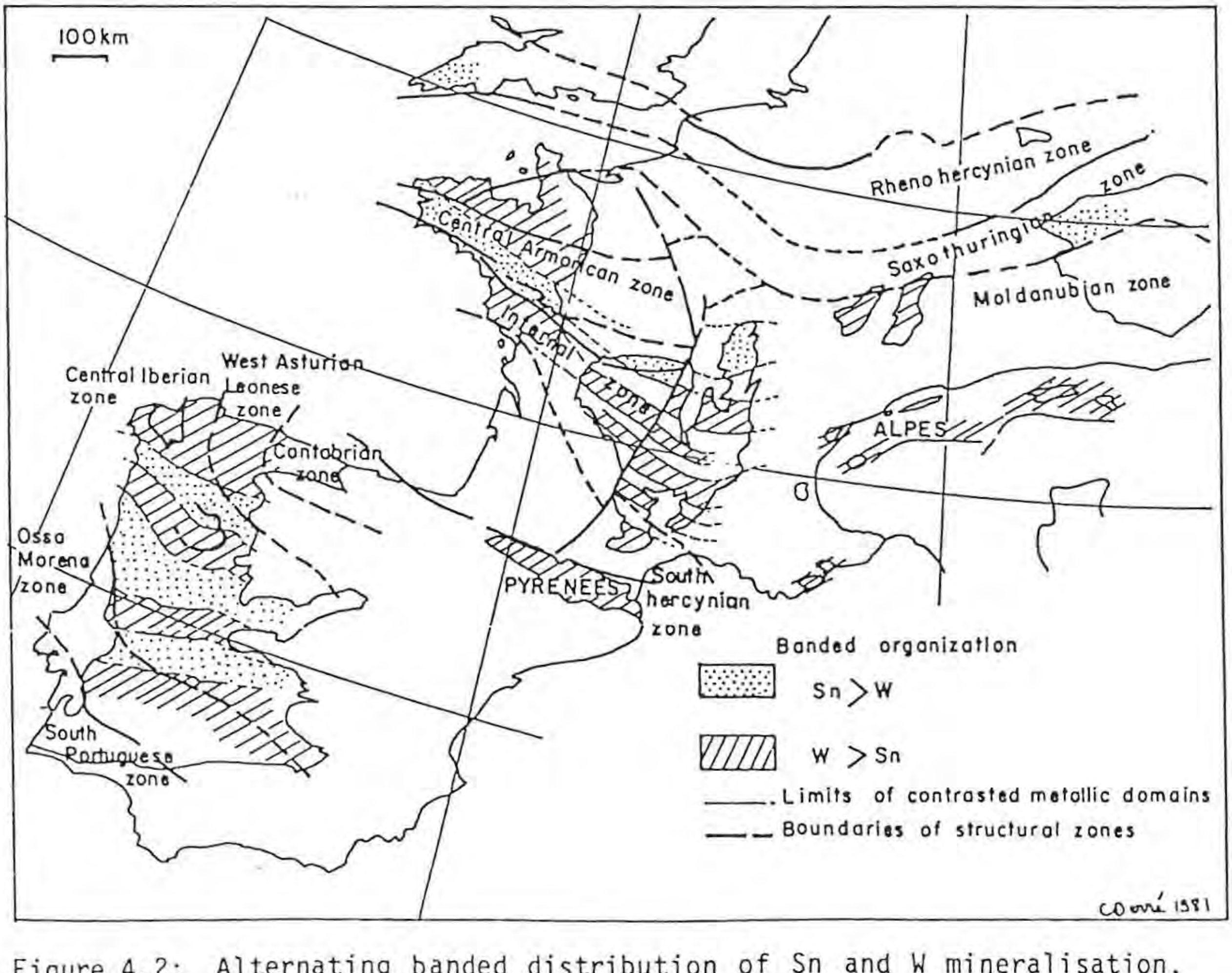


Figure 4.2: Alternating banded distribution of Sn and W mineralisation, oblique to Variscan orogenic structures (Derré, 1982)

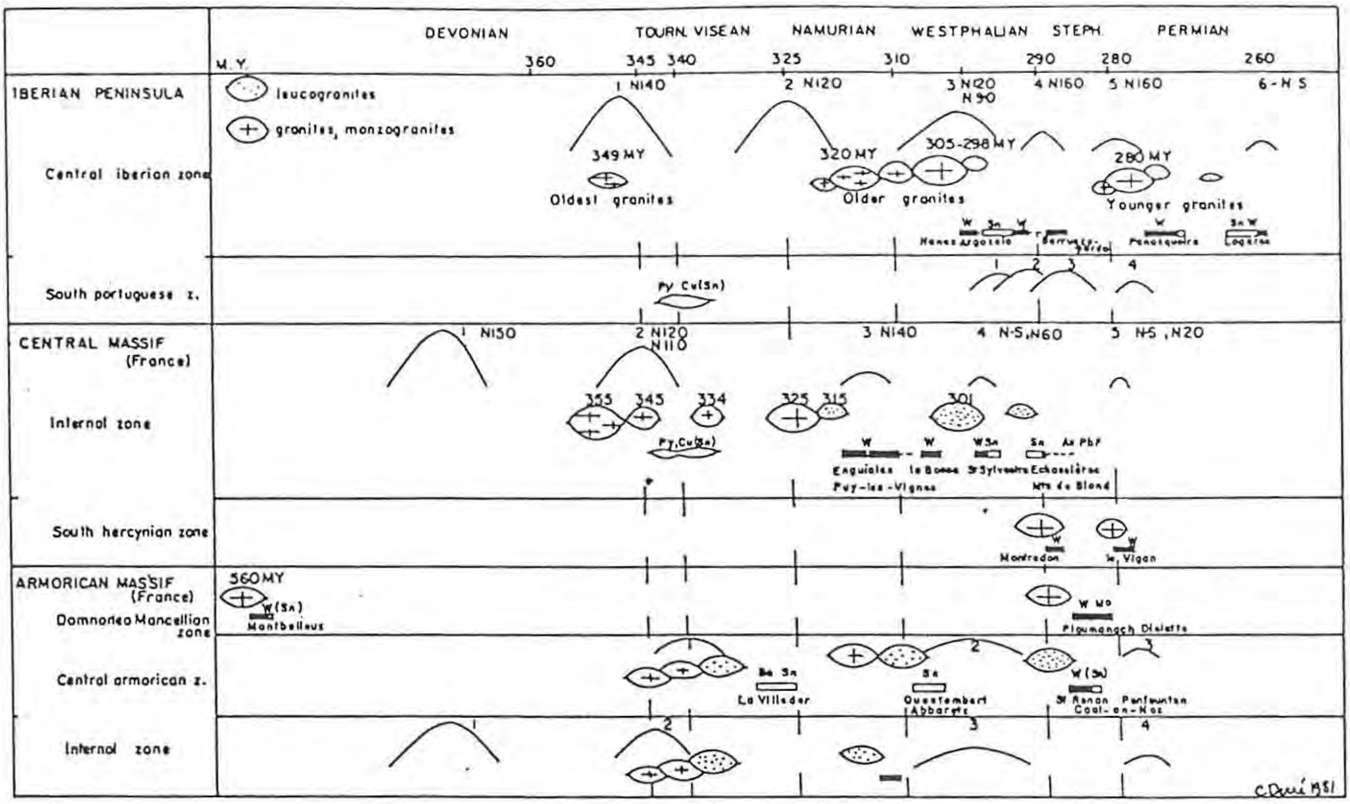


Figure 4.3: Metallogenic epochs of Sn-W mineralisation and alignment of associated granitoids (Derré, 1982)

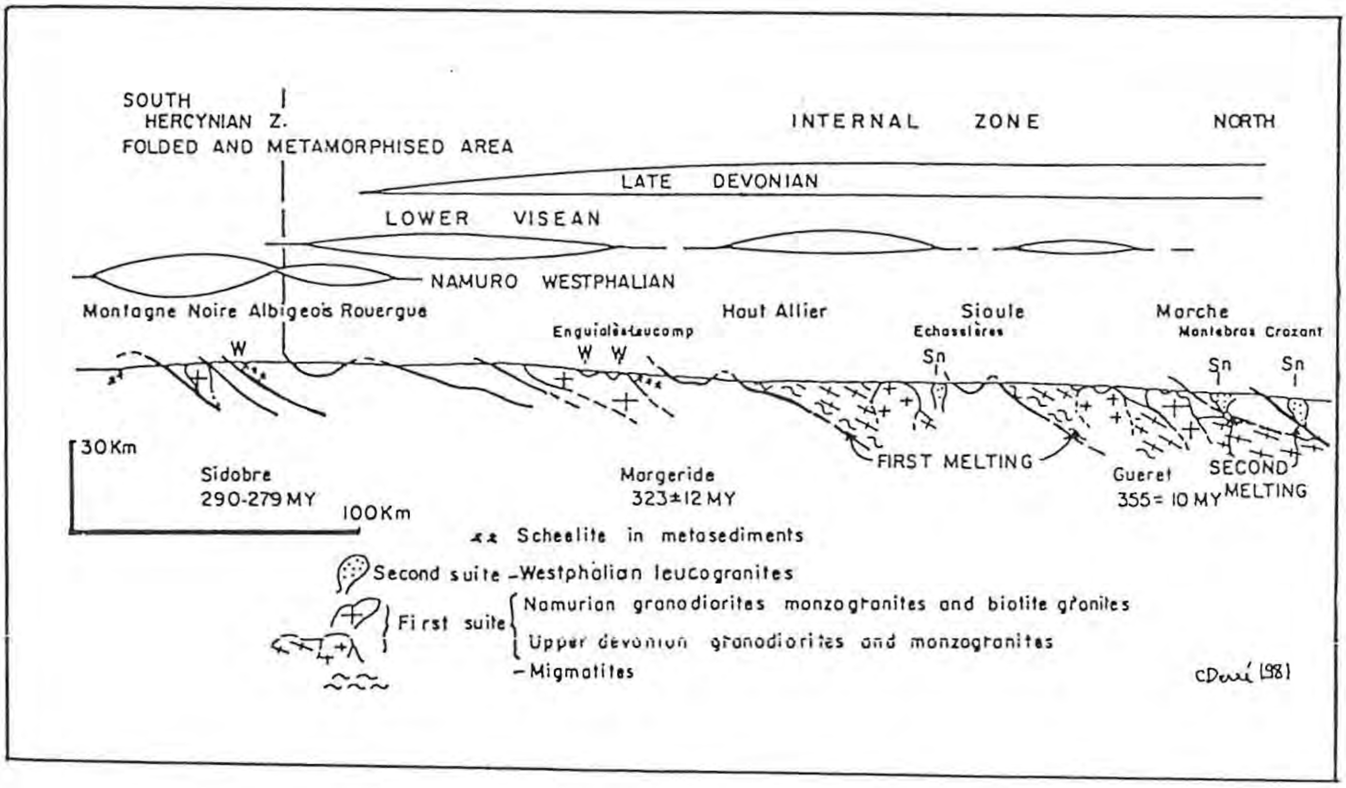


Figure 4.4: Schematic N-S cross-section of the Massif Central (France) illustrating the emplacement of the two phases of mineralised granites (Derré, 1982)

with intrusive contacts, occurring as stockworks, vein complexes, or single veins, characterised by tourmalinisation and greisen-type alteration (Stemprok, 1981). Skarn-type deposits are formed locally where limestones occur adjacent to igneous contacts. In these and greisen deposits scheelite is the primary tungsten mineral. Vein deposits generally contain cassiterite, wolframite, and variable amounts of sulphides, including arsenopyrite, pyrite, pyrrhotite, sphalerite, galena, chalcopyrite, bismuthinite and stannite. A broad metal zonation in both a vertical and lateral sense is sometimes discernable - with a central concentration of Sn-W, and peripheral base metals.

Table 9: Types and principal characteristics of Sn-W deposits associated with Variscan granitoids in western Europe (adapted from Derré, 1982) (N.B. Translated from French)

TYPE	PRINCIPLE CHARACTERISTICS	AGE
1. VEIN wolframite	PORTUGAL: Panasquira, principle producer of Portugal 36000t concentrates 1934-1975. Precambrian to Early Cambrian greywacke-schists. Non outcropping greisenised - leucogranite cupola.	Autonian
	Boralha, 2nd largest producer of Portugal, 12000t Wo_3 Veins bordering polyphase granitic complex	Autonian
	Vale das Gatas 670t conc. 1959-1970. Several veins (57-63) within older granite cross-cut by younger granite	Autonian
	SPAIN: Barruecopardo veins with scheelite in two-mica leuco-granite	Middle Stephanian
	FRANCE: Enguiales-Leucamp, 1500-1700t WO_3 produced Veins in schists near to porphyritic monzogranite	Late Namurian
	Montredon - Labessonnie, Reserves 10,500t WO_3 Veins in micaschist and orthogneiss	Late Westphalian
	Phy-les-Vignes, 5,400t WO_3 produced 1905-1956	Late Namurian
2. VEIN Cassiterite	PORTUGAL: Montesinho: largest producer 1,825t conc. 1959-1970 Argozelo-Ribeira: 2nd producer, 2,268t cass. conc. and 1,121t wolfram conc. 1959-1970	Late Westphalian

Table 9 continued

TYPE	PRINCIPLE CHARACTERISTICS	AGE
	SPAIN: Penouta, possible reserves 2000t Sn in veins, 1500t Sn in alluvials. Vein stock-work, alluvial and eluvial deposits	
	San Finx, 2000t Sn produced Sn/W ratio between 85/15 and 40/60	
3. TACTITES scheelite	FRANCE: Salan (Pyrenees), largest producer in France over 15000t WO_3 combined production and reserves	Late Westphalian
4. ALBITISED GRANITE CUPOLAS Disseminated cassiterite and veins with wol-framite	FRANCE: Echassieres, Reserves 7,200-9000t Sn 48,000t Li. plus veins with wolframite (6000t WO_3 produced)	Late Westphalian
	SPAIN: Santa Comba, W/Sn ratio 10/7 4000t WO_3 1500t Sn produced? Fontao W/Sn ratio 3/2, 3,600t WO_3 1200t Sn produced?	
5. PEGMATITE Cassiterite	PORTUGAL: Amarante 1467t conc. produced Lagares 150t conc. produced	Autunian

4.1.1 Panasqueira Sn-W deposits, Northern Portugal

The Panasqueira district is discussed here as it is the largest producer of tungsten in western Europe, (around 1250t wolframite, 62t cassiterite per annum) and has been thoroughly studied in terms of its geology, mineralogy and fluid inclusions by Kelly and Rye (1979).

The Panasqueira veins overlie a hidden cupola of strongly altered and greisenised granite, apparently intruded approximately 290Ma ago (Fig. 4.5). The host Ordovician Beria schists of greenschist facies, exhibit a broad spotted hornfels contact metamorphic aureole, indicative of more extensive intrusion at depth. The area is also intruded by post-granite dolerite and aplite dykes.

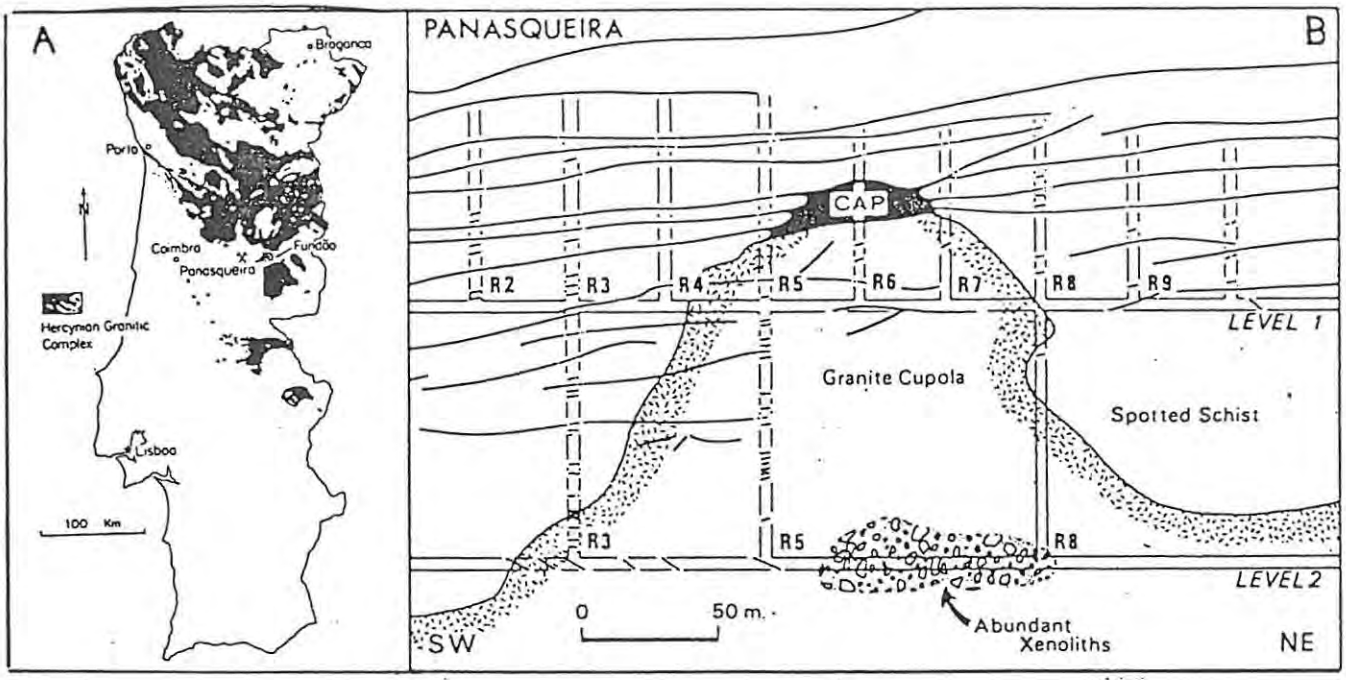


Figure 4.5A: Location of Panasqueira, north Portugal, in relation to the Variscan granites

B: The W-Sn flat-lying Variscan vein system, the granite cupola and silica cap (Kelly and Rye, 1979)

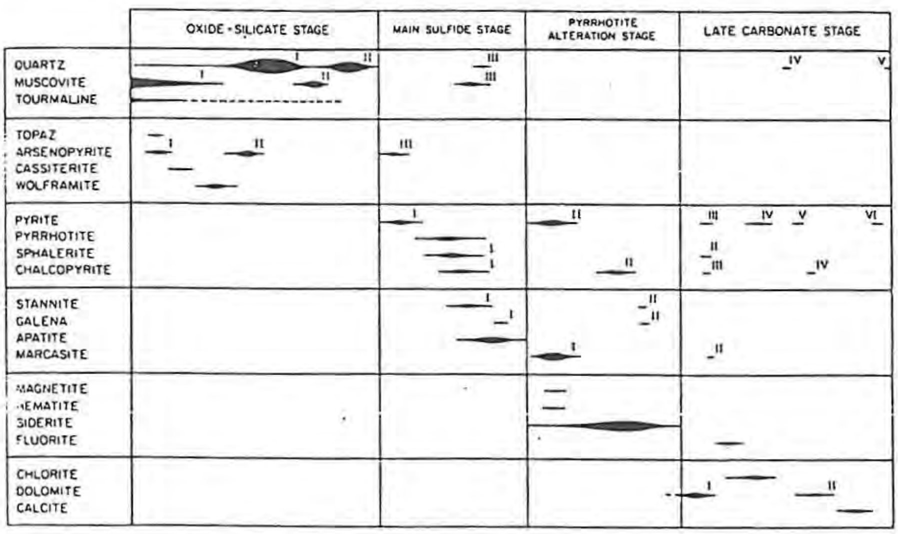


Figure 4.6: Paragenesis of the Sn-W veins, Panasqueira. Roman numerals identify multiple generation of a given mineral (Kelly and Rye, 1979)

The vein system at Panasqueira is unusual in that it is nearly horizontal, with the ferberite-quartz veins filling openings created by vertical dilation of pre-ore joints, post-dating metamorphism, and thought to have formed by the reduction of overburden pressure (Kelly and Rye, 1979). A highly silicified cap occurs at the roof of the cupola, but its mineral content is subeconomic. Economic veins are laterally extensive but restricted to a narrow vertical zone (100-300m) in the schists above the cupola (Fig. 4.5).

Kelly and Rye (op. cit.) also conducted fluid inclusion sulphur- and oxygen-isotope studies. The vein fluids were NaCl brines (5-10% equiv. Wt% NaCl) with deposition within the range 360-230°C. Vein deposition during the late carbonate stage was at lower temperatures (less than 120°C), and salinities (less than 5 equiv. wt% NaCl). The isotope studies indicated the fluids were predominantly composed of meteoric water in the carbonate veins, but the higher temperature Sn-W deposition was related to deeper-seated magmatic water. The ore mineral depositional phases indicate mixing of magmatic and meteoric waters as the system cooled beyond the sulphide stage (Fig. 4.6).

4.1.2 Erzgebirge, Czechoslovakia and German Democratic Republic

The Erzgebirge tin province (Fig. 4.1) is discussed briefly as it constitutes one fo the classic greisen-type areas of Sn-W mineralisation in the world. The general environment includes a long lived mobile belt consisting of Proterozoic to Lower Ordovician metasediments and Lower Carboniferous (360-310Ma) granitoids. The metasediments (progressively metamorphosed to the NE from greenschist to granulite facies) consist of phyllites, metagreywackes, metaconglomerates, metaquartzites, carbonate rocks, schists, gneisses etc within an NE anticlinorium plunging Sw (Baumann, 1970). The granitoids batholiths were intruded as a series of 15-20 complexes within a ENE trending zone 30 x 100km, within which younger phases are elongate NW. The thickness of metasedimentary roof pendants varies, so that in the NE apical stocks are exposed, whilst in the SW larger granitic complexes are present (Baumann et al., 1974). Older and younger granitoid phase are distinguished, both composed predominantly of monzogranites (Taylor, 1979).

Mineralisation is spatially related to 20-30 discrete centres, mostly located within a NE trending zone 15 x 150km. It is manifest as Sn, W

±Mo, Bi, As, Fe, with minor Cu, Pb, Zn, Ag and located mainly on NE striking lineaments. The economic mineralisation is associated with the younger granite complexes, which are also geochemically specialised in elements such as F, Sn, Rb, Li (Baumann et al., 1974).

The principal types of mineralisation are:

i) Massive greisens associated with quartz veins (classic greisen), major economic concentrations of Sn, W, ±Mo, As, Fe, Cu, Bi, Zn.

ii) Skarn assemblages associated with granitoids - minor Fe, Sn, W, Zn mineralisation.

iii) Sulphide rich veins related to i) above, of minor economic importance containing Fe, Sn, Cu, Pb, Zn, Ag.

iv) Stratabound massive sulphide ore and impregnations associated with rhyolite ± schists gneisses and carbonates (Baumann et al., 1974).

A vertical zonation of minerals is suggested for the vein systems. Topaz-mica (endogreisen) is succeeded by tourmalinisation, chlorite-sulphide assemblages and sulphides (exogreisens).

4.1.3 Cornubian tin field, south-west England

Comprehensive reviews of the mineral deposits of the Cornubian tinfield are afforded by Hosking (1970), Dunham et al. (1978), and Jackson (1979), whilst Exley et al. (1983) and Floyd et al. (1983) have synthesised the Variscan magmatism and attendant mineralisation.

Mineralisation is spatially related to the roof and border zones of a series of high level, post-tectonic granite plutons, which together form the Cornubian Batholith (Fig. 4.7). Six major and several minor masses of granite are discernable, and, together with their associated dykes and sheets, were emplaced partly passively, by cauldron subsidence and block stoping, and partly forcefully, into a deformed Devonian-Carboniferous volcano-sedimentary sequence (Jackson, 1979). The plutons are composite in nature, with roughly arcular or ellipsoid shape, and invariably have sharp and discordant contacts with the surrounding rocks. Emplacement ages are spread between 303-265Ma (K-Ar), with a Rb/Sr isochron of 285

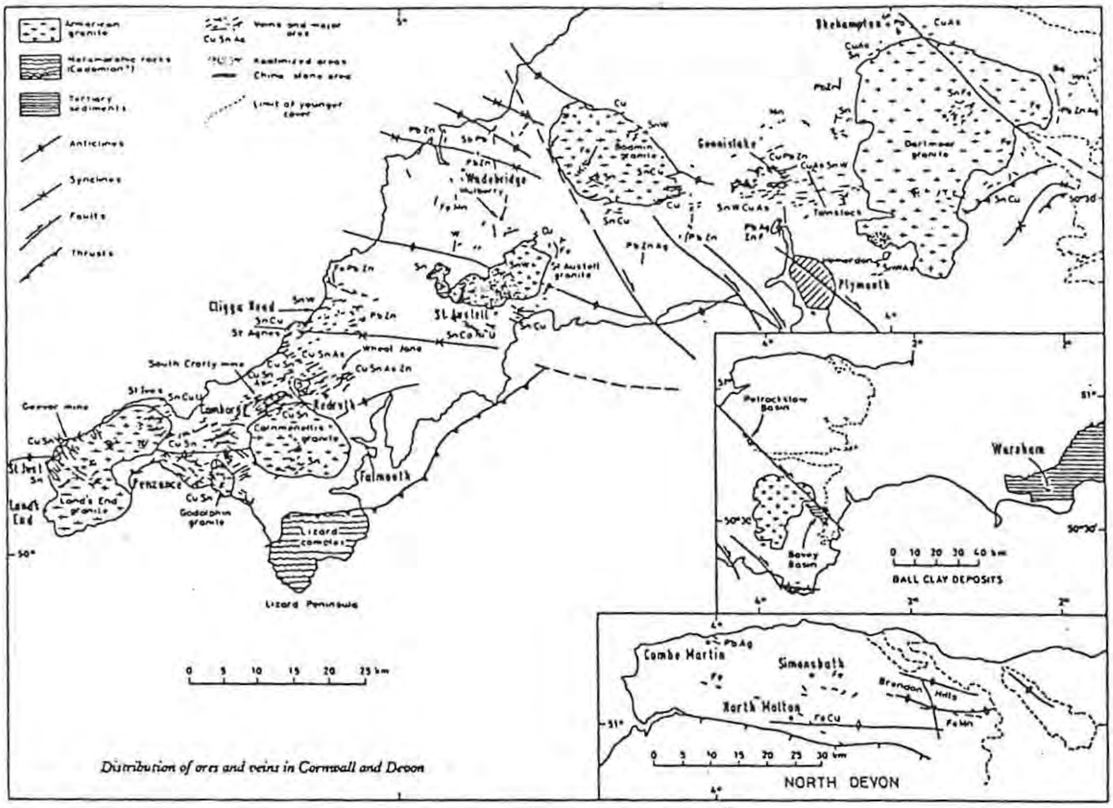


Figure 4.7: Distribution of ores and veins in SW England (Dunham et al., 1978)

$\pm$12Ma (Exley et al., 1983). The "best" age of granite emplacement appears around 295Ma, followed by pegmatites (285Ma), greisen veins between 285-280Ma, granite porphyry dykes at 275-170Ma followed by mainstage mineralisation at $\pm$270Ma (Halliday, 1980).

Petrography and geochemistry

The temporal relationship between the plutons has been elucidated by field and petrographic studies. Exley et al. (1983) distinguish six principal types (A-oldest to E-youngest) on the basis of petrography (Table 10). Type B plutons are the predominant granite-type exposed. The granites were further grouped by cluster analysis on a geochemical basis (Table 11) into three district series:

i) coarse- and fine-grained biotite granites, including the megacrystic Li-mica granite of St. Austell (Fig. 4.7, i.e. types B, C, and D, Tables 10, 11);
ii) equigranular Li-mica and fluorite granites (E and F plus Li-mica leucogranite and Carnmenellis aplites);
iii) granite porphyry ('elvan') dykes (Exley et al., op. cit.).

Table 10: Petrographic summary of main granite types (Exley et al., 1983)

Type	Name	Texture	Minerals (approximate mean modal amounts in parentheses): K-feldspar	Plagioclase	Quartz	Micas	Tourmaline	Other
A	Basic microgranite	Med. to fine Ophitic to Hypidiomorphic	(amounts vary)	Oligoclase–andesine (amounts vary)	(amounts vary)	Biotite predominant some muscovite	Often present	Hornblende Apatite Zircon Ore Garnet
B	Coarse-grained megacrystic biotite granite	Med. to coarse Megacrysts max. 5–17cm, mean about 2cm. Hypidiomorphic, granular	Eu- to subhedral Microperthitic (32%)	Eu- to subhedral. Often zoned. Cores An_{35}–An_{36}. Rims An_{4}–An_{33} (22%)	Irregular (34%)	Biotite, often in cluster (6%) Muscovite (4%)	Eu- to anhedral. Often zoned. 'Primary' (1%)	Zircon Ore Apatite Andalusite etc. } (1%)
C	Fine-grained biotite granite	Med. to fine Sometimes megacrystic, Hypidiomorphic to aplitic	Sub- to anhedral Sometimes microperthitic (30%)	Eu- to subhedral. Often zoned. Cores An_{16}–An_{35} (26%)	Irregular (33%)	Biotite (3%) Muscovite (7%)	Eu- to anhedral 'Primary' (1%)	Ore Andalusite Fluorite (<1%)
D	Megacrystic Li-mica granite	Med. to coarse Megacrysts 1–8.5cm, mean about 2cm. Hypidiomorphic, granular	Eu- to subhedral Microperthitic (27%)	Eu- to subhedral Unzoned, An_7 (26%)	Irregular. Sometimes aggregates (36%)	Li-mica (6%)	Eu- to anhedral 'Primary' (4%)	Fluorite Ore Apatite Topaz (0.5%)
E	Equigranular Li-mica granite	Medium-grained. Hypidiomorphic, granular	Anhedral to interstitial Microperthitic (24%)	Euhedral. Unzoned, An_4 (32%)	Irregular. Sometimes aggregates (30%)	Li-mica (9%)	Eu- to anhedral (1%)	Fluorite Apatite } (2%) Topaz (3%)
F	Fluorite granite	Medium-grained, Hypidiomorphic, granular	Sub-anhedral Microperthitic (27%)	Euhedral. Unzoned, An_4 (34%)	Irregular (30%)	Muscovite (6%)	Absent	Fluorite (2%) Topaz (1%) Apatite (<1%)

Table 11: Average analyses of granitic rocks from southwest England (Exley et al., 1983)

	Type B			Type C		Type D		Type E granite	Type F	Granite porphyry	Micro-granite	Av. granite*
ref no[1]	18	9	39	42	21	13[2]	35	5	15[2]	34	27[3]	
wt%												
SiO_2	72.43	72.63	71.20	73.70	74.08	73.01	72.73	71.10	74.20	72.80	72.80	70.58
TiO_2	0.21	0.28	0.35	0.06	0.07	0.14	0.13	0.06	0.07	0.20	0.04	0.36
Al_2O_3	15.03	14.65	14.20	14.10	14.76	14.72	14.85	16.11	15.81	14.50	16.40	14.47
Fe_2O_3	0.32	0.50	0.80	0.60	0.19	0.47	0.34	0.35	0.08	1.85 }	0.84[4]	0.66
FeO	1.48	1.24	1.38	0.44	0.86	0.74	0.94	0.81	0.17	1.21 }		1.92
MnO	0.04	0.05	0.03	0.03	0.03	0.03	0.03	0.07	0.01	0.05	0.09	0.06
MgO	0.44	0.48	0.60	0.05	0.18	0.14	0.33	0.09	0.08	0.26	0.05	0.52
CaO	0.84	1.12	1.12	0.56	0.44	0.44	0.41	0.59	1.31	0.28	1.28	1.09
Na_2O	3.11	3.11	2.82	2.86	2.74	3.42	3.21	3.73	4.06	0.12	2.77	3.08
K_2O	5.06	4.36	5.11	4.77	5.73	5.36	5.03	4.84	4.66	7.66	3.95	4.67
Li_2O	0.06	0.07	0.08	0.07	0.04	0.18	0.11	0.27	0.01	0.03	0.94	
P_2O_5	0.25	0.18	0.24	0.32	0.25	0.33	0.15	0.50	0.46	0.26	0.48	0.18
B_2O_3	—	—	0.41	0.47	—	—	0.27	0.14	—	—	—	
F	—	—	—	—	—	(0.38)	0.38	1.22	(1.36)	—	1.40	
H_2O+	1.01	—	0.73	1.38	0.88	—	1.13	—	—	—	—	
ppm												
Nb	—	17	30	40	—	57	—	93	81	21	67	20
Zr	121	137	185	40	34	(50)	65	46	(11)	94	38	154
Y	41	48	30	20	40	—	—	10	—	18	—	—
Sr	94	92	95	22	43	41	175	61	64	34	47	100
Rb	419	462	480	760	444	982	695	1218	615	814	2293	170
Ba	195	397	230	15	102	(83)	150	204	(43)	699	197	380
La	31	16	—	—	12	8	—	~3	—	14	15	—
Ce	38	—	—	—	2	34	95	36	19	68	27	—
U	—	—	—	—	—	—	—	19	—	20	24	6
Th	—	—	—	—	—	—	—	22	—	31	—	19
Pb	46	47	15	10	42	—	—	16	—	6	5	19
Ga	—	40	30	30	—	—	40	40	—	20	35	19
Zn	62	72	45	35	48	—	103	48	—	45	31	39
Ge	—	—	—	—	—	—	—	11	—	4	11	—
Sn	23	14	19	17	29	—	40	36	—	71	14	3
Cs	28	34	—	—	48	—	—	127	—	33	223	4
K/Rb	100	78	88	52	107	45	60	33	63	78	14	170

W 2
Mo 1

* Sources: Wedepohl, K.H. (Ed) 1969: Handbook of geochemistry
Tauson, L.V. 1974: MAWAM Vol. 1 p. 222–223
Kozlov, V.D. 1974: MAWAM vol. 1 p. 202–203
Klominsky, J and Absolonova, E. 1974: MAWAM vol. 1 p 191
(MAWAM: Metallisation associated with acid magmatism)

Analysis of Table 11 indicates the Cornubian rocks are enriched in several elements compared to others and the world "average" granite. For example Li, Rb, F are often in the % range, and are concentrated within micas, especially in D-, and E-type granites. High trace alkali values include Cs, Li, Rb. Low K/Rb, coupled with high Ba/Sr and Rb/Ba (over 5) are an indication of extreme differentiation of some of these granites. Zr and TiO_2 show good linear relationships with concomitant decrease in both from B to E to leucogranites and aplites, and the Zr/TiO_2 ratio further attests to the differentiated nature of the granites (see El Bouseily and Sokkary, 1975 for a discussion on Rb, Ba, and Sr in granitic rocks).

Ore-forming metals such as Sn (Li-mica granites) and Pb, Zn (biotite granites) show an overall enrichment relative to world granites. Unfortunately there is a lack of W analyses. Both boron (B) and fluorine (F) show high concentrations, corresponding to the presence of tourmaline, (and other boric silicates) and fluorite, whilst enrichment in Cl is found in inclusions (Exley et al., 1983).

Mineralisation and post-magmatic alteration

A wide variety of Sn- and W-bearing deposits occur at Cornwall. Hosking (1970) classified them into pegmatites, pyrometasomatic deposits, greisen-bordered vein swarms, stockwork deposits, fissure veins of Cornish-type lodes, and pipe and carbona deposits. Figure 4.8 illustrates a diagrammatic representation of the primary deposits, the various ore depositional environments, and the distribution and relative chronology of the mainstage mineralisation. Table 12 summarises the main features of the deposit types.

The fluids and solutions involved in the evolution of the batholithic and associated rocks resulted in a continual sequence of widespread alteration and mineralisation, and there has been much overlap and repetition of late- and post-magmatic processes.

Primary alteration processes can be differentiated (as pre-joint development) from secondary (post-joint development). The former are characterised by tourmalinisation and greisenisation. Tourmaline is present in different rocks, effectively reflecting magmatic changes related to fractionation, producing the sequence granite-pegmatite-aplite-

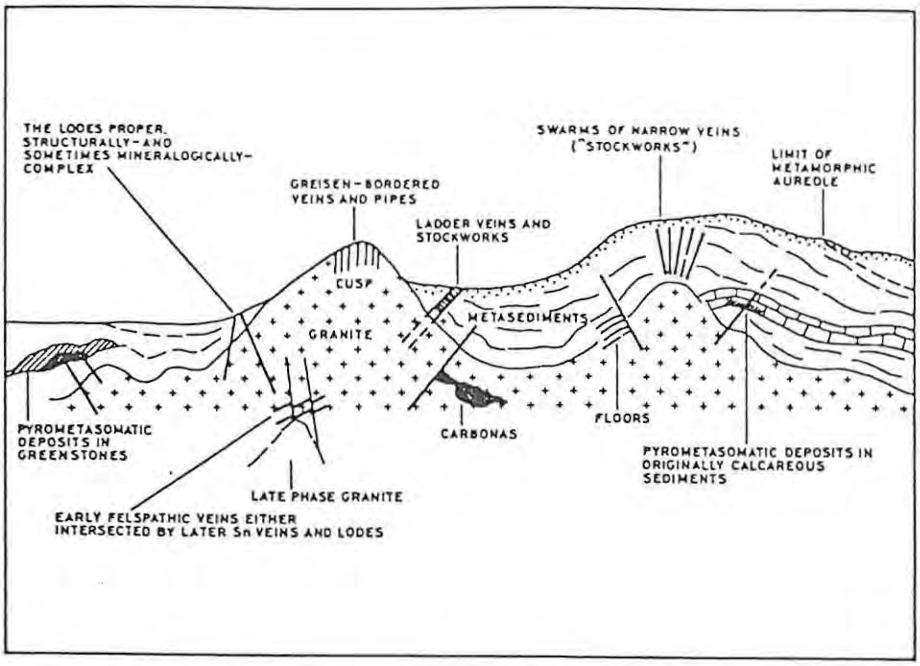
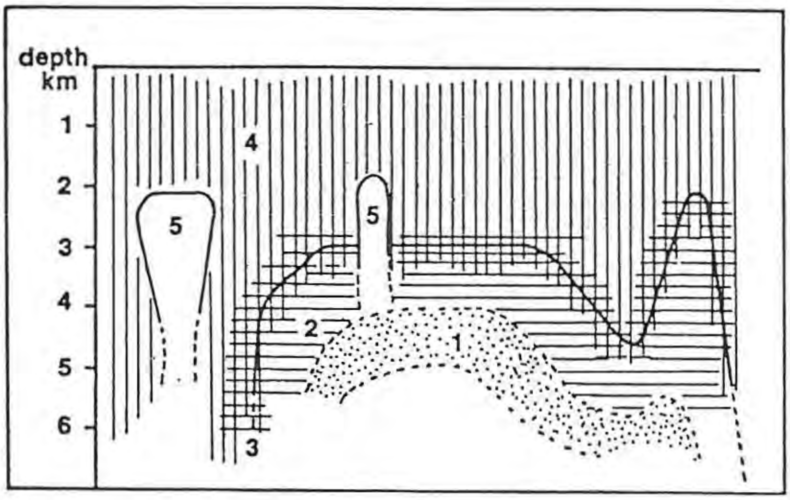
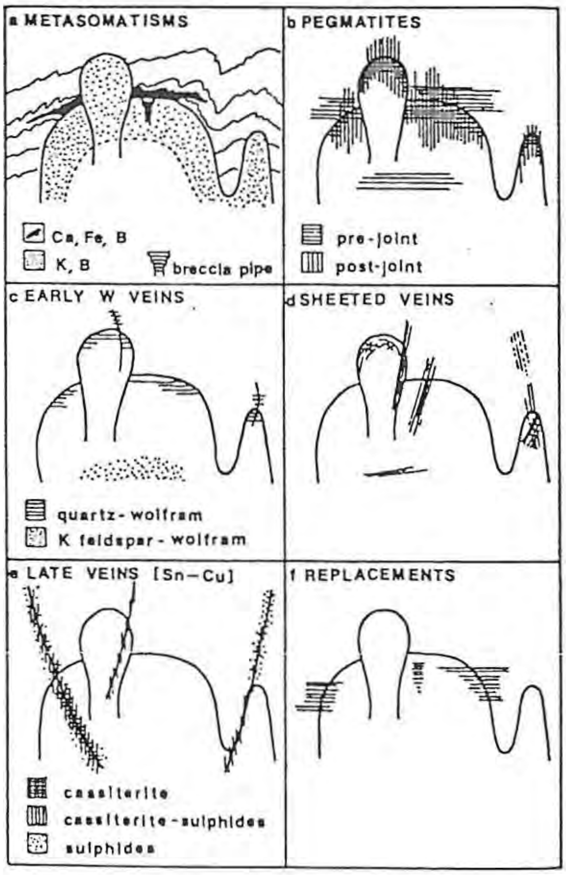


Figure 4.8A (above): Diagrammatic representation of primary tin deposits in south-west England (Hosking, 1970).
B (below): Ore depositional environments, Cornubian Batholith: 1) internal zone; 2) internal contact zone (endogranitic) 3) contact zone 4) external contact zone (exogranitic) 5) porphyry stocks and cupolas (after Jackson, 1979)





C (Right): Distribution and relative chronology of mainstage mineralisation phenomena (Jackson, 1979)

Table 12: Salient features of Cornish Sn/W ore deposits

CLASSIFICATION	CHARACTERISTICS
Pyrometasomatic	Few, relatively unimportant. Skarn type min po-aspy-sp-py-cpy within garnet-diopside-actinolite-wollastonite-axinite skarn. Local occurrences of Sn as malayaite ($Ca\ Sn\ SiO_5$)
Pegmatites	Insignificant, scattered small lenses close to granite contacts. Local wolframite, aspy.
Greisen-bordered Vein swarms	Host subparallel quartz veinlets in granite, walls of greisen or narrow selvages of chlorite rosettes, tourmaline needles, feldspar and fluorite. Usually have central vein, ores cassit, wolf, aspy, cpy, cc, stannite ($Cu_2Fe\ SnS_4$). Greisen walls may contain f.g. cassit. and sulphides. Vein swarms dip steeply, usually pinch out in granite. Grades 0.15-0.40% Sn.
Stockworks	Networks of narrow mineralised structures occurring within any country rock-type. Some assoc. with major hydrothermal veins, or else preferentially along set of joints. Veins contain cassiterite, wolframite and arsenopyrite.
Fissure veins	Primary source of ore. Tabular, generally less than 2m wide, strike lengths up to ±5km, dip extent to ±1km. Fill congugate fracture pattern in the country rocks. No consistent relationship between granite-shape and vein trend, no radial pattern around a pluton. Cross veins, ± perpendicular to lodes, usually barren. Polyphase deposition indicated by banded structures, internal brecciation and recementation. Main ore minerals chalcopyrite, cassiterite, arsenopyrite.
Pipe and carbona deposits	Uncommon, usually small, endogranitic and in association with fissure veins. Pipes are vertical cylindrical, ±130m high and 4m wide, have core 1cm wide vein quartz and clay surrounded by altered granite with cassiterite, ± sulphides and fluorite. Carbonas similar, but have irregular shape and attitude and contain tourmaline

quartz-schorlite rock. Primary greisenisation is widespread in younger varieties of granites, but proceeded only as far as the replacement of perthite by 2° mica and quartz.

Secondary processes are largely confined to joint systems. A second stage of greisenisation, marked by greisen flanked veins, is found in all granites. The joints are commonly occupied by veins containing a gangue of quartz, tourmaline, chlorite $\pm$ oxide or sulphide ore minerals. Greisenised borders are only a few centimetres wide, marked by a general enrichment in minerals such as muscovite, tourmaline, fluorite, and topaz. Greisenisation, essentially the breakdown of feldspar to secondary mica, occurs at 350-450°C (Grim, 1968; Shcherba, 1970). Secondary tourmalinisation succeeds secondary greisenisation in the same joint system.

These steeply dipping joint systems and faults were also the locale of ore mineral deposition (stockwork and lode-types, Table 12). A very complex and variable chronology exists, with several generations of vein mineral deposition, and, in addition to replacements, there are reversals in trend (Exley et al., 1983). The general paragenesis is as follows (Table 13): the earliest minerals are found commonly as replacements in the greisen flanked veins, including cassiterite, wolframite, arsenopyrite and molybdenite, indicating greater enrichment of the fluids in Sn, W, As, Fe, Mo and S compared to those responsible for the greisens themselves. Stages of mineralisation are recognisable at 400-500°C, with fluids of moderate salinity (20-24 equiv. wt% NaCl); and 280-400°C, fluids of variable salinity (8-14 to 24-40 equiv. wt% NaCl) (Jackson et al., 1982). Hypo-, meso- and epithermal zones of mineralisation (and alteration) are recognised (see Table 13; Jackson, 1979, et al., 1982). The overall ore distribution is complex, and concepts based on concentric zones (e.g. W/Sn core, Pb/Zn outer) determined by T-P conditions, arranged around "emanative" centres provided by granite cupolas is probably an oversimplification. A broad pattern can be discerned on a regional basis (Fig. 4.9). The three mineralising processes outlined in Table 13 were superimposed by kaolinisation, conjecturally a combination of deep weathering and hydrothermal alteration (i.e. acidic conditions at less than 350°C) (Jackson, 1979; Exley et al., 1983).

Table 13: Generalised ore-mineral paragenesis in southwest England (Exley et al., 1983)

	Greisen-flanked veins	Hypothermal				Mesothermal		Epithermal	
Zone		1	2	3	4	5a	5b	6	7
Homogenisation of inclusions (°C)	250–500						150–350		150
Salinity of fluids (Equiv. wt % NaCl)	8–40						0–10		25
Gangue minerals	Feldspar Mica Tourmaline		Chlorite	Haematite	Quartz	Fluorite	Chalcedony Barytes, Dolomite, Calcite		
Ore minerals		Arsenopyrite Wolframite Cassiterite Molybdenite Specularite		Scheelite Stannite	Chalcopyrite Pyrite Sphalerite		Pitchblende Niccolite Smaltite Cobaltite Bismuthinite Argentite Galena		Tetrahedrite Bournonite Siderite Haematite Marcasite Jamesonite Stibnite
Economically important elements		As W Sn		Cu			U Ni Co Bi Zn Ag Pb	Fe Sb	
Usual occurrence	Faults, Veins, Stockworks	Lodes, Caunter lodes Faults, Veins, Stockworks, Carbonas, Pipes				Lodes, Caunter lodes, Cross-courses, Faults, Veins		Mainly Cross-courses Faults, Veins	
Wall-rock alteration		Greisening Tourmalinisation		Feldspathisation	Chloritisation Haematisation	Silicification			

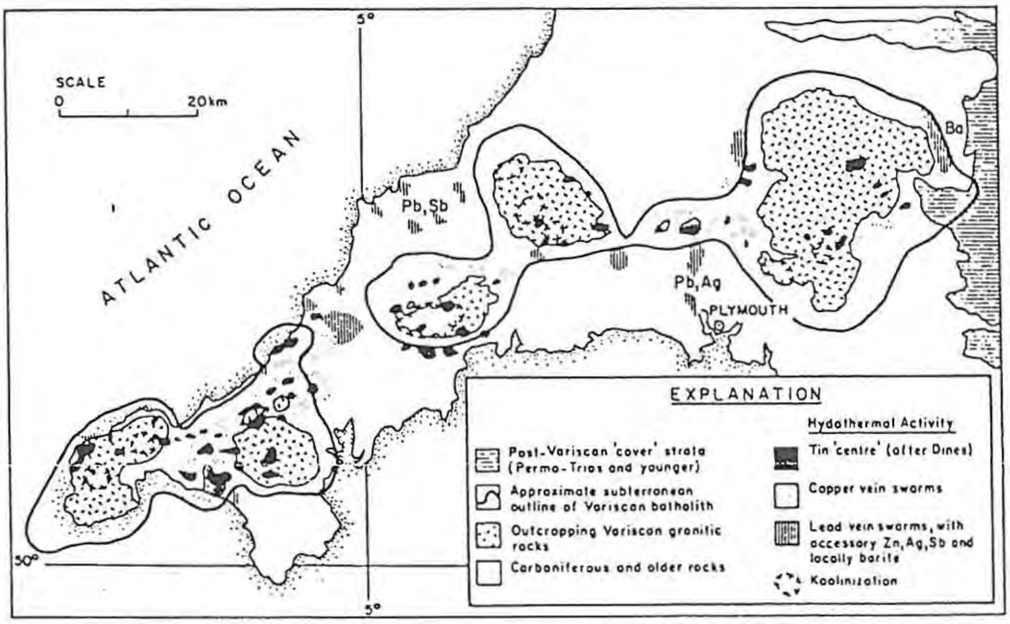


Figure 4.9: Map of the mineralised centres of SW England showing broad zonal relationships of tin, copper, and lead mineralisation. There is a clear indication of a spatial association of mineralisation with the roof zones of the Variscan granites (Moore, 1982)

4.3 Malay Peninsula

The Malay Peninsula is part of the large "tin girdle" of SE Asia, which forms an arc some 3400km long and up to 800km wide. The region accounts for about 45% of world Sn production, and including China about 30%W. Malaysia is primarily a Sn producer (1980 : 61,400t, 25% World), and although W is spatially related, deposits tend to be subeconomic. Malay Sn production is predominantly (95%) from placer deposits (Table 14). with the remainder from primary hard rock deposits, of interest here. There is only

Table 14: Malaysian Sn production (typical grades and tonnages)

	mt	Conc(t)	Sn recovered (t)	Grade recovered
Dredging	28,6	1500	1125	39ppm
Hydraulicking	1,0	220	160	151ppm
Opencast	5,0	27000	17300	±0.35%
Underground	0.16	2000	1500	±1%

one significant hard rock mine (Pahang Consolidated Mine in the Pahang-Trengannu field, Eastern Malaysia). The two other tinfields (Kinta Valley, Kuala Lumpur) are located in the western region of the Peninsula. Figures 4.10, 4.11 indicate the location of the major tinfields, tin and tungsten deposits.

4.2.1 Features of Sn-W mineralisation

A wide variety of primary deposits are known within the major tin-fields, but with few exceptions production has been insignificant (Figs. 4.10, 4.11, Table 14). Of importance are the different styles of mineralisation and metal associations in the Main Range and Eastern Sn-belt granitoids (Table 15), as well as a clear cut spatial association of tin mineralisation and ilmenite-series granitoids (Fig. 4.12, Ishihara et al., 1979).

The main source of detrital cassiterite in the Main Range Sn-belt comes from greisenised vein swarms in the outer zones of granite plutons. Most of these primary deposits (e.g. in the Kinta Valley) are thin stringers of cassiterite and cassiterite-quartz transgressing the granite, or small lodes of quartz-cassiterite, pyrite and arsenopyrite near the granite/

Table 15A: A comparison of tin mineralisation of the Main Range and Eastern Belts, Malay Peninsula (Hosking, 1977)

WEST BELT	EAST BELT
Sn pegmatites and aplites occur. Nb/Ta species may also be present.	No Sn pegmatites known.
Sn skarns with cassiterite and/or malayaite, and/or Sn-andradite not uncommon.	Sn/Fe skarns present. Malayaite absent.
Contact, stratabound, pipe-like and vein-like skarns known.	Massive contact and stratabound Sn skarns only recorded.
Many rich and poor stanniferous veins, lodes and replacements, that occur singly or in swarms, and that have a very modest dip length, are known. Commonly associated with granite cusps.	Sn veins and lodes, singly or in swarms, with limited dip length, distributed rather sparsely in the Belt.
No Sn lodes or veins with considerable dip length known, but some hydrothermal pipes in the marble may have considerable dip lengths.	Sn lodes, with considerable strike and dip extensions, and of major importance, occur at Pahang Consolidated Mine, and perhaps elsewhere in N.E. Malaya. Sn deposits, of similar dimensions, also occur at Kelapa Kampit, but these, in part, may be stratabound ones.
Sn xenothermal veins and pipes are not uncommon, particularly in the Perak and Selangor fields of Malaysia.	Only one Sn xenothermal deposit known and that is the Manson Lode, in Kelantan, Malaysia.
Cassiterites that display red to pale colour pleochroism are common.	Cassiterites generally display dark-brown to pale-brown pleochroism, or are non-pleochroic.
No wood tin recorded.	Wood tin recorded from 2 localities.
Stannite not uncommon.	Stannite only recorded from the Manson Lode and from Kelapa Kampit mine.
Sb and Be species recorded from a number of Sn deposits.	No Sb species recorded from Sn deposits. Phenakite recorded from Kelapa Kampit mine.

B) Comparison of granitoid chemistry and tin mineralisation (Hutchison, 1979; Ishihara et al., 1979)

THE MAGNETITE-SERIES AND ILMENITE-SERIES GRANITOIDS OF THE MALAY PENINSULA REGION BY NUMBER OF MEASUREMENT EXCLUDING THOSE DETERMINED BY BULK CHEMICAL ANALYSES

Area and age (number of analysis)	Magnetite-series	Ilmenite-series	Tin mineralization
Southern Thailand			
Northwest of Krabi district, Cretaceous (n = 22)	8	14	Intense
Main part, Carbo-Jurassic (n = 48)	0	48	Moderate
Whole area (n = 70)	8(11%)	62(89%)	
Peninsular Malaysia			
Southern part, Cretaceous (n = 12)	11	1	None
Central belt, Permo-Triassic (n = 9)	9	0	Rare
Eastern belt, Permo-Triassic (n = 28)	6	22	Moderate
Main Range belt, Permo-Triassic (n = 45)	4	41	Intense
Eastern and Main Range belts	10(14%)	63(86%)	

SOME CHEMICAL CHARACTERISTICS OF PENINSULAR MALAYSIAN GRANITES

Granite	Crystallization Index Range	K_2O wt % range and mode ( )	K_2O/Na_2O range and mode ( )
Late Cretaceous	0 to 6	2.9 to 5.2 (4.3)	1.3 to 2.8 (1.5)
Main Range	0 to 14	1.4 to 9.0 (5.0)	1.0 to 5.2 (1.7)
Benom (Central)	0 to 14	4.3 to 6.2 (4.7)	1.0 to 2.4 (1.4)
Eastern Belt	0 to 25	2.4 to 5.4 (4.1)	0.5 to 2.0 (1.3)

(Data are lacking from granites of the Western Stable Shelf.)

TRACE ELEMENT CONTENTS IN PPM OF PENINSULAR MALAYSIAN GRANITES. AVERAGE VALUES WITH RANGE IN BRACKETS

Element	Late Cretaceous	Main Range	Benom	Eastern Belt
Rb	217 (146–289)	531 (334–738)	323	243 (186–348)
Sr	200 (193–207)	53 (4–164)	186	91 (19–158)
Zr	131 (128–135)	116 (60–243)	16	120 (107–142)
Ba	816 (808–825)	374 (6–877)	694	879 (253–1441)
Sn	5 (5–6)	7 (5–11)	6	5 (4–6)
Nb	5 (5–6)	8 (7–9)	7	6 (3–9)
W	1	4 (1–7)	6	2 (1–3)

Two analyses of the Langkawi granite (Bignell, 1972) give Rb 348 and 562 and Sr 40 and 27 ppm.

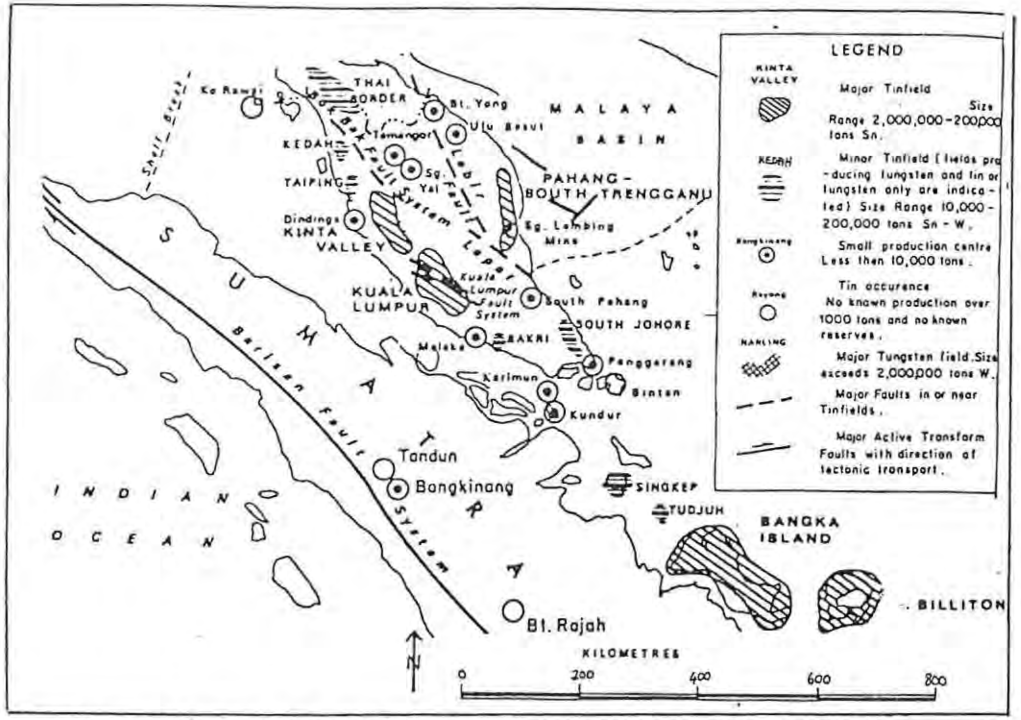


Figure 4.10: Locality map of the most important tin deposits of the Malay Peninsula (adapted from Hutchison and Taylor, 1978)

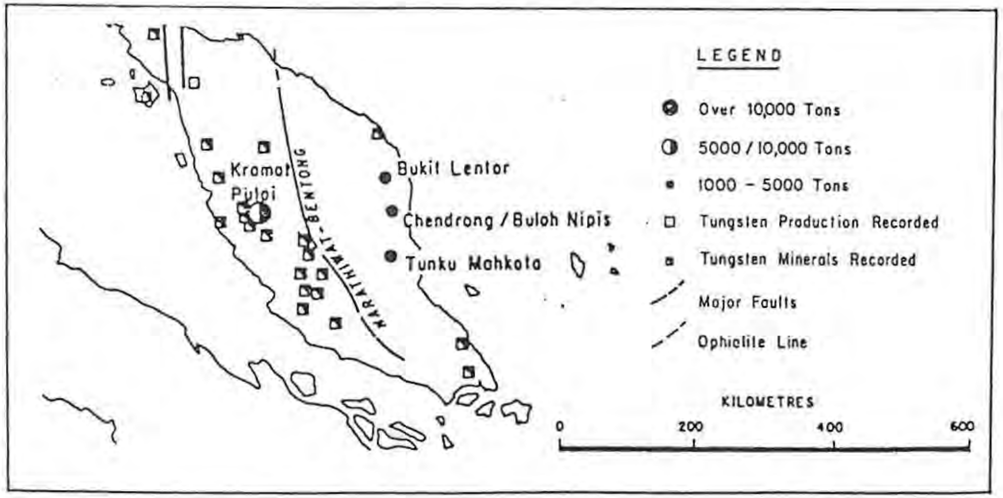


Figure 4.11: Locality map of tungsten deposits of the Malay Peninsula (adapted from Hutchison and Taylor, 1978)

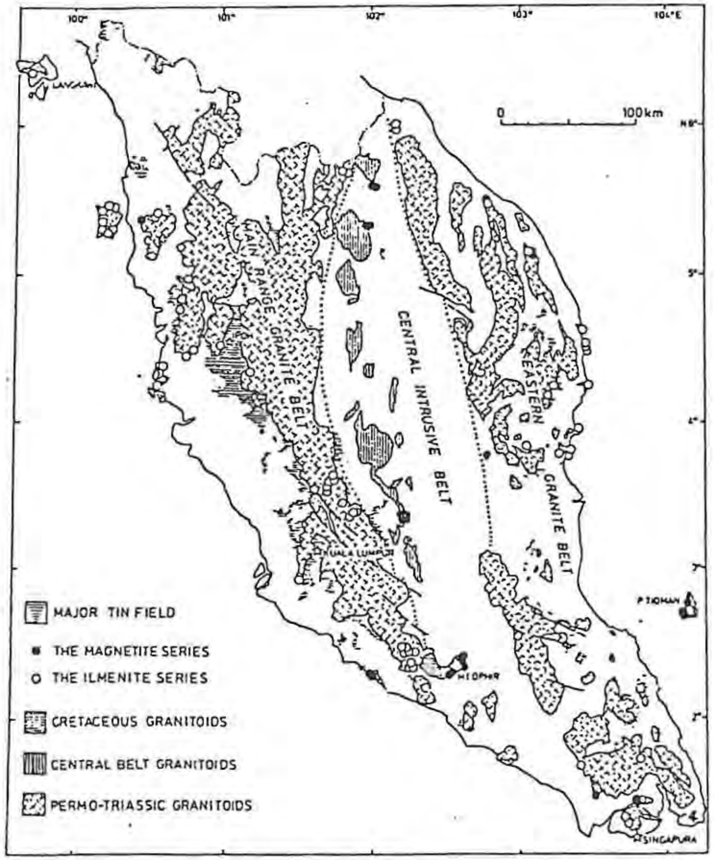


Figure 4.12: Distribution of magnetite-series and ilmenite-series granitoids of Peninsula Malaysia in relation to the major tin-fields (Ishihara et al., 1979)

country rock contact (Hosking, 1970, 1975; Hutchison, 1983). Other areas have mineralised veining hosted by schistose rocks, Sn/W skarns that developed within calcareous host sediments, and isolated Sn/W breccia pipes (hydrothermal breccia pipes with mineralised matrix) and carbonas (irregular replacement ore bodies). Minor pegmatites hosting cassiterite and columbite also occur in the Kedah field (see Fig. 4.10).

The Eastern belt hosts quartz vein and skarn-related Sn mineralisation, but notable no pegmatites. Generally mineral occurrences are associated with well developed andalusite-cordierite bearing thermal aureoles. At the Pahang Consolidated Mine Carboniferous marine sediments are intruded by Permian granites. Related sheeted vein systems localised along fault planes host cassiterite and sulphides, but although these veins and faults extend down to the host granite, mineralisation ceases around 350m from the granite. There is a clear distinction between this style and that of the Main Range, where lodes occur within the contact zone of the granite cupolas (i.e. largely endogranitic). Skarns are also developed where these granites intrude calcareous sediments, and are marked by iron-ore bodies with disseminated cassiterite close to the granite intrusion. Iron is considered to have either been derived from the sediments or from an intermediate composition (dioritic) magnetite-series granitoid. Tin is envisaged to have been emplaced in association with a later ilmenite-series granitoid, which also generated sufficient heat to sustain a hydrothermal system (Hutchison, 1983).

In the Malay Peninsula W is generally subordinate to Sn, and although the association is widespread, the closeness on a local scale varies widely. Tin, as cassiterite, is principally found in greisenised cupolas of granites and associated vein swarms extending about 100m down dip, as well as in skarn deposits. Tungsten appears to exist mainly as scheelite in contact metasomatic environments. Hosking (1973, 1975) has generally categorised the SE Asian tin deposits (Fig. 4.13), as well as, in more detail, those of the Main Range (Fig. 4.14). However considerable detailed studies of the often poorly exposed deposits are still required to elucidate their nature and possible economic potential.

4.3 Southern Africa

South Africa does not rate as a major tin or tungsten producer on a global scale, although it caters for most of its local tin requirements.

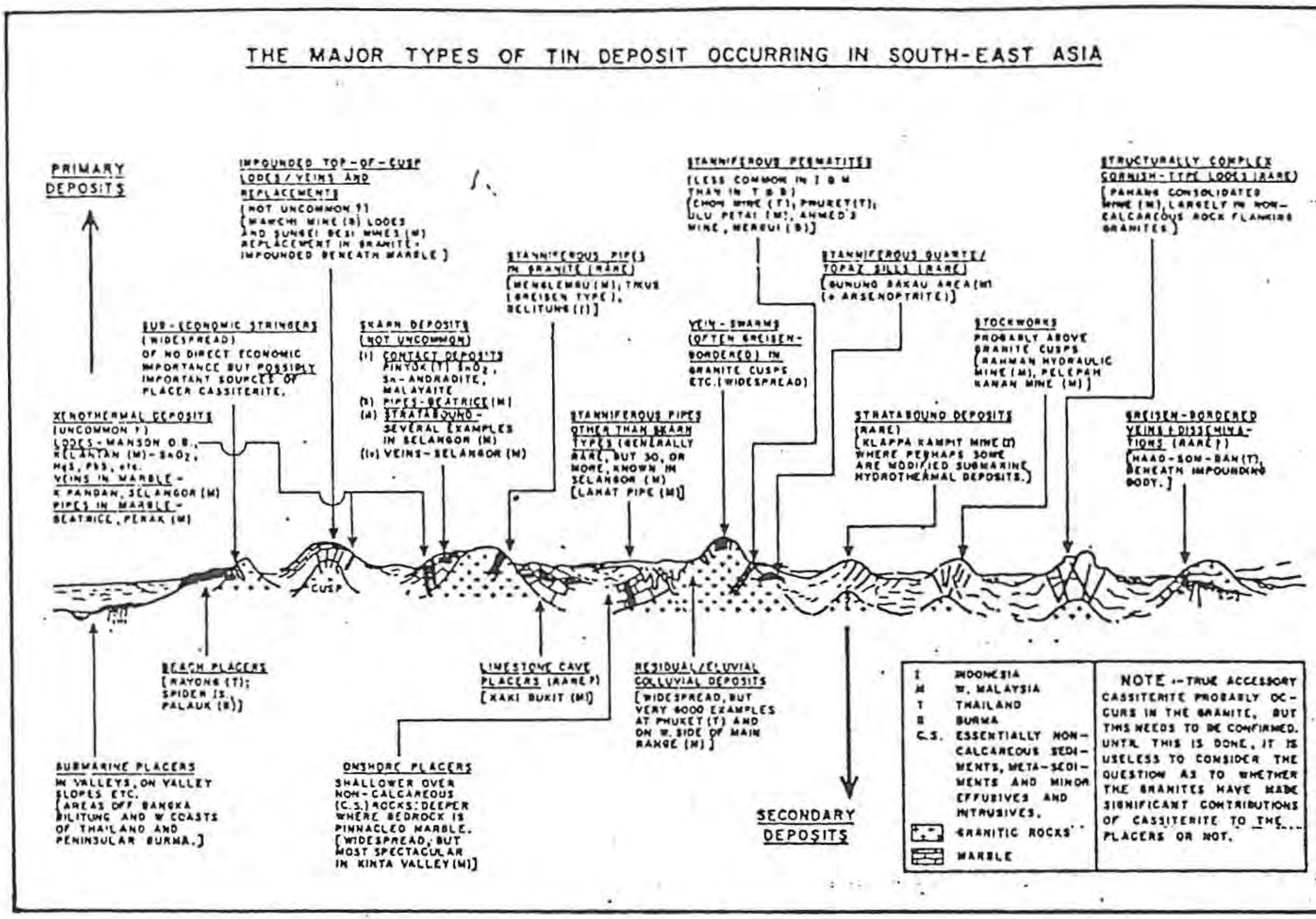
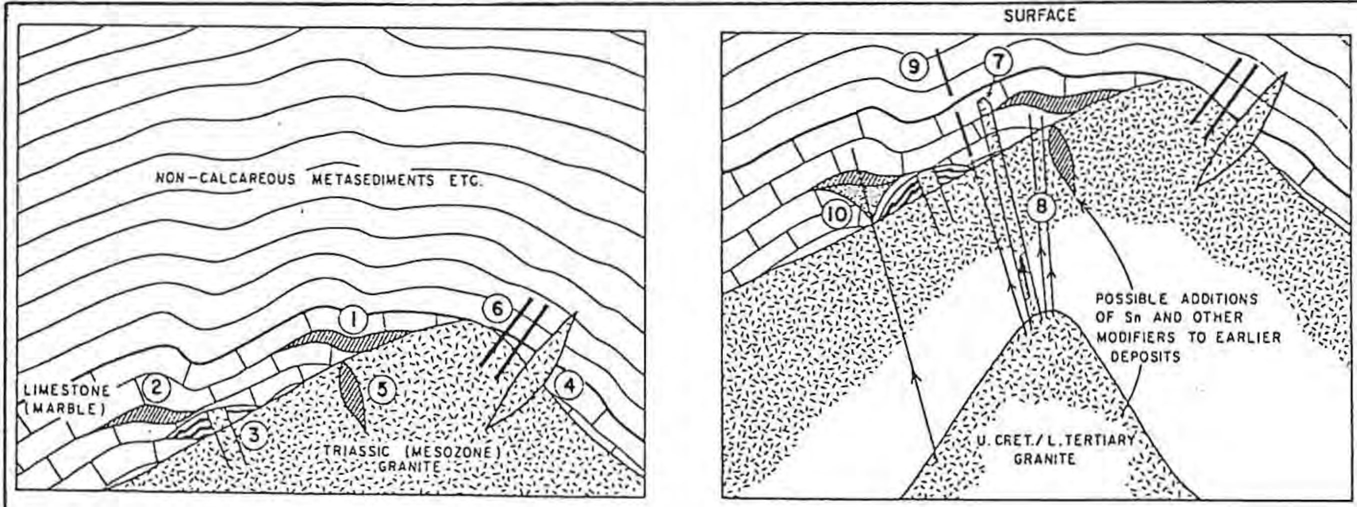


Figure 4.13: Major types of tin deposit occurring in SE Asia. Malay examples are highlighted (Hosking, 1977)



The postulated major results, in West Malaysia, of superimposing mineralisation due to a high-level granitic invasion on that due to granite emplaced in the mesozone.

1) Strataform skarn.
2) Folded bed of non-calcareous sediment or metasediment.
3) Pegmatite, in part in a skarn host.
4) Pegmatite in granite/non-calcareous host rocks.
5) Tin orebody impounded beneath limestone.
6) Tin-bearing lodes/veins in granite/non-calcareous host rocks.
7) Pegmatite in granite/limestone host rocks.
8) Tin-bearing lodes/veins in granite, etc.
9) Xenothermal tin-bearing deposit in limestone. Possibly associated with vein- or pipe-like skarns.
10) Metasomatic tungsten-bearing deposit impounded in limestone by a metasedimentary folded bed.

Figure 4.14: A categorisation of tin deposits in the Main Range Belt, involving superimposed mineralisation due to high level granitic invasion on that related to earlier plutonism (Hosking, 1977)

The Republic is badly lacking tungsten resources. Numerous tin deposits in South Africa and Namibia have been or are currently mined. Present day production stems largely from the Potgietesrus tinfields, in association with acid phase rocks of the Bushveld Complex (see 4.3.3). In South Africa other occurrences are known near Upington in the high grade rocks of the Namaqua Metamorphic Complex (e.g. Van Roois Vlei), from pegmatite to the west within the same Complex, in association with tungsten in stratabound occurrences in the O'kiep area (see below), and as minor deposits associated with the Cape granites (see 4.3.2).

Van Rooi's Vlei Sn-W vein deposit is located within the gneissic Kakamas Terrane of the Gordonia Subprovince (formerly Namaqua 'front', Stowe, 1983). The deposit consists of several en echelon quartz-tourmaline vein zones intruded into a tight syncline of medium to high grade metamorphosd Proterozoic supracrustal rocks. Recent exploration (Shell Metals) has proved a small ore body of rough dimensions 300m strike, 500m deep, and 5-20m wide, with a resource of 2.5mt grading 0.38% WO_3 and 0.21% Sn in the major vein zone. The mine has been technically approved, and only awaits improved tungsten prices before development will commence. Of note is that considerations to ore genesis show affinities to exogreisen mineralisation, possibly related to S-type ilmenite series granites (Bentley, 1984).

Elsewhere in South Africa tungsten has been mined sporadically from small scheelite deposits or else in association with cassiterite mineralisation, and even then conditional to market conditions. Most production appears to have come from stratabound deposits in the O'kiep District of Namaqualand, within the so called Wolfram Schist unit 'interlayered' with the Nababeep and Concordia granite gneisses. Grades were quite high (0.29-2.50% WO_3), and around 2900t of concentrate were produced from 1941-1956 (Keyser, 1976).

Other small occurrences occur to the north within the Namaqua Pegmatite belt, and in association with shear zones transecting large rafted amphibolitic xenoliths within Vioolsdrif Suite granodiorites of the Richtersveld Province.

Present day Sn production in Namibia is from the Uis pegmatite deposit. Sn and W production has been recorded from the now closed Brandberg West and Kranzberg Mines. These deposits and similar occurrences are discussed in more detail below (4.3.1). This section overall is confined to discussions of the deposits/occurrences related to the Damara (Namibia) and Saldanian (South Africa) Provinces, with a brief overview of the Bushveld deposits as they presently provide the major tin resource of South Africa, and are considerably better studied. Fuller reviews on tin mineralisation in South Africa are given by Crocker et al. (1976), Crocker and Callaghan (1979) and on tungsten by Keyser (1976).

4.3.1 Damara Province

Tin and tungsten occurrences in the Damara Province are very poorly documented, and the following relies mainly on personal observations, coupled with descriptions of the Uis Mine by Richards (1980) and Kranzberg by Grant (1980).

The Damara "tin province" is located within the Northern and northern Central zones of the Province, within which four "belts" of mineralisation can be delineated (Fig. 4.15). These, progressing NW, are the Karibib Belt (pegmatites hosting Li, Be, Cs, Ta-Nb mineralisation), the Khan-Rössing Belt (mostly uraniferous alaskites, but extending NE to encompass the Erongo Complex and related Sn-W greisen, quartz vein and pegmatite mineralisation), the Uis Belt (Sn-Ta in pegmatites), and the Brandberg Belt (Sn-W in greisens, quartz veins, pegmatites, skarns and replacement bodies). Some metal zonation is evident within these belts. For instance the Uis pegmatite belt is Ta-rich at the SW margin and Sn-rich towards the NE extremity. The Brandberg Belt, which is some 75-80km long by 20km wide, is essentially W-rich, although Sn is present throughout, and of significance in the SW. The Belt is defined by conspicuous ring structures (Fig. 4.15).

An important feature is the distribution of late- to post-Karoo alkaline complexes, which are aligned along NE trending lineaments (Fig. 4.15). Of note are the Messum, Brandberg and Erongo complexes, the latter having associated Sn-W mineralisation of which the Kranzberg W deposit (see below) and the Spitzkoppe are the most economically significant.

Pegmatites (as at Uis Mine) are both syn- and late- to post-tectonic, as evidenced by deformed pegmatites parallel to the regional foliation, and others that are cross-cutting. These pegmatites are assigned Damara ages (560-500Ma). To the north of Brandberg the Ousis pegmatites are

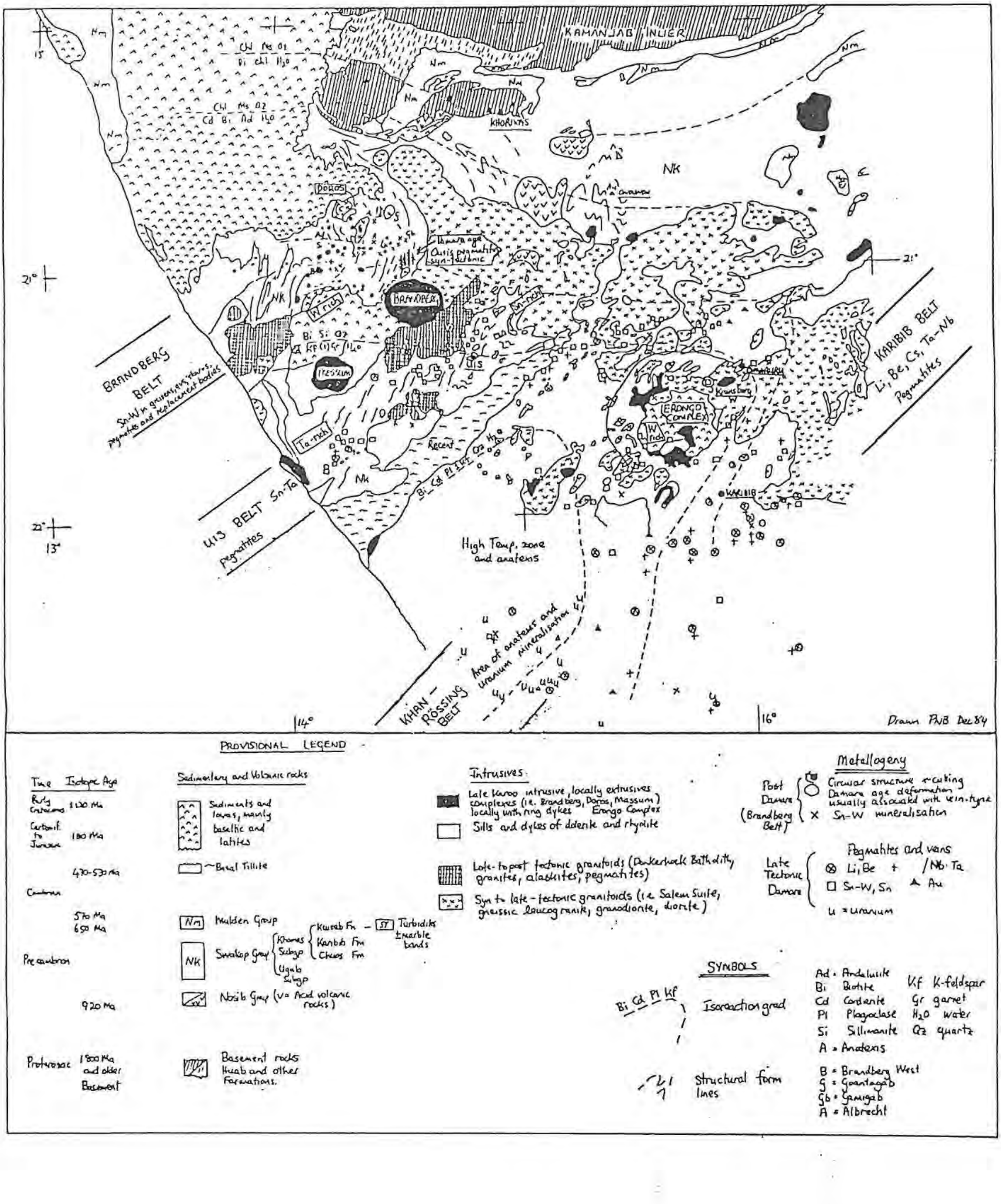


Figure 4.15: Distribution of Sn and W mineralisation in the northern central and northern zones, Damara Province (after Pirajno, unpublished data)

syn-tectonic, and are correlable with those at Uis. It is conjectural at present as to whether the diverse Brandberg Belt mineralisation is solely related to late-tectonic Damara granitoid emplacement, or else is 'remobilised' Damara (Uis-type) mineralisation, emplaced during Karoo magmatism (F. Pirajno pers. comm., 1984). The W-rich nature of the vein systems may suggest affinities to the Erongo Complex, and an overall link to the alkaline complexes may be postulated.

Kranzberg Mine (presently non-operational)

The Kranzberg Mine is located 34km south of Omaruru township. Tungsten mineralisation (largely as wolframite, grading 0.3-0.5% WO_3) is associated with post-Karoo granites of the Erongo Complex (Fig. 4.16), which appear to have been emplaced preferentially along the NE-SW trending Waterberg lineament. Other rock types in the area include Damara granites, comprising porphyritic and equigranular varieties of the Salem granite, and which have also intruded the Khomas schists along a NE-SW trend (Grant, 1980).

Three types of mineralisation are discernable in the area:-

i) Kopje mineralisation - greisen bodies formed at the contact zone of the Khomas schists and Kopje granite (a foliated leucocratic q-fs-bi granite of the Salem suite). The zones of greisenisation and disseminated? mineralisation are localised along vertical to steep dipping fractures.

ii) C-zone mineralisation - W associated with the greisenised contact zone between Salem granites and Khomas schist, dipping 55°SE. Greisenisation and attendant mineralisation is restricted to often-flat fractures within the granite (i.e. endogreisen). The greisens consist of a qtz-mica-fluorite-topaz assemblage ± chalcopyrite, arsenopyrite, molydbenite, scheelite, pyrrhotite and bismuth minerals, and grade 0.3-0.5% WO_3 (principally from wolframite?, Grant, 1980). Mineralisation appears largely restricted to the roof zone of the granite. Hanging wall Khomas schists are extensively tourmalinised, as are the overlying Karoo sediments, significantly inferring that mineralisation may be of post-Karoo age.

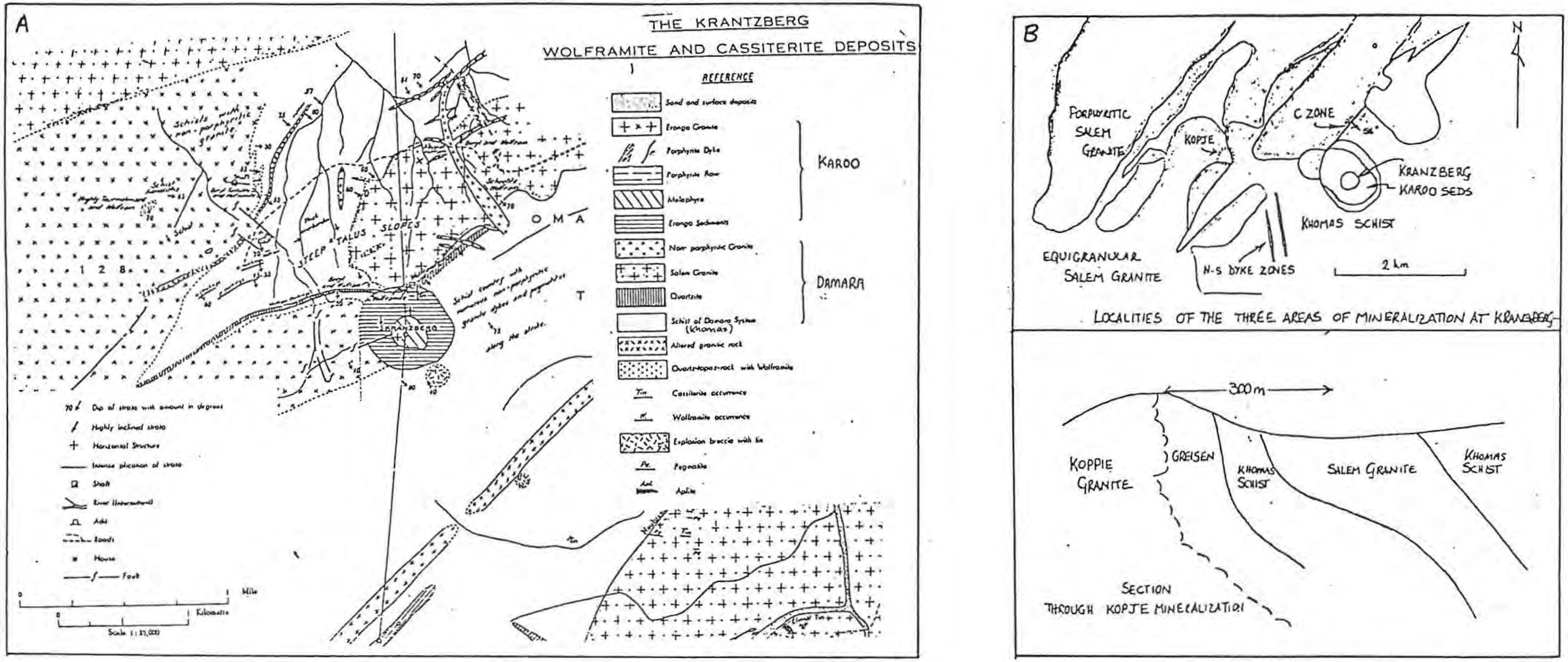


Figure 4.16A: Geological map of the Kranzberg wolframite deposit (anon, in Grant, 1980)

B: Localities of three mineralised areas and section through Kopje mineralisation, Kranzberg mine (Grant, 1980)

iii) N-S vein mineralisation - weakly mineralised zones 3-10m thick and of some 500m strike length occur within the Khomas qtz-fs-bi schists. Mineralisation in pegmatites, quartz and granitic (greisen?) veins, and aplite has associated tourmaline and silica alteration of the schists. Minor wolframite ± pyrite occur on fracture planes that pre-date the silicification.

The present writer considers the evidence (i.e. alteration of Karoo sediments, as cited by Grant, 1980) for post-Karoo age mineralisation as equivocal. There is no doubt that remobilisation and/or redeposition of pre-existing mineralisation may occur, coupled with emanation of fluids capable of effecting tourmalinisation and greisenisation of country rocks. Only more detailed geochemical and petrographic studies will elucidate whether or not there has been any significant replacement of Damara mineralisation (i.e. the spatial relationship of mineralisation to Kopje granite/Khomas schist contacts infers a Damara age) or, more unlikely, whether Karoo-aged mineralisation has preferentially localised along structures at the granite/schist contact.

Uis Pegmatite Mine

The Uis Mine is currently the largest operating mine in the world extracting Sn from pegmatites. Reserves amount to about 250mt @ 0.2%SnO_2, and up to 1980 22702t of Sn concentrate had been produced. The Uis Mine is located at the NE extremity of a 125km long by 30km wide pegmatite belt (Fig. 4.15), which is interpreted as a NE trending graben (Richards, 1980). Satellite imagery, aerial photographs and aeromagnetic surveys have all delineated structures substantiating this interpretation. Furthermore the belt, analagous to other areas of the central and northern zones, has been the locale of transform faulting, and repeated reactivation and tectonism along long-lived lineaments, with localised emplacement of Karoo alkaline complexes and dolerite dykes.

The pegmatites are hosted by Khomas Subgroup schists, comprising a highly variable sequence of metapelites, metagreywackes, calc-silicates and quartzites. These attain upper amphibolite facies metamorphic grade, and there is evidence for fairly widespread regional contact metamorphism from the spotted nature of the schists and andalusite-cordierite (retrograded to pinite) assemblages. Locally, in the vicinity of pegmatites, the schists are heavily tourmalinised (to 60 modal %; Richards, 1980).

Syn- to post-tectonic Salem granites in the Uis area include various phases of foliated porphyritic muscovite-, porphyritic biotite-granite $\pm$ tourmaline, leucogranite, and coarse grained, less foliate, porphyritic biotite granite. Pegmatites are especially abundant at the foliate granite/schist contacts.

At Uis Mine there is a swarm of over 100 individual pegmatite bodies, all 'layered', unzoned, and coarse to very coarse grained. Mining operations are centred upon eight of these pegmatites, which are up to 100m wide and 1000m long. The mineralised pegmatites are syn-tectonic, they vary in size, and are emplaced within tension gashes, with noticeable pinch-and-swell structures. Average grades of Sn are around 0.15%, but over 1m sections may range from 0.002-3.20% Sn. Cassiterite distribution is as random fine-grained disseminations throughout the pegmatite. There is a close association of cassiterite with albite, muscovite, sericite and schlorite, and grains often tend to concentrate along fractures in quartz and orthoclase (Richards, 1980). Highest concentrations of Sn are found in association with sporadic greisenisation within a given pegmatite. Two generations of unmineralised quartz veins (pre- and post-pegmatite) are also recognised.

The pegmatites all have a characteristic alteration halo which persists for up to 200m from the body, and is marked by an inner 20-50m zone of intense greisenisation (quartz-muscovite-albite-tourmaline). This is further evidenced by geochemical enrichment in the wall rocks of Al, K, Fe, Ti, P, Li, F, B, Sn, Ta and Nb, and depletion in Si, Mg, Ca, Na. Similar geochemical trends exist in exogreisen alteration assemblages in New Zealand (Pirajno and Bentley, in press). Mineralogical changes such as feldspar to sericite, biotite to chlorite and tourmaline, and plagioclase to albite are discernable in thin section (Richards, 1980). Two discrete phases of biotite and tourmaline were identified, and could be interpreted in terms of two pulses of contact metamorphism and boron metasomatism, or else evidence of magmatic intrusion and alteration in Damara <u>and</u> Karoo times. The former appears more likely, with mineralisation and, alteration effects superposed on earlier hornfelsing due to emplacement of syn-tectonic Salem-type granite. Stages in the evolution of the Uis pegmatites can be envisaged as follows:

1)	Generation of crustal-derived syntectonic Salem-type granite.		
2)	Intrusion of parent body at depth, spotting of country rocks (contact metamorphic effects)	differentiation in situ	a. muscovite-garnet granite + b. pegmatites (coexisting exsolved aqueous fluids) $\pm$ aplites Late tectonic emplacement along the margin of the

3) Deformation and metamorphism continues with - growth of cordierite (biotite) within foliation - folding and development of foliation	c. Limited concomitant to late greisenisation along granite margins and associated with pegmatites (metasomatic activity)	Salem-granite and migration of pegmatitic fluids along fractures emanating from intrusive margins

Brandberg West Sn/W Mine

The Brandberg Mine (presently closed) is located on the margin of a ring structure where it intersects a major NW trending fault (Fig. 4.17). Mineralisation is largely within quartz veins hosted by Khomas Sub-group quartz-biotite schists, which display biotite-cordierite hornfels. The quartz veins appear to have a sheeted arrangement (Plate 4A), localised along fractures related to the diapiric intrusion of granite into the core of a SW plunging anticline, and which subsequently collapsed and formed the ring structure. The principal strikes are E-W and NE-SW. Recent diamond drilling has elucidated the existence of an intrusion (greisenised granite) below the open pit.

The vein system overall extends 500-600m to the NE, covering an area some 1200m long by 500m wide, although in the NE block mineralised veins are more randomly orientated. The veins within the open pit are mostly localised within the lower marble and upper schist (Fig. 4.18) and have a classic greisenised halo with muscovite and tourmalinised selvages which overprint biotite hornfels (Plate 4B, Fig. 4.19). The principal ore minerals are cassiterite, wolframite, scheelite, plus sulphides such as chalcopyrite, bornite, pyrite, pyrrhotite, marcasite, 2° chalcocite, Pb-Bi sulphosalts. Tourmaline and fluorite are very common. Wolframite (up to 0.10m) crystals occur mostly on vein margins, with cassiterite in the central parts of veins and within wallrocks in association with tourmaline. The veins range from 0.1-1.5m in size and have typical grades of around 0.1% SnO_2 and 0.2% WO_3, including wallrock. Apparently in 1980, prior to closure, reserves were estimated at 4mt grading 0.25% $WO_3 + SnO_2$.

Within the open pit, apart from the attendant greisenisation around given veins, a zone of sulphide mineralisation and associated argillic

Figure 4.17: Simplified geological map of the Brandberg North area, showing mineral occurrences, ring structures and prospect localities (Pirajno, unpublished data)

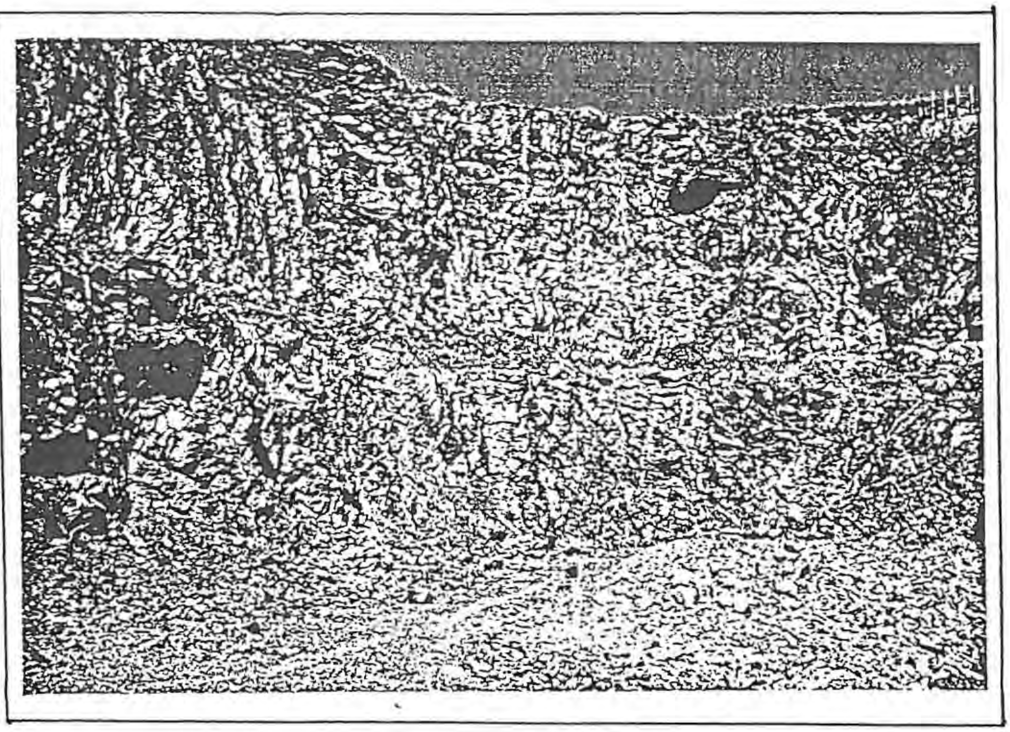


Plate 4A: View of the NE side of the Brandberg West open pit. Note the anastomosing sheeted vein system

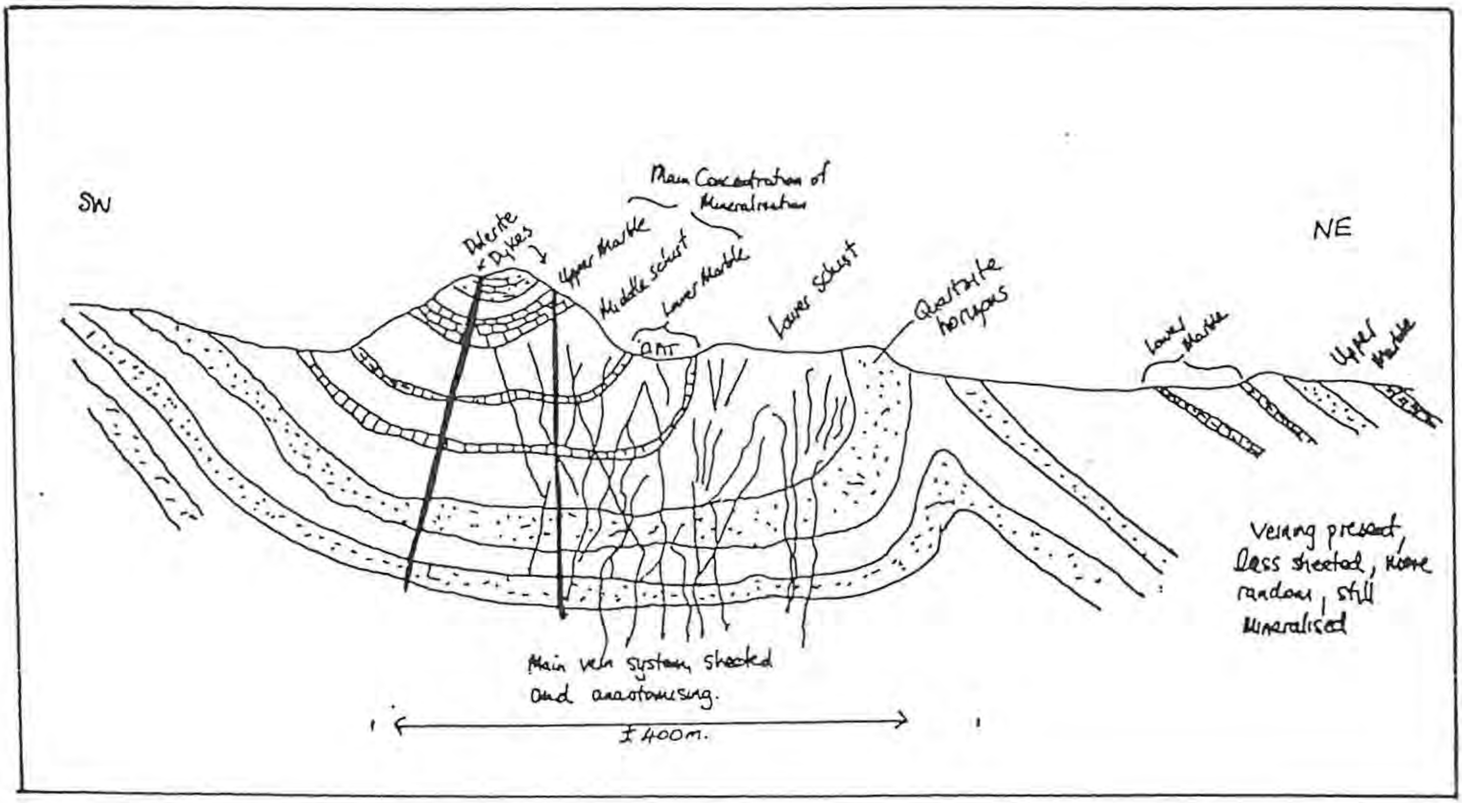


Figure 4.18: Geological sketch cross section, Brandberg West open pit

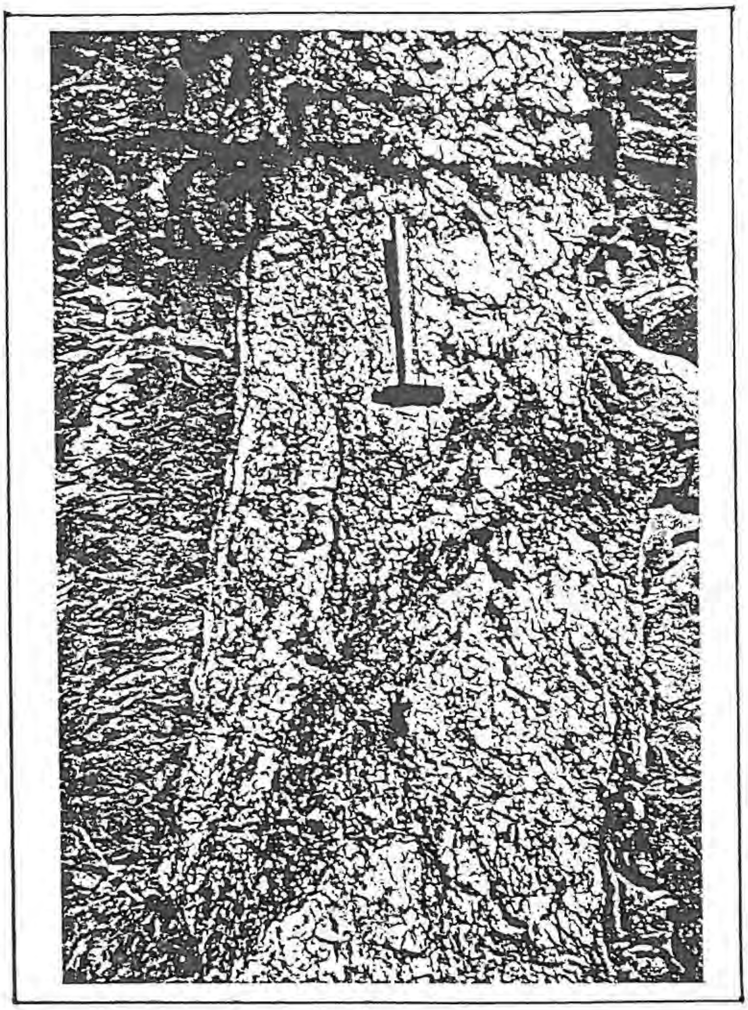


Plate 4B: Typical quartz vein, Brandberg West open pit. Note the distinctive muscovite selvage, and tourmalinised wall-rock. The veins carry cassiterite, wolframite, scheelite, plus minor sulphides (Fe, Cu, Pb, Bi)

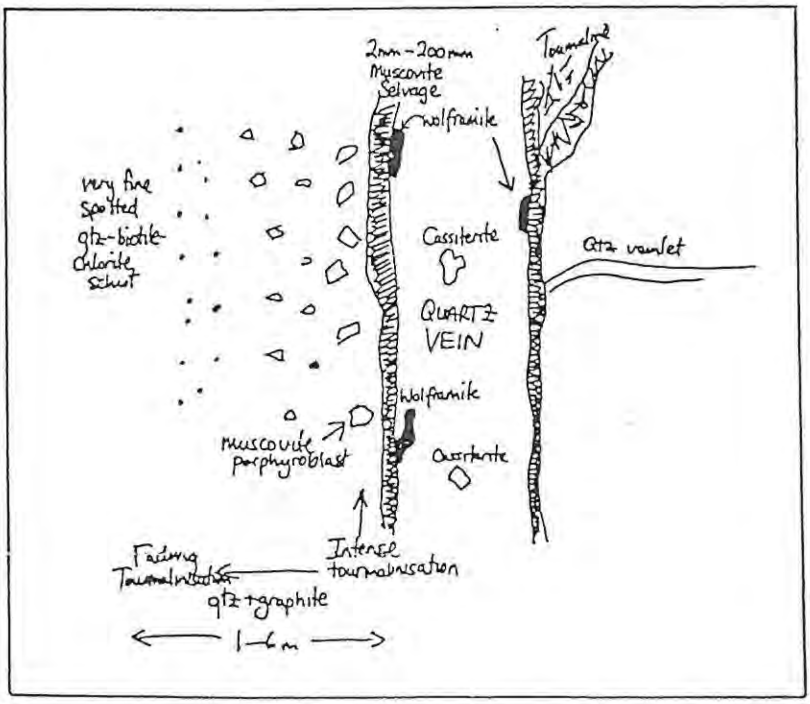


Figure 4.19: Schematic illustration of typical wall rock alteration, Brandberg West open pit

alteration occurs on the northern side, where a post-mineralisation dolerite dyke (+Cu mineralisation) has been emplaced along similar trending fractures (E-W) that host quartz veining. The fractures that have channelled the dykes have also locally displaced the mineralised veins.

Brandberg North area

The simplified geology of the area north, and north-west of the Brandberg is shown in Fig. 4.17. The area is underlain by turbiditic sediments and intercalated marbles of the Khomas Subgroup. Basal Chuos Formation mixtite is locally preserved, and there are sparse intercalated felsic volcanics. The predominant lithologies are pelitic, psammitic and calcareous rocks of the Kuiseb Formation. Two persistent marble members 10-60 and 50-170m thick make a subdivision of the succession possible although in the Brandberg North area it is often difficult to distinguish the marble bands, making regional correlation difficult.

The sedimentary succession is intruded by various granitic rocks, which occupy vast areas to the south and east. These comprise syn- to post-tectonic Damara granites (Salem-, Sorris Sorris-types) and late- to post-Karoo granites, and which exhibit partly concordant, partly discordant relationships with the country rock. To the northwest widespread Karoo-age (120-130Ma) lavas (the Etendeka Plateau) include basaltic to dacitic flows. The Doros Complex in the north is a conspicuous feature, consisting of an alkaline complex of mostly layered olivine-rich melanogabbros emplaced during late Karoo times. A recently discovered layered mafic-complex, emplaced concordantly within the Khomas Subgroup outcrops NW of Goantagab (Fig. 4.17).

Metamorphism in the region is low grade, attaining a maximum of lower greenschist facies. The effects of thermal hornfelsing by post-tectonic granites are clearly superimposed on the metamorphic fabric, being manifested as spotted schists (biotite-cordierite porphyroblasts). Notably in the Brandberg North area the porphyroblasts cut the foliation, whereas at Uis (to the southeast of Brandberg, Fig. 4.15) they are larger and are parallel to the foliation, inferring syn-tectonic formation, and continued growth during progressive deformation. In more aluminous zones of schist andalusite may form (c.f. cordierite in Mg-rich zones). The other important feature is that the thermal aureoles related to the intrusions of granite invariably are localised within clearly defined circular structures (visible on Landsat imagery and aerial photographs) that cross-cut schistosity, and also occur within Salem granites. These structures are formed by collapse from polyphase diapiric intrusion, and often have directly associated veining and alteration (greisenisation) $\pm$ mineralisation. As such they present a prime exploration target for subsurface mineralised bodies.

Four periods of deformation are recognisable in the Brandberg North area (Fig. 4.20). F_1 is equated with open to isoclinal folding, with a prominent slaty cleavage (S_1). This was followed by open to tight folding with localised flexural slip (F_2), gently warping (F_3) and then kinking (F_4). The latter actually may be flexuring due to northwest-southwest compression (Weber and Ahrendt, 1983), resulting in a steep NW dipping crenulation cleavage. The deformational phases are attributable to the Damara orogenic episode.

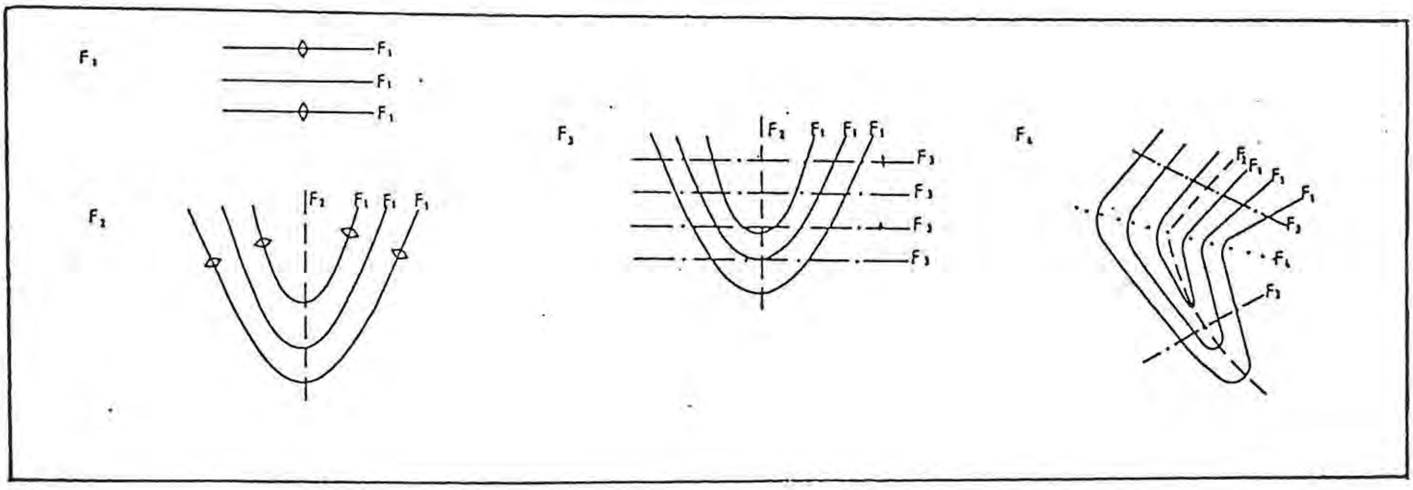


Figure 4.20: Sketch of deformational events, Brandberg North

A variety of Sn/W occurrences are observable in the Brandberg North area, including vein deposits intimately related to ring structures and greisenisation, pegmatites associated with the Ousis granite (see below), and replacement-skarn type bodies.

The Ousis pegmatites are emplaced adjacent to the Damara-aged, syntectonic Ousis granite (Fig. 4.17, Plate 4C), which is a highly differentiated biotite-muscovite granite. The pegmatites locally host Sn/Ta mineralisation (e.g. De Rust Mine), and are considered genetically related to the granite, which significantly displays geochemical specialisation, having very high Sn content (up to 220ppm), W(to 55ppm), high Rb/Sr ratios (around 5.0), K_2O (3.6-4.0wt%), Na_2O (3.7-4.0wt%) and SiO_2 (68-70wt%). Generally the pegmatites display foliation and a crenulation subparallel to the enclosing schists. The margins of the pegmatites may be tourmalinised and have a distinctive micaceous selvage (Plate 4D), probably related to late-stage greisenisation.

The contact of the Ousis granite and Kuiseb schist is marked by garnet-muscovite-tourmaline assemblages in the granite, and the emplacement of syntectonic, locally mineralised, greisen and pegmatite veins subparallel to and cross-cutting the foliation of the country rocks (Plates E, F).

Various forms of exogreisen vein hosted mineralisation are found within the Kuiseb schists. At Gamigab prospect cassiterite-hematite mineralisation is hosted by E-W trending veins, cross-cutting the NE-trending schistosity. The veins appear emplaced within an antiform, and importantly are restricted to within the schist, with adjacent marble units unmineralised (in contrast to the replacement-skarn type of mineralisation, see below). However the marbles, when proximal to the veining, display distinctive sideritic alteration (Plate 4G).

The schist is visibly thermally hornfelsed, with biotite clots that cross-cut the foliation. Of note is the intense hematite alteration adjacent to quartz veining, and hematite, calcite $\pm$ cassiterite infilling of fractures. This may indicate a (conjectural) second phase of mineralisation related to late stage fluids derived from Karoo intrusives (e.g. at Gamigab there is a Karoo aged volcanic plug nearby, with abundant Fe alteration).

Often exogreisen vein swarms are localised within ring structures. The veins are generally hosted by Khomas schists, and follow pre-existing joint or fracture patterns. Ore minerals include cassiterite, wolframite and sulphides such as chalcopyrite and pyrite. The veins are often intensely brecciated, indicating some tectonic reactivation and fluid remobilisation. Alteration around the veins is mostly muscovite/sericite-tourmaline-clay, whilst proximal marble units may be sideritised.

Replacement-skarn type mineralisation is found adjacent to and within the marbles of the Kuiseb Formation, manifested mostly as Fe-rich pods bearing cassiterite. In the Goantagab area (Fig. 4.17) these pods overly

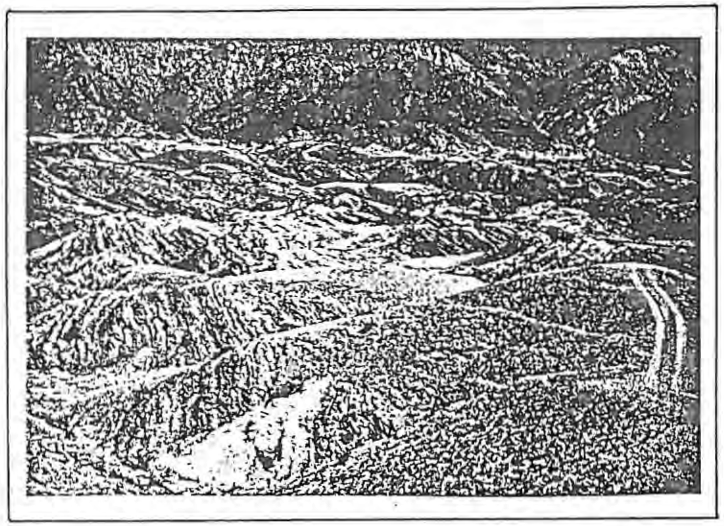


Plate 4C: Ousis pegmatite within Kuiseb quartz-biotite schists. The Brandberg is in the background

Plate 4D: Ousis pegmatite showing greisenised selvages (muscovite-tourmaline). The top of the pencil magnet indicates a grain of cassiterite.

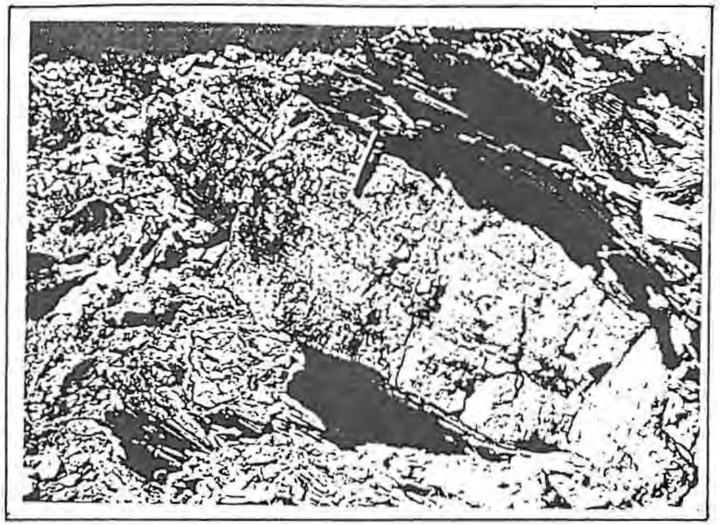


Plate 4E: Ousis granite-metasediment contact, with greisen vein parallel to foliation

Plate 4F: Ousis granite-metasediment contact. Greisen vein intruding biotite schist and squeezed along foliation

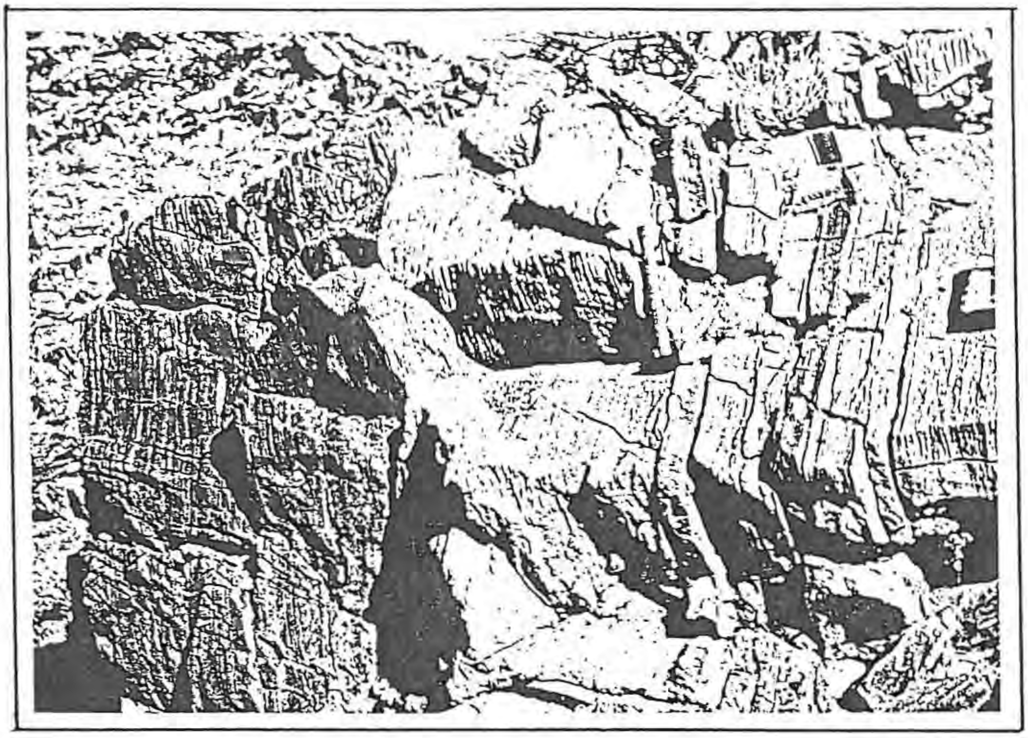


Plate 4G: Distinctive sideritic alteration of marble, adjacent to mineralised veining in schists (Gamigab prospect)

cassiterite- and sulphide-bearing vein systems within biotite schists flanking an antiform (Fig. 4.21, Plate 4H).

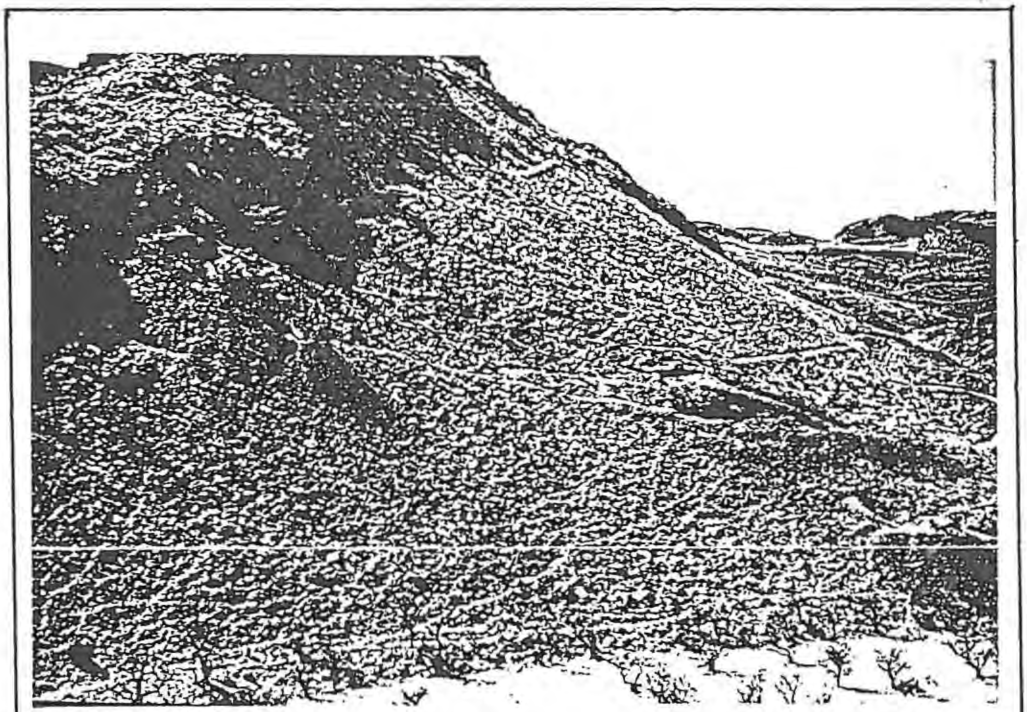


Plate 4H: View SE showing locale of Sn-hm pods which replace the upper marble horizon (forms bluffs top left). The scree slope (schists) has visible quartz veining carrying cassiterite and sulphides

Mineralisation of this nature appears spatially related to ring structures. Flexural slip (F_2, Fig. 4.20) appears to have aided brecciation (and 'conditioning') along the schist-marble contacts. This brecciation has attendant quartz veining, which tend to 'foliate' around the margins of the ring structure. The replacement bodies have hardly any attendant alteration apart from ferruginisation, and consist mostly of hematite $\pm$ cassiterite and magnetite (Plates 4I, 4J). Locally they are particularly rich (e.g. 21.4% Sn over 6.9m, 21% Sn over 0.3m, in drill core), but are overall too small to exploit commercially (the Damara natives extract Sn profitably). Quartz vein mineralisation includes cassiterite, galena, $\pm$ pyrite, pyrrhotite and rutile. Cassiterite content decreases with distance away from the ring structure (Fig. 4.21).

Overall the belts of Sn/W mineralisation in the Damara Province indicate the likelihood of numerous periods of magmatic and tectonic activity (during Damara and Karoo times), which no doubt have enhanced metal concentrations. The generation of these granites is attributed to anatexis and melting of the lower crust, although the mechanisms behind the emplacement of the Salem suite are likely to be quite different from that of the post Karoo alkaline complexes such as Brandberg, which can feasibly be attributed to hotspot/mantle plume activity in the upper mantle. The province as whole may well be characterised by geochemically specialised granites (i.e. granites with high background Sn/W and other incompatibles). As an example, in the Brandberg North area there are widespread granite intrusions, which to date have been grouped as Salem-type, Sorris-Sorris type, within which the specialised Ousis granite has been differentiated. Until further geological mapping and geochemical, petrological and radiometric dating studies are undertaken it is speculative to assign any particular granite as genetically related to the mineralisation, and furthermore present indications point to a fairly complex genetic and geochemical cycle in the evolution of the mineralisation.

The Ousis granite is known to be highly differentiated, and Sn-rich (up to 220ppm Sn). The emplacement of greisen and pegmatite veins related to this granite in contact zones parallel to the biotite schist foliation indicates a syntectonic emplacement, possibly prior to consolidation and metamorphism of the sediments. This substantiates that at least some of the observed mineralisation is of Damara age.

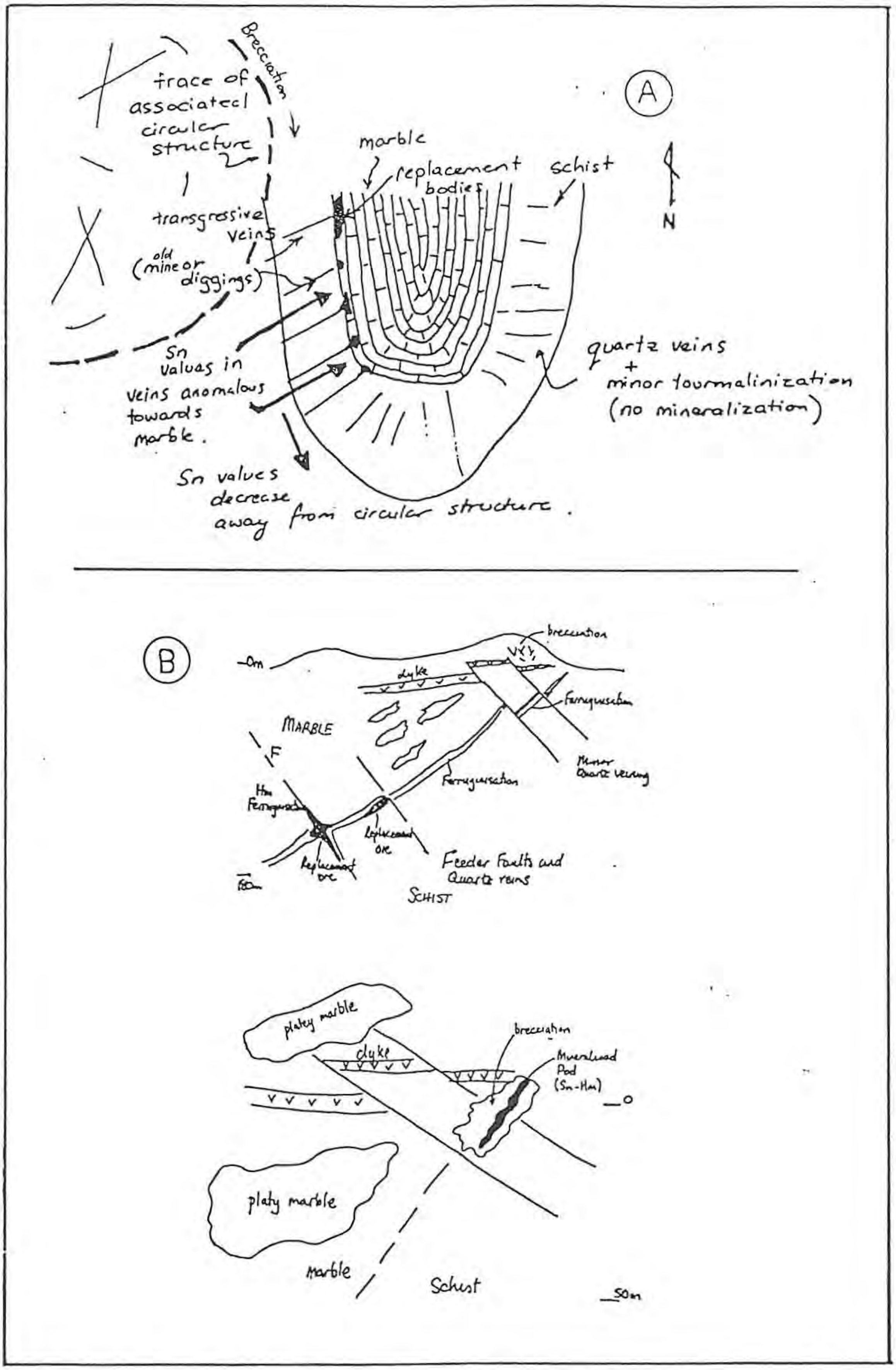


Figure 4.21: Sketches of the types of mineralisation found at the Goantagab prospect

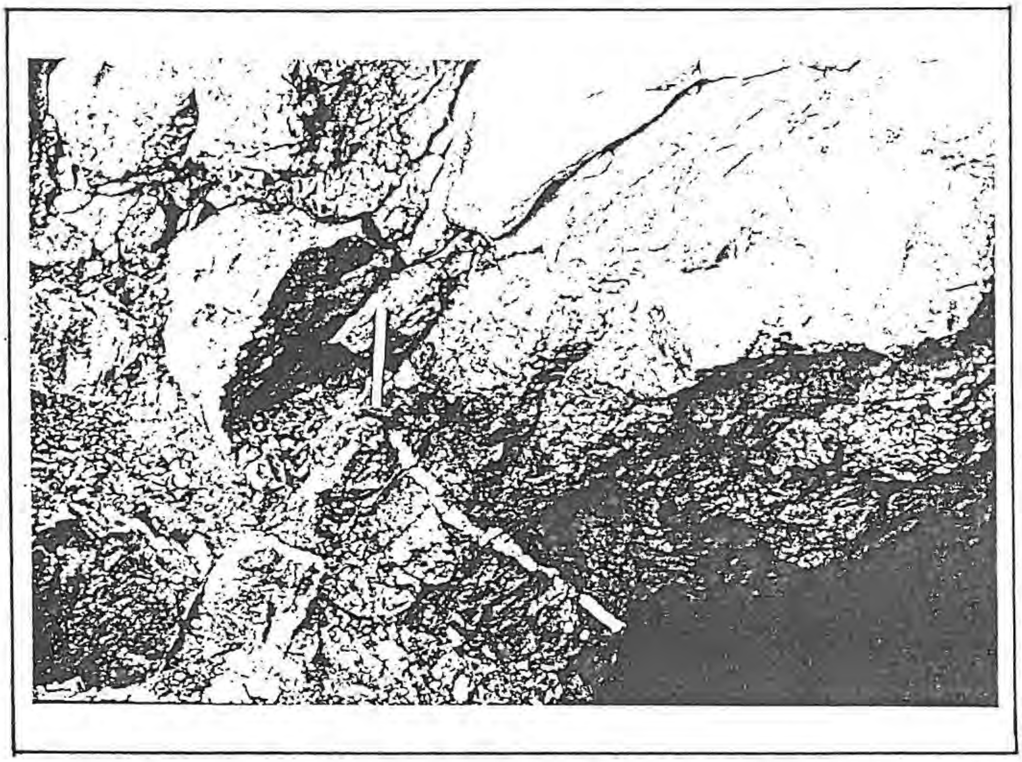


Plate 4I: Hematite $\pm$ cassiterite mineralisation marginal to a mined out pod, within marble at the schist/marble contact (Goantagab Prospect)

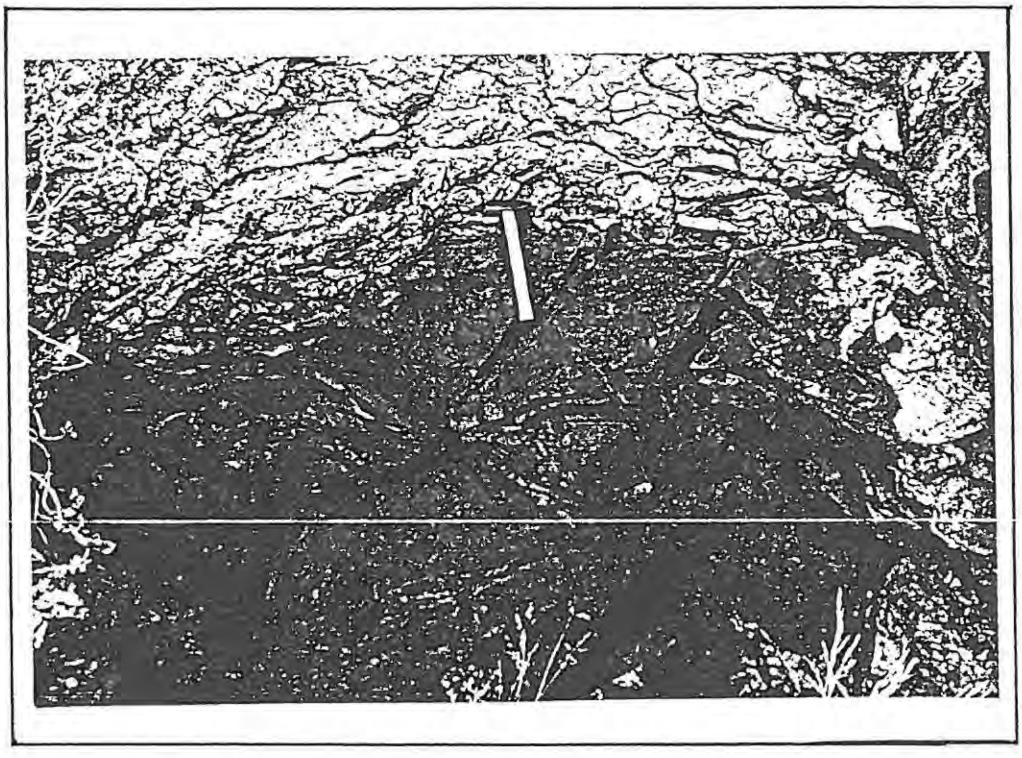


Plate 4J: View of an intact cassiterite-hematite replacement pod. Grades are high (up to 6-7% Sn) but bodies tend to be too small for economic exploitation by corporate enterprises

The late-tectonic Salem granite includes porphyritic biotite- and hornblende-biotite ± sphene granites. In general more siliceous lithologies that are less porphyritic are found towards the core of intrusives. The granite overall is a potential Sn-mineraliser, with a background content of about 10ppm Sn. The marginal zones of plutons are often characterised by xenoliths of more mafic "enclaves" (parental material) and digested country rock. Localised greisenisation is observable, with over short distances fresh microcline-biotite granite progressively muscovitised and capped by silicification, quartz veining and tourmalinisation. Contact zones are irregular, and interfinger with the country rock. Biotite usually superposes hornblende in these areas, and garnet and spotted biotite + cordierite hornfels may occur. Depending on the composition of the country rock sillimanite may occur right at the contact (if Al-rich) or else magnetite (if mafic). The Salem Suite granites also have associated mineralised ring and collapse structures, although it is unsure as to whether these are related to late stage intrusive phases of the suite or younger Karoo intrusives.

The Sorris-Sorris granite is also late tectonic, and intrudes the Salem-type granites (Salem xenoliths are incorporated within the Sorris Sorris). It has no notable Sn-enrichment, but displays greisenised border zones, with, similar to the Salem types, hornblende-biotite-microcline granites being altered to biotite-, then muscovite-granites, with later quartz-tourmaline veininv with greisenised margins.

A common denominator to all the mineral occurrences is the presence of greisen-type alteration, and this immediately places constraints on the nature of hydrothermal fluids involved in the ore-forming processes (see section 5.0). In the Brandberg North area, as much as it is difficult at present to pin point the exact granitoid-mineralisation relationship, one aspect is very clear, - the low grade quartzo-feldspathic and carbonate metasediments constitute an ideal environment for the formation of a wide spectrum of mineralised environments - as evidenced above. Furthermore the recognition of circular collapse structures and associated greisenisation and quartz veining clearly narrows down target evaluation. The variety of mineralised settings observed in the Brandberg North area can be assessed in terms of the probable depth of emplacement of the granitoid intrusions, coupled with the potential interaction of mineralised fluids with favourable lithological horizons.

4.3.2 Saldania Province

Tin and tungsten mineralisation in the Saldania Province is poorly documented, especially in terms of modern concepts concerning granite genesis, and related greisen and porphyry mineralised environments. This probably stems from the lack of economically viable deposits (to date) in the region.

Tin, as cassiterite, is the predominant ore mineral, whereas tungsten usually occurs as accessory wolframite. Most of the known Sn deposits are situated in a NNW trending zone extending from the slopes of Helderberg near Somerset West in the south east, past Kuilsrivier and Durbanville, to an isolated occurrence near Yzerfontein on the west coast in the NW (Figs. 4.22, 4.23; Crocker and Callaghan, 1979). Minor Sn has been detected (Beeson, 1978 in Crocker and Callaghan, 1979) in stream sediments taken adjacent to the George plutons to the east.

In a spatial context the deposits range from endogranitic in the S.E. (Kuisrivier- Helderberg area) to exogranite the NNW (Koeberg, Hogekraal, Kanon Cop etc.). This feature was attributed by Scholtz (1946) to the granite in the SE pitching slightly NW. Proceeding NW from Helderberg the ore deposits along the mineralised zone are located within Malmesbury sediments at progressively higher levels relative to pluton roofs (Fig. 4.24). In this respect there is a possible vertical zonation of the ore assemblages. At Helderberg endogranitic assemblages comprise cassiterite, traces of wolframite, löellingite tending to arsenopyrite, chalcopyrite, and minor pyrite, molybdenite, tourmaline, apatite, zircon and traces of gold and silver. At Kuilsrivier endo- and exogranite assemblages are cassiterite with associated wolframite, löellingite, arsenopyrite, traces of molybdenite and tourmaline. Assemblages of cassiterite, arsenopyrite, pyrite, sericite, with traces of gold, silver, copper anmd bismuth are hosted within Malmesbury sediments to the NW. Overall the essential features are the abundance of sulphides and the lack of fluorine-bearing minerals.

Roof pendants of Malmesbury metasediments are hornfelsed (spotted, chlorite-cordierite), and tourmalinised where transected by quartz veins. Muscovite (often Li-bearing) is common as vein selvages, within veins, and country rocks. Contacts between the metasediments and the older granites are invariably intrusive, with hornfelsing in the

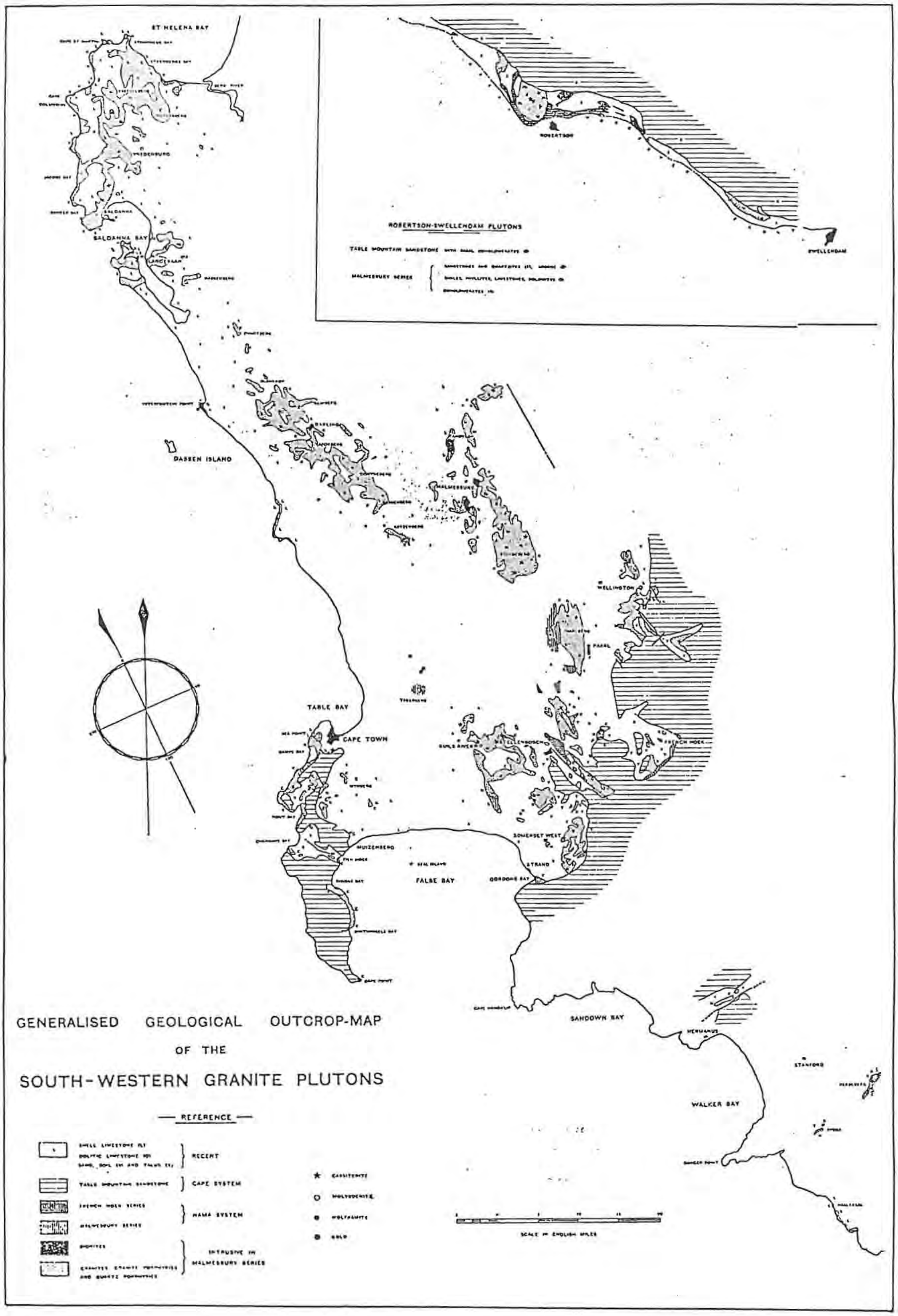


Figure 4.22: Generalised outcrop pattern of the south-west Cape granites (Scholtz, 1946)

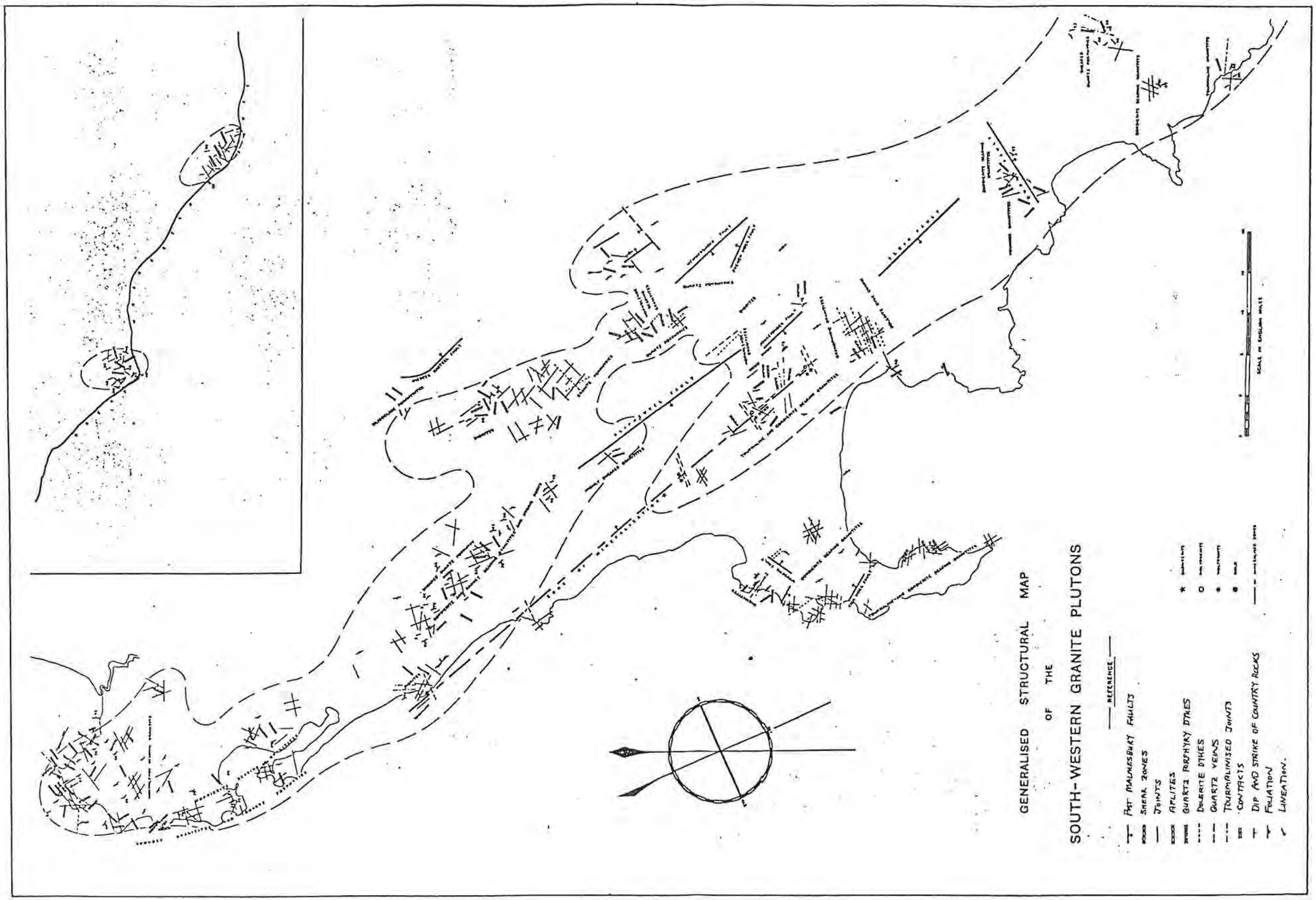


Figure 4.23: Generalised structural map of the southwest Cape granite plutons (Scholtz, 1946)

metasediments grading into zones of quartz stringers and veining, and granitic (greisen?) veins, and within the granite aplitic veining, minor pegmatites and tourmaline nodules (Krige, 1921).

The Kuilsrivier areas (Figs. 4.24, 4.25) is the best studied of the tin deposits, and yielded some alluvial production on Langverwacht Farm (Krige, 1921). Primary cassiterite mineralisation occurs in quartz veins up to 1.5m wide, dipping 15-20° east and associated with zones of intense shearing in coarse porphyritic biotite granite. The zones are up to 200-300m wide and can be followed intermittently along strike for 3km in a NNW direction. Albitisation and tourmalinisation of the host granites and muscovite selvages to quartz veins indicate a greisen environment, with mineralisation concentrated within the roof zone of a cupola stemming from a large batholith at depth.

It appears that most if not all the Sn mineralisation in the Cape region can be related to greisen-style mineralisation. As mentioned earlier in 3.3.1 it is suspected that there is a likelihood of attendant porphyry mineralisation with the younger intrusive suite, implications of which are discussed in section 5.0.

4.3.3 Bushveld Complex

Reviews of tin deposits associated with the acid phase and roof zone rocks of the Bushveld Complex are given by Crocker et. al. (1976), Lenthall and Hunter (1977), Crocker and Callaghan (1979), and Wilson (1979). Tin has been mined from six fields within the Bushveld Complex (Fig. 4.26), within which three mines, Zaaiplaats, Union and Rooiberg are still producing. Cassiterite is the main ore mineral, occurring within endogranitic pipes and primary disseminations (Zaaiplaats), or in exogranitic fissure veins, fault breccias or replacement bodies (Union and Rooiberg, Fig. 4.27). Further classification reflects host rocks, e.g. granite, granophyre, felsite and sedimentary rocks. Table 16 summarises salient features of the mineralisation, and Fig. 4.28 illustrates a model of the various types of tin deposit. It can be seen that a wide variety of deposits have been discovered within the Bushveld Complex (see Crocker and Callaghan, 1979), and only a brief resumé of that exhibited by the producing mines is presented here.

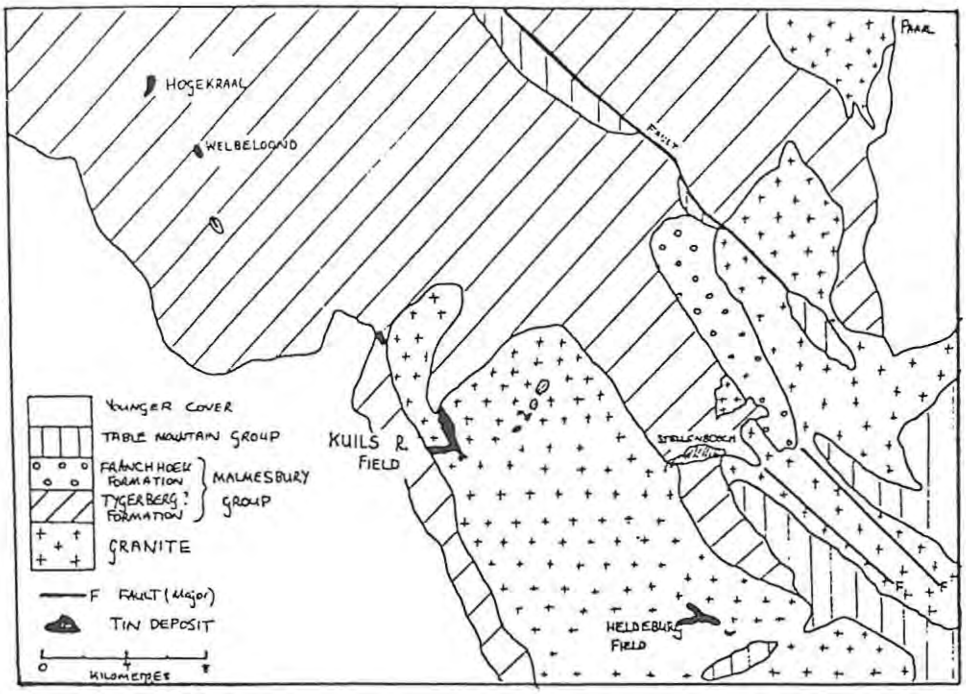


Figure 4.24: Simplified geological map of the Kuilsrivier region, showing tin deposits within granite and Malmesbury Group metasediments (adapted from Krige, 1921)

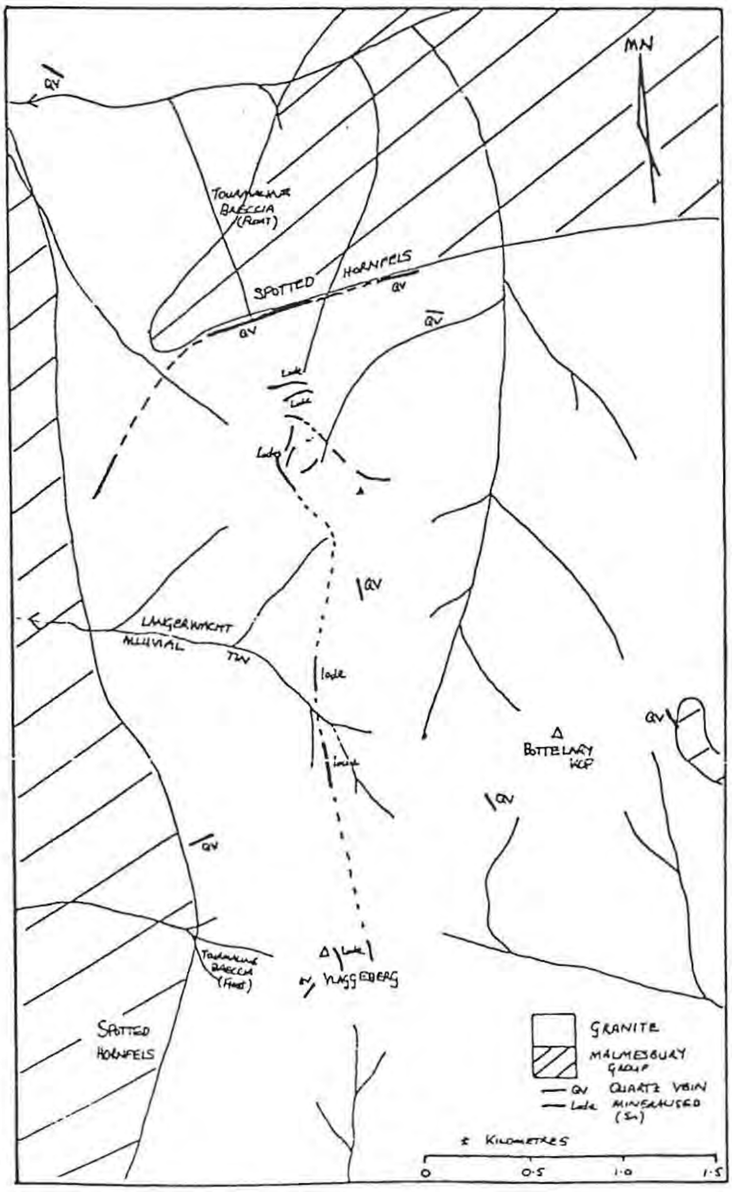


Figure 4.25: Simplified local geology of the Kuisrivier tin deposits (adapted from Krige, 1921)

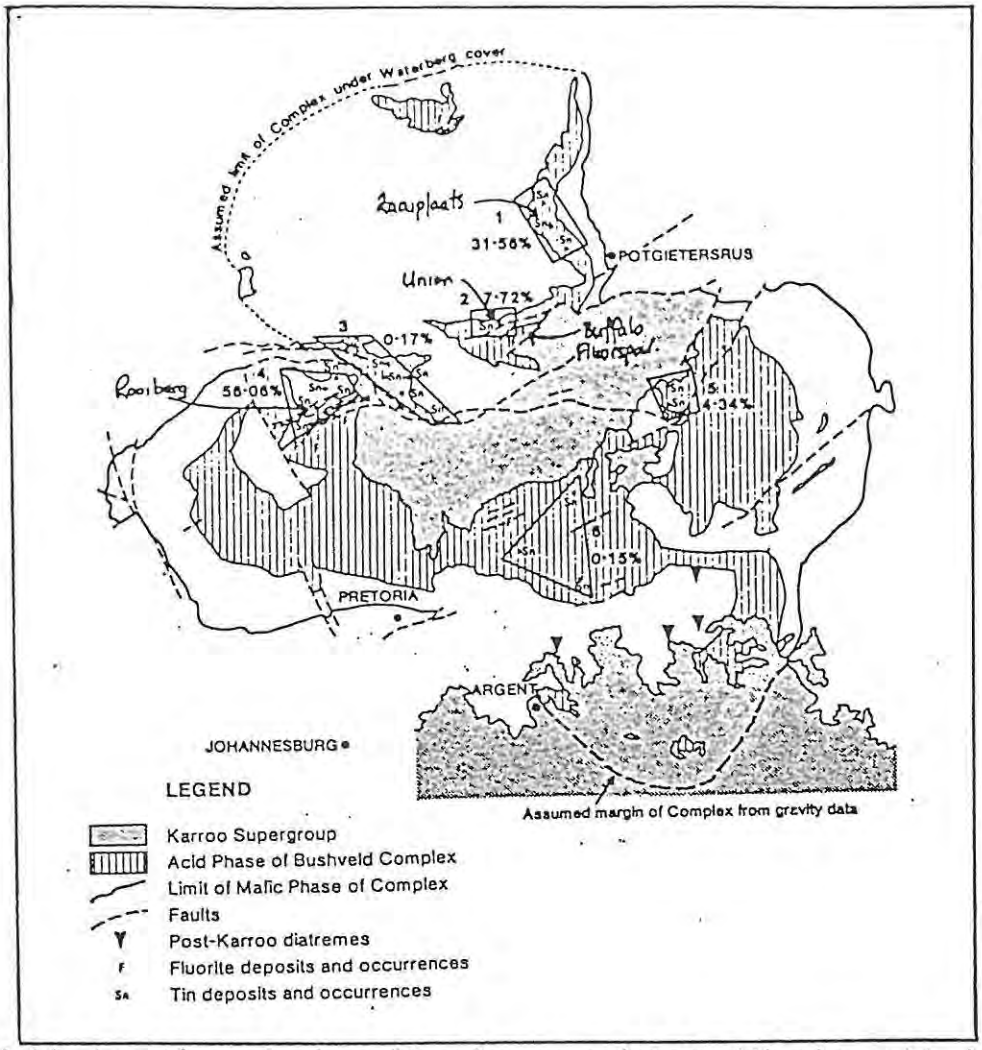


Figure 4.26: Location of mineralisation associated with the acid phase of the Bushveld Complex. The six main tin fields are shown with an indication of the proportion of the total tin production contributed (till 1976) 1, Potgietersrus; 2, Nystroom; 3, Elands; 4, Rooiberg; 5, Olifants; 6, Moloto (after Hunter, 1976; Lenthall, 1974)

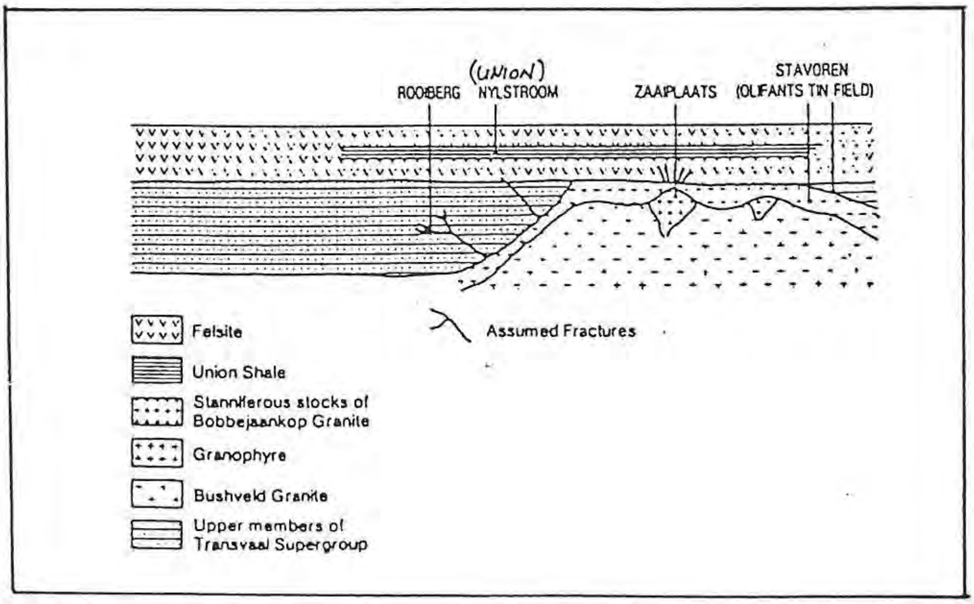


Figure 4.27: Schematic representation of the main locales of tin mineralisation in the Bushveld Complex (Hunter, 1976)

Table 16: Some features of tin mineralisation associated with the acid phase of the Bushveld Complex

Six principal fields: Mineralisation mainly Sn (0.6-3.0%), ± minor W, As, Cu, Fe, Pb, Zn. Alteration: sericite, quartz, chlorite, tourmaline, granite with miarolitic textures.

Tin field/ deposit	Host rock	Type of ore body	
Potgietersrus/ Zaaiplaats Solmans Temple	Bobbejaankop and Lease granites	Disseminated endogranitic either within granite (low grade) or pipes (very high grades). Some associated W, Ag, and base metal sulphides	Flat lying 12-15m thick zone, on flank of ENE warp. Pipes highly erratic, unpredictable
Rooiberg/ A, B, C, Mines	Pretoria Group sediments A Mine - bedded qtzite B Mine - Stock-work replacement of qtzite. C Mine - fractures in qtzite and shale	Replacement bodies in arkoses and shale, localised annular structures on shears, veins in shears. Generally intense tourmalinisation	Mineralisation related to Lebowa Suite intrusion, localised on NW axis folds, and thrust faults on syncline flanks. These have acted as feeders.
Nylstroom/ Union Mine	Rooiberg felsites, and interlayered Sterk River shale	Mineralisation within or adjacent to fissures in felsite or shale.	On northern limb of W plunging antiform. Extensive faulting (Doornhoek and Welgevonden Faults)
Eland	Granophyre	Mineralisation in fissures transecting pegmatite and granophyre. Alluvial and eluvial deposits have been of significance	Prospected in 1973
Olifants/ Mutue Fides Stavoren	Bobbejaankop Granite, Rooiberg felsites, Granophyre	Fissure and lode deposits, small pipes. Local greisenisation associated with mineralisation. Diverse sulphide assemblage - py, stannite, cpy, aspy, mo, bi, gl, sp, sch, wolf.	Not worked anymore
Moloto	Bobbejaankop Granite, granophyre	Minor deposits, mostly fissure and lode deposits	Prospected in 1973

Zaaiplaats

Endogranitic mineralisation at Zaaiplaats Mine is restricted to the apical portions of the Bobbejaankop Granite (a late stage differentiate of the Lebowa Granite Suite), and to a fine-grained "hood facies" - the Lease Granite (Fig. 4.29). Primary disseminations within these zones constitute low-grade ore (0.15-0.20% Sn). These extensive zones contain cassiterite grains to 6mm, and generally taper off in grade with depth from the apical portions of the granite. Locally the granite may be altered to epidote-chlorite-sericite assemblages.

Highly erratic and unpredictable pipes with concentrations of up to 30% Sn provide irregular highgrade ore (Plate 4K). The cylindrical pipes may reach up to 1000m in length with diameters of less than 1m to over 10m (mostly 1-2m). The pipes are distinctive annular structures with mineralogical zoning - an illite-sericite core, cassiterite-tourmaline-quartz margin, grading through sericitised and albitised granite to relatively fresh microcline granite. The pipes are distinct in that they only carry tin in areas where disseminated ore zones within the Bobbejaankop Granite are transected.

Relatively rich mineralisation (0.4-5.0% Sn) also occurs as lenticular disseminated ore bodies (Lily Pads) within the Lease Granite, and generally located against or close to the pegmatite roof zone. An alteration halo comprising albite-epidote-sericite-illite-cassiterite assemblages going inwards is usually associated with these bodies.

Union Tin Mine

Mineralisation is confined to within or adjacent to fissures in Rooiberg felsite (ignimbrite) and the interlayered Sterk River shale. The Mine is situated on the northern limb of the regional Swaershoek anticline, which plunges west at a low angle. The area is marked by extensive faulting, in particular the Doornhoek and Welgevonden Faults (Fig. 4.30). The Doornhoek Fault duplicates rocks of the Rooiberg Group, with an ondip displacement of 800-1000m.

Tin mineralisation, as cassiterite, is associated with hematite, magnetite, and lesser amounts of chalcopyrite, arsenopyrite and galena, invariably adjacent to and within fractures and fissures discordant to the Sterk River shale (Plate 4L).

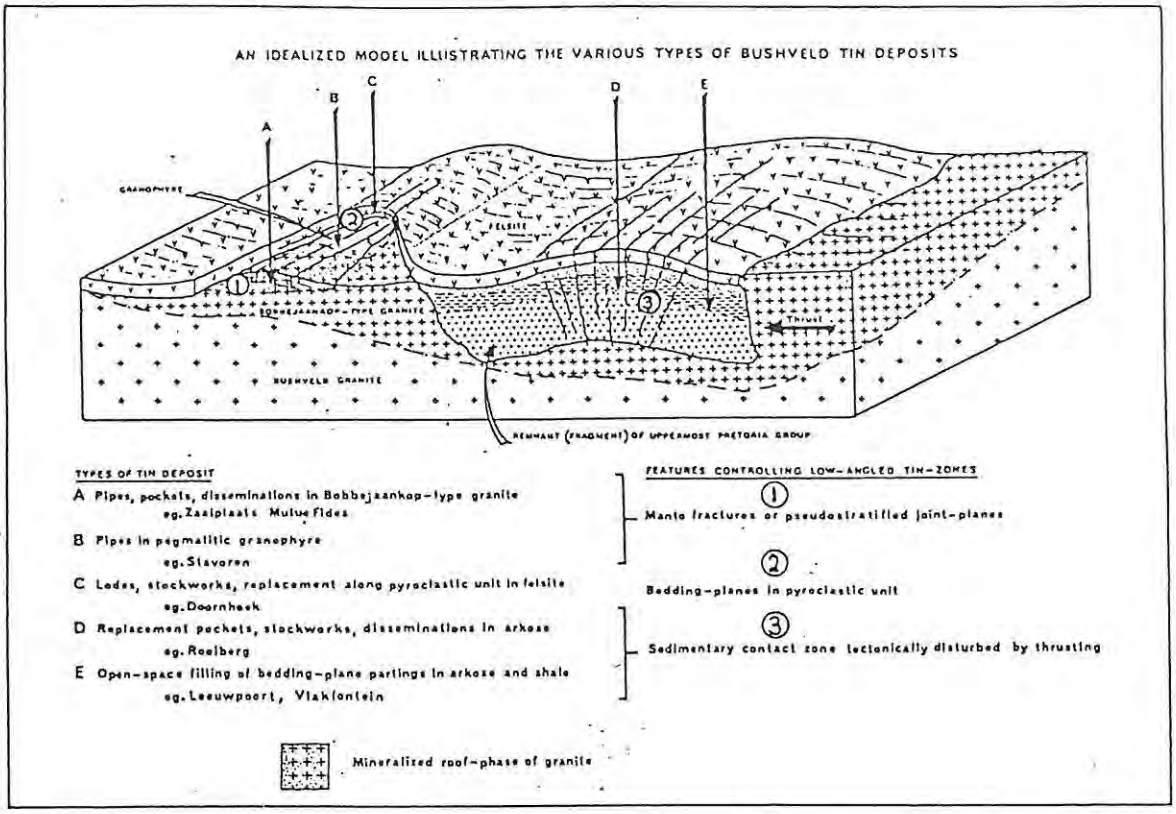


Figure 4.28: Idealised model illustrating the various tin deposits in the Bushveld Complex (Stear, 1977)

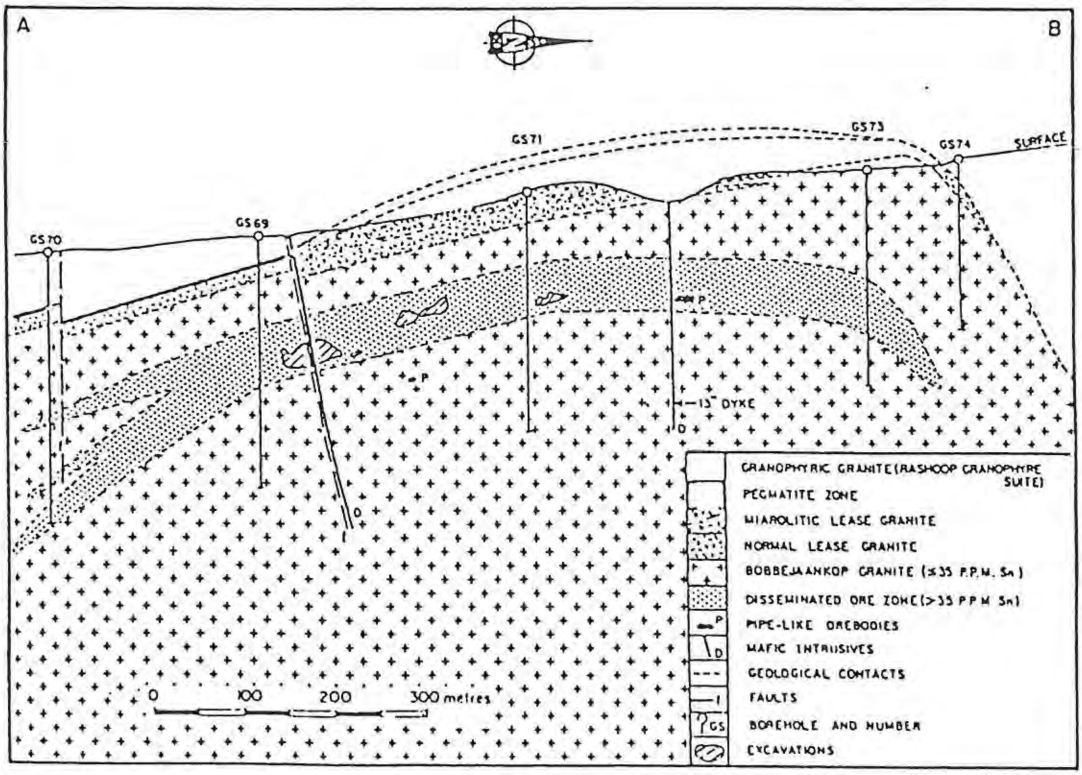


Figure 4.29: Geological cross section of Zaaiplaats Tin Mine, showing distribution of the principal tin mineralisation (Coetzee, 1983)

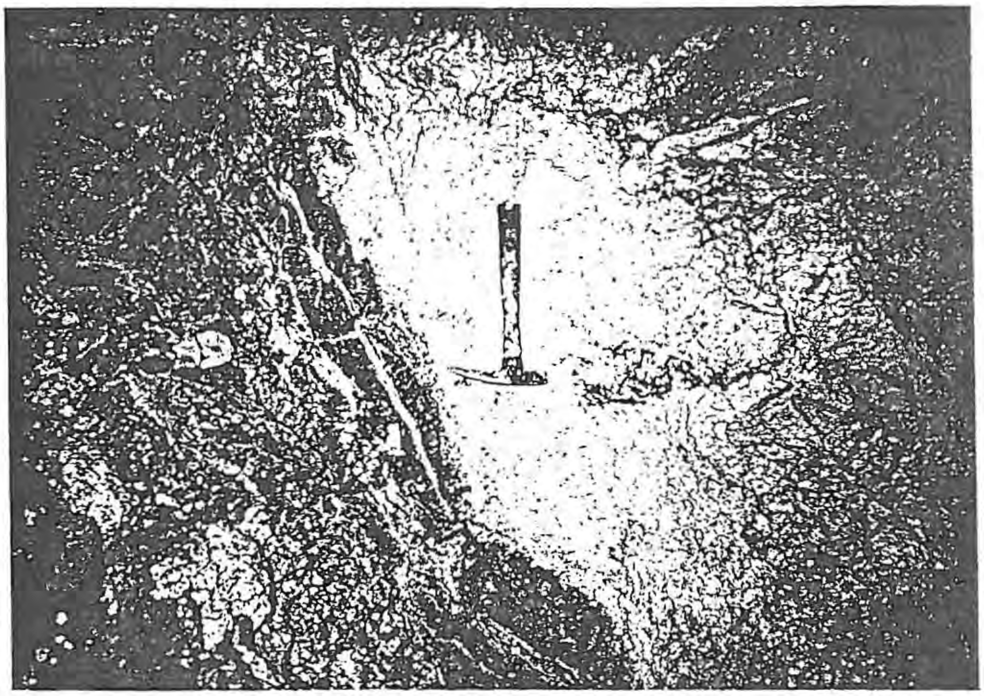


Plate 4K: Cassiterite-bearing pipe, Zaaiplaats Tin Mine. An inner sericitised core has grades of ±1% Sn, whereas the outer halo may contain up to 30% Sn

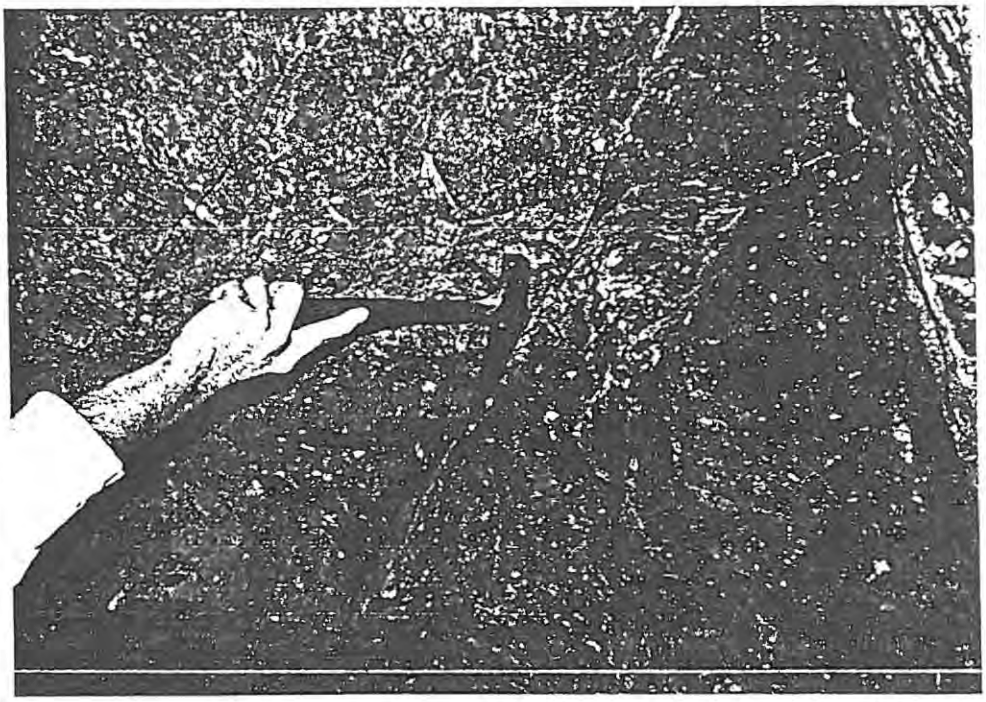


Plate 4L: Typical discordant fracture (the N3 Lode) within Sterk River Shales which hosts cassiterite mineralisation, Union Tin Mine cassiterite disseminations, locally with quartz veins and disseminated sulphides characterise this lode-type. No. 1 type Mine geologists distinguish three main lode types: N, No. 1, and 120-types (Fig. 4.31). The N-type lodes dip east at $\pm 45°$, and are the principal cassiterite carriers. Chloritised wall rocks with fine lodes transect the shale bedding, and are sub-parallel to the regional Welgevonden Fault. These lodes have a stong hematite-magnetite association and show wide variations in cassiterite content where mineralised. The 120-type lodes strike parallel to bedding, and dip 55°N. A genetic relationship with the Doornhoek Fault is envisaged, and reverse fault/movement may have remobilsed some mineralisation, as it is generally low grade (0.1-0.3% Sn) and at best erratic.

Rooiberg

The Rooiberg tin-field is situated 65km west of Warmbad within the uppermost sediments and volcanics of the Pretoria Group of the Transvaal Supergroup (the "Rooiberg fragment"). Cassiterite is presently recovered from four producing mines:- Hartbeesfontein ('A' mine), Niewpoort ('B' mine), Leeuwpoort ('C' mine) and Vellefontein (Fig. 4.32). Numerous other occurrences are known in the fragment. The lower portion of the stratigraphic succession at Rooiberg (Fig. 4.33) consists of a fining-upward sequence of sediments, in which the basal Boshoffsberg Quartzite Member grades into the Blaauwbank Shale Member. This is unconformably overlain by the volcano-sedimentary Smelterskop Formation, which is capped by felsite of the Rooiberg Group.

Tin mineralisation in the Rooiberg fragment occurs as replacement and bedded lode deposits. The replacement deposits occur as spectacular, irregular-pipe like bodies located along, or at the intersections of steeply dipping fractures. Bedded lode deposits occur within a specific stratigraphic interval - in the cross-bedded feldspathic Boshoffsberg Quartzite Member (termed arkosite) and in the overlying transitional shaly-quartzite below the Blaauwbank Shale Member (Fig. 4.33).

Mineralisation does not extend past the Blaauwbank Shale Member, which appears to have been impervious to ascending fluids. It is preserved within the SE-trending Blaauwbank syncline (Fig. 4.32), one of three synclines formed in the region during emplacement of the Lebowa Granite suite. The axes of the synclines generally parallel the intrusive

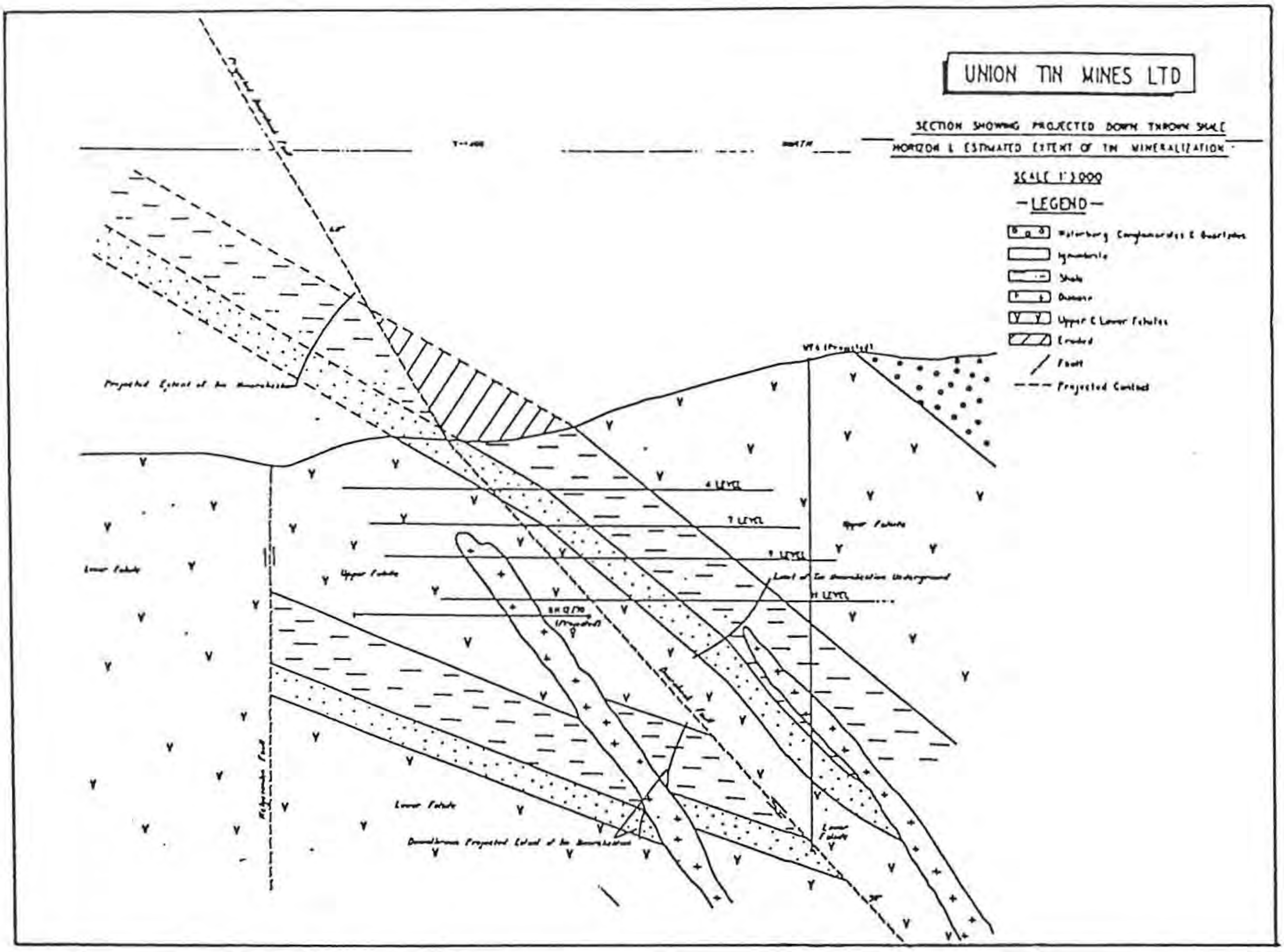


Figure 4.30: Geological section of Union Tin Mine (Mine Staff, 1984)

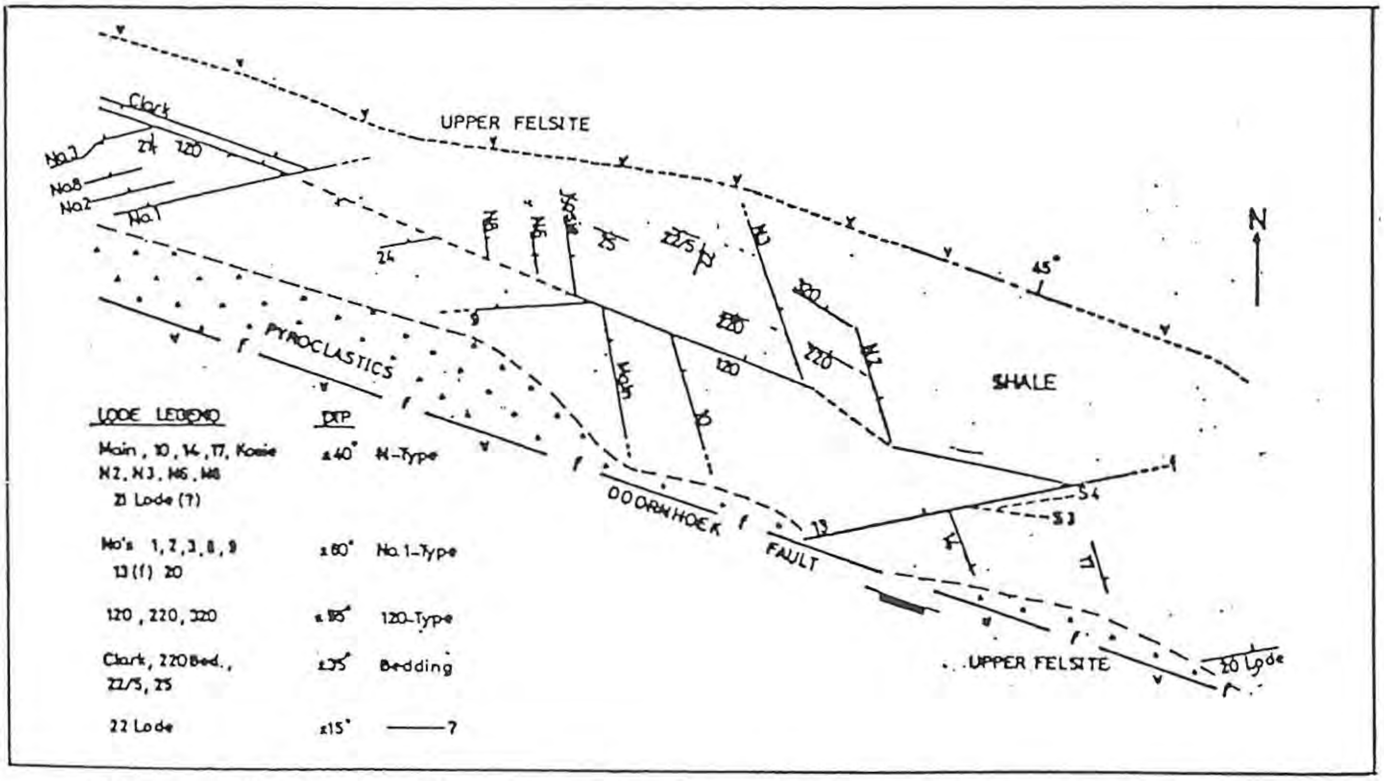


Figure 4.31: Union Tin Mine, principle fracture trends associated with mineralisation (Mine Staff, 1984)

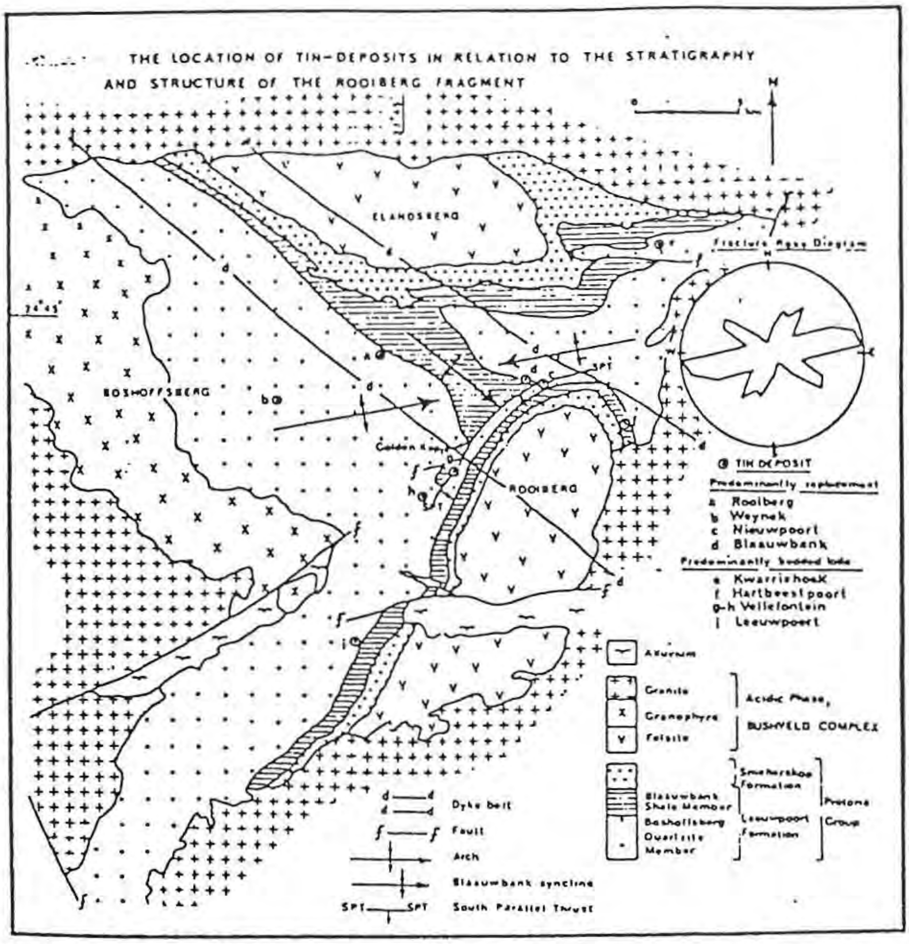


Figure 4.32 Location of tin deposits in relation to the stratigraphy and structure of the Rooiberg fragment (Rooiberg Staff, 1984).

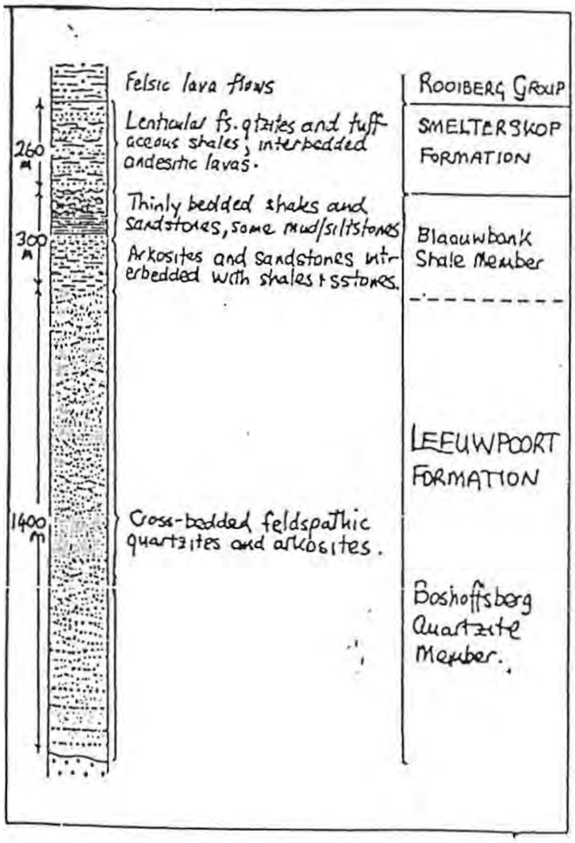


Figure 4.33: Stratigraphy of the Pretoria Group at Rooiberg (Rooiberg Staff, 1984).

contacts of the granite. The other major structural feature, subsequent to the folding, was thrust faulting along the flanks of the synclines, dipping towards the granites (Rooiberg Staff, 1984). The thrust fault zones are often characterised by brecciation with pale pink quartzite fragments in a dark tourmaline- and chlorite-rich matrix. They are considered as primary feeder channels for the ore-forming fluids.

The various mines have different styles of mineralisation, although the bulk of production comes from exploitation of the stratabound 'bedded' lode deposits. 'A' mine has cassiterite mineralisation associated with 'pockets' which are randomly orientated and bedded, around 90% being controlled by fractures in the arkosites (Plate 4M). 'B' mine is characterised by discordant fracture controlled mineralisation, as well as stratabound cassiterite along bedding planes, with occassional 'boudinaged' pockets. Similar mineralisation occurs at 'C' and Vellefontein mines.

Greisenisation of the sediments is a common and widespread phenomenon at Rooiberg. Typical alteration assemblages include chlorite, sericite (after feldspar), muscovite, recrystallised quartz, fluorite, siderite, ankerite and tourmaline. Sulphides are present along mineralised fractures (e.g the Magazine fissure lode), and in zoned alteration haloes associated with replacement pockets. Pyrite and chalcopyrite are common, the former hosting substantial Ni (6000ppm) and Co (5000ppm) when in bedded lode deposits. Minor pentlandite and bravoite may also be associated with the pyrite. Other sulphides including galena, arsenopyrite, pyrrhotite and sphalerite have been reported from the mines ('B' mine in particular has aspy and antimonides), whilst minor scheelite is found in pockets at 'A' mine, and Fe-oxides such as magnetite in 'C' mine.

Alteration zonation associated with the replacement pockets is highly variable. Cores may consist of cassiterite-tourmaline or quartz-siderite-ankerite (Plates 4N, 40), with variable sulphide content. Invariably there is an outer tourmaline halo. A feature of these bodies is that prominent cross-bedding of the arkoses may be traced into the outer rims of the structures (Plate 4P), good evidence of the hydrothermal replacement nature of the fluids (and vapours).

PLATE 4N: Typical pocket mineralisation and zonation, 'A' Mine. Note the well defined quartz-carbonate-tourmaline core, outer pyrite halo, and peripheral tourmaline. The controlling fracture runs vertically up the photograph. Tourmalinised bedding planes are visible top left.

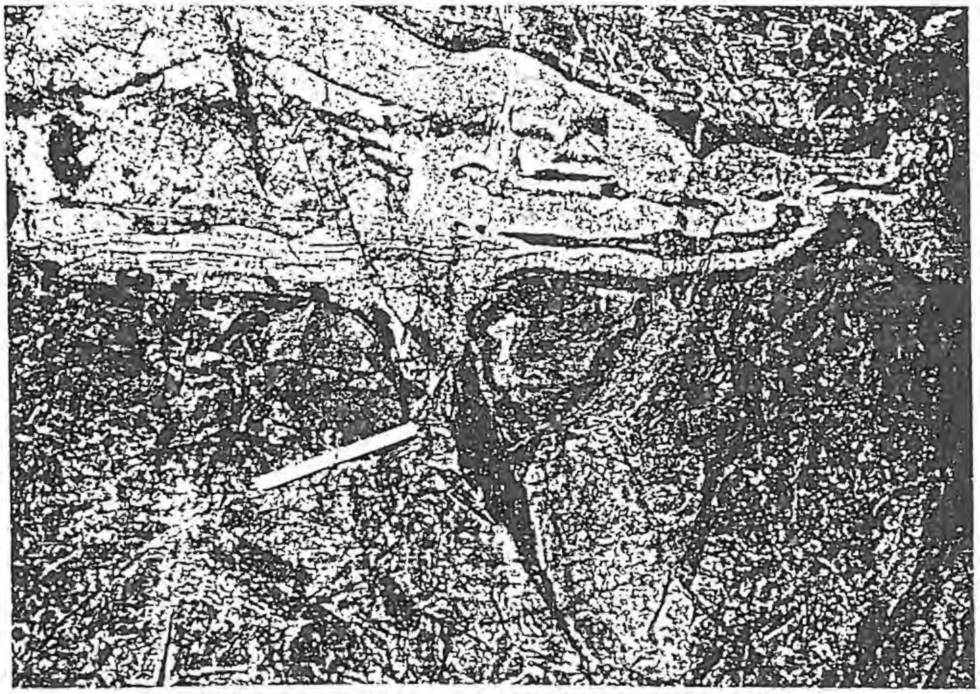


PLATE 4P: Annular pocket superimposed on cross-bedded arkosite. The bleached alteration is mostly silicification, with minor carbonisation, and zones of tourmalinisation. Note the "feeder pipe" at base

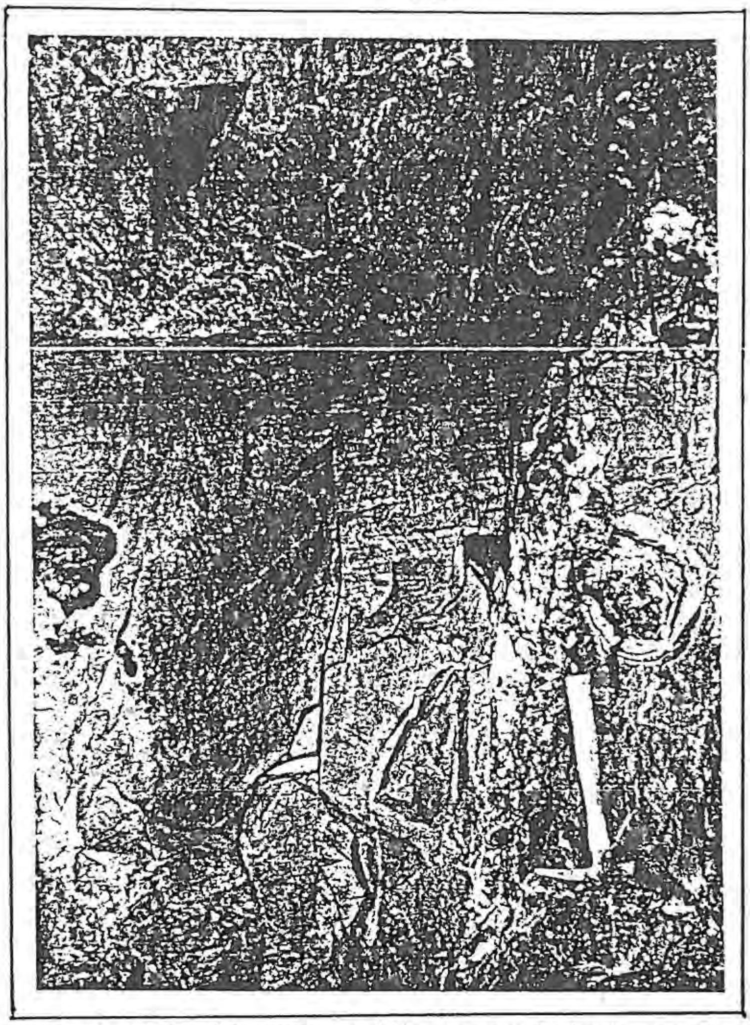


PLATE 4M: Brecciated fissure with associated replacement pocket (right), truncating finely tourmalinised (+cassiterite) bedding planes in arkosite (left). Rooiberg 'A' Mine.

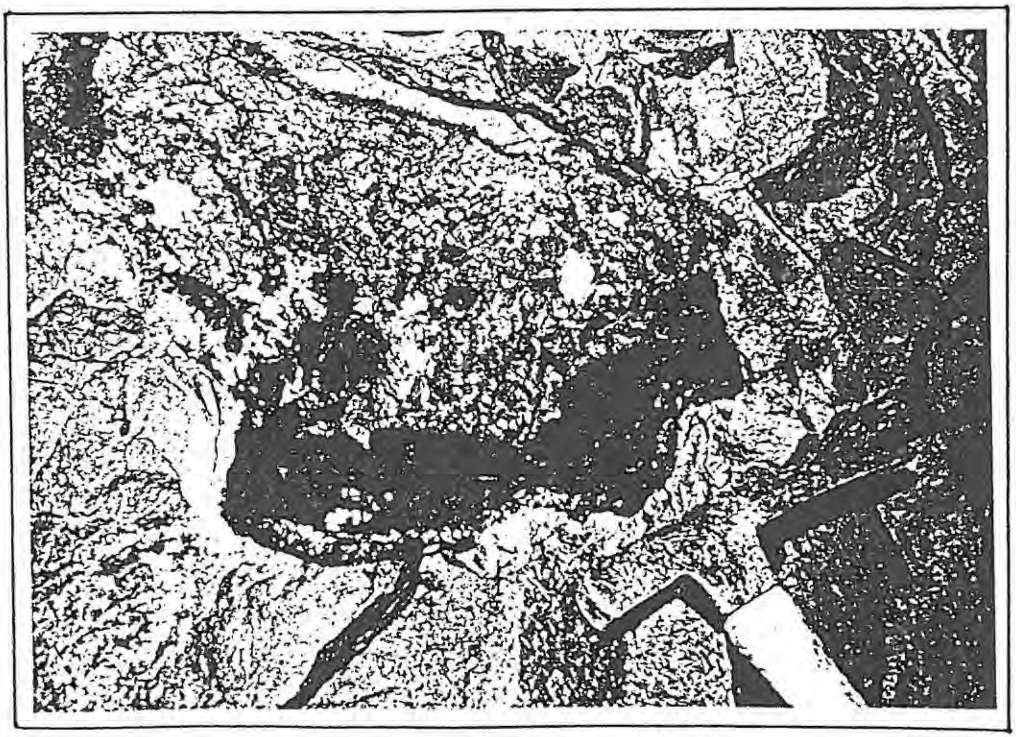


PLATE 40: Ankerite-quartz-pyrite core to pocket, with cassiterite-tourmaline outer halo. Pyrite occurs at outer margins of each zone.

Genetical Aspects

Overall the tin deposits of the Bushveld Complex are genetically related to end member differentiates of the Lebowa Granite Suite. Variety in setting is imposed largely by the intrusion of these granites into different roof rocks - in particular leptites of the Rashoop Granophyre, felsites and shales of the Rooiberg Group, and various sedimentary successions of the Transvaal Supergroup. Mineralogy of ensuing mineralisation varies with the nature of host rocks, the presence of suitable structural channel ways, and whether or not volatilisation and/or hydraulic fracturing and/or throttling of late stage fluids occurred. Some aspects of the genesis of the deposits are outlined below, as they often constitute unique mineralised settings, although there are some overall similarities to the other deposits described earlier.

Geochemical studies of the Bushveld acid magma phases (Lenthall and Hunter, 1977; Crocker, 1979; Wilson, 1979) point to an anatectic source, brought about by selective fluorine-fluxing of sialic crustal rocks. Differentiates of the Lebowa Granite Suite, such as the Bobbejaankop granite, are derived from hornblende granites, and are marked by volatile-rich biotite granites, often localised in the core of intrusions. Late stage deuteric and post magmatic alteration in the late crystallising cores occurred as a result of continued enrichment of the melt in residual and volatile constituents. This deuteric alteration, which results in perthitisation of feldspars, chloritisation of biotites and oxidation of released iron (from feldspar lattices) gives rise to the typical "red granites" almost always associated with mineralisation.

Endo- and exo-granitic mineralisation can be distinguished, the former as at Zaaiplaats (Sn) and Buffalo Fluorspar (F) Mines, the latter as at Union and Rooiberg Tin mines. Volatile action and possible pre-existing structures probably play important roles in whether or not mineralising fluids intrude roof zone lithologies. Lenthall and Hunter (1977) consider that remobilisation and movement of high level granite may occur where large volumes of volatiles collected in high zones of the granite roof, and migration of the volatiles down pressure and temperature gradients depressed the melting point of the magma. Resolidification after essentially lateral movement would occur when the melting-curve was re-crossed. Observed disseminated mineralisation at Zaaiplaats is in roof zones of the granite, and the perplexing pipes may represent the last stage expulsion of vapour plumes of highly volatised fluids and gases.

Where roof-rocks were fractured, either by confining pressures exceeding lithostatic pressures, or by tectonic movements along pre-existing or forming new fractures and fissures, volatiles were able to escape and penetrate roof-zone lithologies. There appears to be a case for preferential mineralisation of certain lithologies - e.g. the shale at Union Tin, or arkose at Rooiberg. Earlier discussions have highlighted either the fracture or bedding plane control on cassiterite-tourmaline deposition. At Rooiberg, it is feasible that arkosite hosted mineralisation predates the main phase of mineralisation (fracture controlled) related to the Lebowa Granite. Observation of zoning in replacement 'pods' indicates probably two or three pulses of fluid, which would account for the often complex interrelationship between phases. The classic ore depositional sequence in exogreisen systems (world-wide) is of oxides (cassiterite- wolframite-scheelite-hematite), then sulphides (po, cpy, py, aspy, sulphosalts), followed by carbonates (calcite, siderite, or as, at Rooiberg, ankerite) (Bentley 1982).

The lower grade mineralisation generally mined at Rooiberg and Union could well be a function of distance from source. Rooiberg mineralisation was no doubt enhanced by pre-existing fracturing as suitable channelways, plus the porosity of the arkosite. It is feasible that the lack of permeability of shales in the Union area may have prevented higher grades forming.

Crocker (1979) draws attention to mineralogical associations such as hematite-quartz-fluorite within granites indicating roof zone fracturing and the likelihood of tin being in roof rocks. The concurrence of fluorite and cassiterite in granites infers no fracturing of roof zone rocks. Telescoping of mineral assemblages may occur if fracturing is very late in the mineralising phase.

Finally, attention is also called to the concurrence of the fluorite deposits (and some tin) on intersecting NW/ENE (i.e Vryburg Arch-Murchison trend) lineaments and fractures. It is held (Hunter, 1976; Wilson, 1979) that these are long lived fractures or tectonic zones within the Kaapvaal Province (see Fig. 3.29). Even today hot springs and high F-content ground waters occur in areas of intersection of these lineaments. The role of intracontinental rifting in the development of the supracrustal sequences of the Kaapvaal Province is equivocal, however the close association of basin axes, extrusive centres, kimberlite occurrences, and the above described deposits may be fundamental to major extensional tectonics at various stages of the Proterozoic. The concept also holds for stratabound mineralisation (e.g. Rooiberg arkosites), whereby rifting would allow access of mineralised fluids (volatile) emanating from evolved granitic melts to penetrate into a domain possibly including water-sediment interfaces, or at least where not much compaction had occurred, and readily porous sediments were available for ore fluid permeation and deposition.

5.0 COMPARISON OF ENVIRONMENTS AND IMPLICATIONS FOR Sn-W EXPLORATION IN SOUTHERN AFRICA

In view of the preceeding discussion, there are many common features between the various deposits, regardless of the given uniqueness of any substantial ore deposit (Table 17). Although only a limited number of geotectonic environments have been assessed, there is good evidence that specific metallogenic provinces can be assigned to a particular geotectonic setting, and that the chemistry of related granitoids plays a significant and definable role in the generation and emplacement of the ore deposits. Furthermore the role of greisenisation is apparent in all the Sn/W deposits discussed (Table 17). Even though the morphology of the end ore occurrence is highly variable ie. exogranitic quartz veins, pegmatites, skarn or replacement bodies as opposed to endogranitic veins, pegmatites, primary disseminations, pipes and carbonas, the underlying mechanisms are largely similar in all cases. There appears a close spatial and temporal association of Sn and W deposits with areas of crustal thickening and generation of S-type ilmenite series granitoids, especially in collision belts. These deposits resemble in terms of morphology and zoning the magmatic hydrothermal vein-type deposits associated with I-type magnetite series granitoid intrusives in subduction related terrains. This is no doubt largely due to the hydrothermal fluids emanating from shallow-seated magmatic systems. As has been emphasised, the main differences lie in the granitoid chemistry, and the predominance of tin and tungsten in the former, and base and precious metals in the latter. If tin and tungsten do occur in arc systems, the deposits tend to be confined to the innermost zones of continental margin arcs where the intrusions manifest a distinct S-type character.

The Sn-W deposits described all exhibit a close spatial association with the roof zones of granites derived from the anatexis of continental crust, and emplaced during the late stages of orogenesis (Table 17). The underlying common denominator and metallic zonation (where apparent) in all the deposits is alteration and metal deposition related to greisenisation (see 2.4).

Table 17: Comparison of the main geotectonic and metallogenic features of the Variscan, Malaysian and southern African Sn-W deposits

LOCALITY	GEOTECTONIC SETTING	PREDOMINANT MINERAL DEPOSITS	HOST LITHOLOGIES	PREDOMINANT RELATED GRANITOIDS, AFFILIATION TO MINERALISATION
Iberian Peninsula France	Bipolar subduction, extensive crustal anatexis, tectonic thickening, thrusting	Greisenised stockworks, vein complexes or fissure lodes. Local skarns/tactites	Variable Lower Paleozoic metasediments and carbonates	S-type ilmenite series. Two phases, Mid-Carboniferous and Late-Carboniferous to Permian. Mineralisation spatially associated with intrusive contacts
Cornwall Peninsula S.W. England	Continental collision. Possibility of granites tectonically emplaced (thrust from SE)	Greisen bordered vein swarms and fissure veins	Lower Paleozoic shelf and deep water flysch sediments	Overall affinities to S-type ilmenite series, derived from crustal anatexis Min^n spatially related to roof and border zones of high level post-tectonic granites (303-265ma)
Malay Peninsula	2 phases of subduction, followed by continental collision	Alluvial tin derived from greisenised vein swarms. Localised skarn-replacement mineralisation	Lower Paleozoic flysch type sediments	S-type ilmenite series granitoids (Permian-Triassic) Min^n spatially related to outer zones of apical portions of intrusive plutons
Damara Province Namibia	Intracontinental rifting hotspot magmatism?, small ocean development, closure with collision and/or limited subduction. Substantial anatexis of continental crust	Greisenised vein swarms, pegmatites, skarn and replacement bodies	Late Proterozoic -Early Cambrian turbiditic and shelf sediments (schists, marbles calc silicates)	S-type ilmenite series granitoids (anorogenic-type) Min^n mostly exogranitic, spatially related to apical and intrusive contacts, circular structures
Saldania Province South Africa	Intra- or inter-continental rifting (more passive margin tectonism). Extensive stretching and thinning of crust, followed by closure, listric faulting and thrusting	Greisenised fissure veins	Late Proterozoic -Early Cambrian turbiditic meta-sediments, lesser carbonates	Dominantly S-type ilmenite series, primary mineralisation in granite roof zones and metasedimentary roof pendants
Bushveld Complex, South Africa	Anorogenic (hot-spot?) massive layered ultramafic complex with associated late stage granitic differentiates	Locally greisenised fissure lodes, replacement bodies, primary disseminations in granite	Granophyre and felsite of Bushveld Complex, Proterozoic carbonates (Transvaal S.G.) and quartzofeldspathic sediments	High level differentiated granites from crustal fusion and anatexis (Prob. S-type ilm. affinities exo- and endo-granitic minerali-sation

5.1 Implications for Sn-W exploration in southern Africa

An immediate conclusion from the above discussion is that there is ample scope and potential for Sn-W exploration in southern Africa. The Bushveld Complex ranks as a priority area for any Sn exploration, and potential target areas are detailed by Crocker and Callaghan (1979), and will not be expanded upon here. In terms of the Cape Province and Namibia there is very good potential for an integrated assessment of the granitoid terranes. By this one emphasises the need for a fresh approach to exploration, utilising existing data (if available), but above all approaching the task in a systematic manner.

In the context of the present work, an important first step in any such exploration programme is the recognition of S-type ilmenite series and I-type magnetite series granitoids. Furthermore this approach could aid the delineation of metallogenic provinces - the SW Cape granitoids appear to show affinities to porphyry Mo-Cu-W? and greisen Sn-W provinces. This necessarily will constrain working exploration models, and ultimately the type of ore deposit explored for.

It is contended here that in the Damara and Saldanian Provinces a prime target should be an intact endo-exogreisen sysem $\pm$ associated skarn-type replacement deposits. The latter appear more likely in the Damara Province (e.g. Brandberg North) because of the presence of fairly substantial marble horizons intercalated with the Kuiseb schists. These deposit types should not be overlooked as they provide small tonnage high grade ore (e.g. around 1000t @ 6-20% Sn) which would bolster the larger tonnage lower grade (e.g. 10-40mt @ 0.2-0.6% Sn + WO_3) greisen bodies. Considerable potential for exogreisen systems must lie within the Malmesbury Group metasediments in the Saldania Province. The methodology in exploring these terranes (which is applicable to any Sn-W exploration programme) is outlined below. There is a necessary bias to S-type ilmenite series granitoids, but, if W is the prime commodity, I-type magnetite series granitoids must also be assessed for porphyry W-Mo-Cu systems (e.g. especially in the Cape Province).

Integrated exploration is strongly advocated, with gradually more refined target assessment and evaluation. The methodology utilised can be broadly divided (for ease of discussion) into geological, remote sensing, geochemical and geophysical parameters.

Geology

Mineralisation is most likely to be spatially related to anatectic S-type ilmenite series granitoids. Metal association is enhanced within the youngest, most differentiated granites within a fractional crystallisation sequence from hornblende granodiorite, through biotite to biotite-muscovite granites. Mineralised areas are usually spatially related to intrusive contacts, and may be manifest as stockwork veins, single or multiple fissure veins with attendant greisenisation, pegmatites, as well as skarns where carbonate lithologies are intruded. Placer deposits of Sn are a good indicator of a primary source. Geological mapping on regional and local scale (in that order) is imperative, with special close scrutiny and allied lithogeochemical sampling (see below) of all roof zones and pendants. The identification of vein systems and greisenisation of wall rocks is important. Reconnaissance drilling leading to detailed drilling should support geological observations, and be substantiated by bulk sampling for realistic grade evaluation.

Petrographic studies should be integrated with all stages of mapping and drilling. This is an important parameter in regional and local assessment of areas, as the detection of thermal aureoles (e.g. biotite-cordierite hornfels) and superimposed greisenisation is often not readily perceived in hand specimen. Such studies also provide a necessary basis and understanding for lithogeochemical studies, and are essential for assessing the likely position one may be within a given system i.e. how far away from a mineralised cupola.

Remote Sensing

Reconnaissance studies must include Landsat imagery and aerial photograph interpretation. Such studies aid delineation of major lithologies and structures, including collapse ring structures (e.g. Brandberg North). The recognition of dyke patterns may elucidate structural patterns that have guided the emplacement of mineralised granite apophyses, and may also have petrochemical relationships to possible underlying Sn-W-bearing granitoids.

Geochemistry

Coupled with geological mapping, geochemical studies are seen as a major facet of exploration for Sn/W ore bodies in the Damara and Saldanian

Provinces. Related granitoids are likely to be specialised in Sn W, with enrichment also in U, Mo, Bi, Pb, Cu, Zn and incompatible elements such as Li, F, B, K, Rb, Nb, etc.). A thorough integrated geochemical survey allied with mapping programmes is advocated. Regional stream sediment and lithogeochemical sampling should delineate any Sn-W (or Mo-Cu) provinces. In the latter regard it is important to note that there are numerous methods by which to geochemically discriminate granitoids - including graphical plots such as $K_2O/Na_2O:SiO_2$; $Na_2O:K_2O$; K:Rb; Rb/Sr:Ti/Zr; Ti:Nb; Ti:Sn; Ti:Mg:Snx10. A regional zonation pattern may also emerge, usually with Sn-W rich cores and base metals outside. Diffuse base metal vein mineralisation may be a distal manifestation of subsurface Sn-W mineralisation. Of note is a common spatial (i.e. lateral) association of W (scheelite) - Au mineralisation (Bentley, 1982), emphasising the need to analyse for Au in these environments.

Geochemisty can also be utilised to delineate alteration haloes, including contact metamorphism and pervasive to weak hydrothermal alteration. An approach integrating petrography, lithogeochemistry and soil geochemistry (K_2O, Na_2, B, F, Sn, W, Mo, As, Cu, Pb, Zn) is advocated. Regional stream sediment and soil surveys must be initiated with orientation surveys to optimise elemental response to given dispersion patterns. In this respect completely different sieve size fractions would be utilised in the Damara and Saldania Provinces due to the vastly different climates (arid-semiarid vs moist). It is likely that coarser mesh sizes (e.g. #30 mesh) may prove more applicable in the former, whilst finer (e.g. #80 mesh) will have a better response in the wetter Cape Province climate. Stream sediment sampling should include heavey pan concentrates. The use of an ultraviolet lamp for detection of scheelite mineralisation in pan concentrates and rocks should not be overlooked.

Geophysics

Not many geophysical techniques are directly applicable to the exploration for Sn-W deposits. Of note is the determination of the magnetic susceptibility of granitoids, which significantly aids discrimination of S-type ilmenite series (less than 1×10^{-4} emu/g) and I-type magnetite series (more than 1×10^{-4} emu/g) granitoids (Ishihara et al., 1979; Ishihara, 1981). This also has direct repercussions in detection of metallogenic provinces. Radiometric surveys, such as gamma ray spectrometry, may be utilised for detecting anomalous K (implying potassic alteration) and U levels in granitoids (U is usually enhanced in specialised and mineralised granites vs non-mineralised). Magnetic surveys may aid detection of skarn-replacement type deposits, whilst to a lesser degree induced polarisation and electromagnetic techniques may aid in the detection of disseminated sulphides (e.g. in porphyry systems) and intrusive contacts respectively.

All these parameters have to be assessed in terms of cost, time, and efficiency so as to be utilised to their fullest potential within the framework of the exploration programme. Exploration in both the Damara and Saldanian Provinces nècessitates confident yet careful, well executed methodology, plus patience and persistence. The only truly effective way of exploring for largely blind deposits is by an integrated approach, and prospecting companies have to be prepared to adopt this policy if they want the optimum long-term pay back on their initial, often substantial, outlays.

6.0 CONCLUSIONS

The most important conclusion of this work is that a careful analysis of the geotectonic setting of granitoids can provide a very good basis for exploration for Sn-W deposits. There are diagnostic geochemical differences between granitoids associated with greisen Sn-W deposits and porphyry Mo-Cu deposits, which can be attributed to the geotectonic environment within which granitoid genesis took place. In particular it has clearly been emphasised that Sn-W greisens can be sought in areas where there has been significant tectonic thickening, crustal melting and anatexis, and that these environments are specifically related to variable manifestations of ensialic orogeny and regionallised hot spot magmatism, as well as collision related tectonics. There appears a very good correlation of Sn-W mineralisation to S-type ilmenite series granitoids. In this respect the Damara and Saldanian Provinces in southern Africa have good potential for the discovery of significant ore bodies - especially the former, where preservation, exposure and erosion levels are favourable.

Mineralisation associated with S-type ilmenite series granitoids is manifest in many forms, of which those inherent to the greisen environment (exogreisen fissure veins, stockworks and replacements plus endogreisen veining and primary disseminations) are the most significant in terms of economic production. It is concluded that careful assessment using integrated exploration methods of granitoid terranes and indicator signs within and peripheral to intrusive contacts and roof zones (such as contact aureole hornfelsing, with superimposed greisenisation) are a viable means of initiating exploration for blind or only partially exposed mineralised systems.

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8.0 APPENDIX 1: Recognition of magnetite- and ilmenite-series granitoids in the field (extracted from Ishihara, 1981 p.480).

" Identification of the two series is easy in the field with the aid of a hand magnet and by observing heavy minerals concentrated on weathered surfaces. If a rock is identified as magnetic by the hand magnet, it generally belongs to the magnetite series. A magnetic susceptibility meter helps significantly in identification. Measurements cannot always be done on fresh outcrop. Care should be paid to pyritization of magnetite related to weak hydrothermal alteration; the pyritization is usually recognized by visual inspection. Hematitization of magnetite on weathered rocks can be ignored, because more than 50 percent of the magnetite remains in completely decomposed, loose granitoids. Complete hematitization has so far only been observed in hydrothermally altered rocks, which can be identified by the presence of altered silicate minerals.

Typically, the rock type can be recognized by the mineralogic characteristics of the ferromagnesian minerals. Biotite is most useful because it occurs universally. For ilmenite series rocks biotite has a reddish brown Z color in thin section and a greasy black color in hand specimens, whereas it looks more greenish in magnetite series rocks, with a Z color of greenish brown under the microscope. Primary muscovite, which constitutes muscovite-biotite granite, is found only in ilmenite series granitoids. Pyrrhotite is readily decomposed in air, so that ilmenite series rocks containing pyrrhotite can be readily identified by limonitic spots or irregular staining, even on fresh outcrop. If bulk chemical data are available before field work, the ferric/ferrous ratio is a useful adjunct to proper identification. "

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