

Fractional Crystallization and the Origin of Tin Deposits in Granitoids

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A comparison between tin-bearing granitoids in an anorogenic setting (Bushveld Complex) and an orogenic setting (Blue Tier Batholith, Tasmania) reveals a number of genetically important similarities. These include: *in situ* fractional crystallization characterised by marked decrease in Ba and Sr and increase in Rb; the accumulation of late melt in a sheet-like form near the roof zone; the association of barren pegmatites overlying the ore; and of aplites; and the occurrence of conformable tin-bearing sheets, often exhibiting greisenization. These features allow the formulation of the following genetic model. A crustally-derived granitoid magma is emplaced and undergoes fractional crystallization from the margins inwards, with bottom crystallization dominating. Disruption of earlier formed solids by rest liquid commonly occurs. Continued fractional crystallization causes enrichment in volatiles and incompatible elements in the late rest melts, which have a sheet-like habit. The efficiency of enrichment of incompatible elements is critically dependant on the degree of separation of melt from solids throughout crystallization. An early, tin-poor vapour may separate after initial water-saturation of the magma is achieved, and this collects under the roof, commonly forming an impermeable barrier to later tin-bearing fluids. Continued fractional crystallization on the floor further enriches incompatible elements, and at a very late stage a Sn-rich vapour separates within the intercumulus phase and becomes concentrated by progressive crystallization of the intercumulus melt. At a late stage of solidification, this vapour loses equilibrium with the earlier formed feldspars and greisenization ensues, accompanied by the crystallization of cassiterite and other ore minerals. The nature of the mineralization changes if through-going fractures tap the late fluids. This model predicts systematic changes in trace element geochemistry with crystallization which provide useful tools for assessing the tin potential of a granitoid, and for indicating the direction of crystallization of the magma, and hence the location of possible ore.

INTRODUCTION

Most of the primary tin deposits of the world occur in or adjacent to epizonal or mesozonal intrusive granitoids, although subvolcanic felsic intrusives and/or lavas may be important hosts to tin mineralization in more restricted areas

(e. g. Bolivia: Turneure, 1971; Sillitoe et al., 1975; and elsewhere Taylor, 1976). The majority of deposits are confined to Palaeozoic, Mesozoic or Tertiary orogenic belts (e. g. Itsikson, 1960), particularly around the Pacific margin (e. g. Hosking, 1970; Mitchell and Garson, 1972), but also to a lesser extent around

the Atlantic margin (e. g. Schuiling, 1967) and in central Europe (e. g. Tischendorf, 1973). However there are important deposits in anorogenic settings, including the ring complexes of Nigeria and the Bushveld Complex of South Africa (e. g. Hunter and Lenthall, 1971).

In order to define a general process for tin concentration in granitoid magmas, and consequent generation of tin deposits, it is important to study granitoids from both orogenic and anorogenic settings. This paper represents an attempt to define such a process by comparison of the host granitoids in a well documented orogenic tin district, the Blue Tier Batholith of eastern Tasmania (e. g. Groves, 1972a; Groves, 1977), with that in a well documented anorogenic tin district, the Bushveld Complex (e. g. Hunter and Lenthall, 1971).

THE BLUE TIER BATHOLITH

Regional Geology and Mineralization

The Blue Tier Batholith is one of a number of post-kinematic composite batholiths that occur in Tasmania at the southern end of the Tasman orogenic belt, and the contained tin field is one of several discrete zones of tin mineralization that are associated with Lower Devonian to Upper Permian batholiths containing epizonal granitoids within the orogenic belt (e. g. Solomon et al., 1972; Taylor, 1974). The geology of the batholith has been previously described by Gee and Groves (1971), Groves (1972a) and Groves (1977), the mineralization by Reid and Henderson (1928), Thomas (1953), Groves (1972b) and Groves and Taylor (1973) and its regional gravity by Leaman and Symonds (1975). The following brief precis of the regional geology of the batholith is from these studies.

The batholith of Upper Devonian age (McDougall and Leggo, 1965), is a composite tabular body, over 75 km long, over 45 km wide and approximately 12 km thick that was emplaced into essentially unmetamorphosed but folded sequences

of quartz greywackes and shales of Siluro-Devonian age (Fig. 1). Individual granitoid plutons have sharp, discordant contacts, narrow zones of thermal metamorphism, and appear to be epizonal, essentially passively emplaced, magmatic intrusions. Limited Rb/Sr isotope studies indicate high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (McDougall and Leggo, 1965) and suggest a possible crustal origin.

The earliest crystallized units are steep-sided plutons of hornblende- and biotite-bearing granodiorite containing abundant enclaves of diorite and country rocks. These granodiorite plutons appear to represent the embryonic margins of the batholith which were disrupted by later crystallized coarse-grained to porphyritic biotite adamellites that now occupy the greatest surface area of the batholith (Fig. 1). In the roof zone of these biotite adamellites there are relatively small, mappable units of muscovite- and biotite-bearing granites that are the last crystallized major units of the batholith. They appear to be sheet-like bodies with feeder dykes in places. The Little Mount Horror and Constables Creek Sheets (Fig. 1) are minor bodies that show small but distinctive differences to the more extensive Mount William, Mount Cameron and Lottah Sheets and the Mount Paris Mass (Groves, 1977). Of the latter bodies, the Mount William Sheet is at least in part the basal section of a sheet whereas the Lottah Sheets and the Mount Paris Mass represent the roof zone: the position of the Mount Cameron Sheets is less clear. All muscovite-biotite granites are even grained, reddish to yellow-brown rocks containing less than 5 % total mica and commonly with miarolytic cavities. They contain large flakes of apparently primary muscovite, and rocks from the upper parts of the sheets contain accessory, apparently primary topaz, tourmaline, fluorite and cassiterite. Biotites are typically Li- and F-rich but Cl-poor (Groves, 1977). The granites occupy less than 10 percent of the surface area of the batholith, and probably represent an even smaller volume (perhaps 2 % of the batholith;

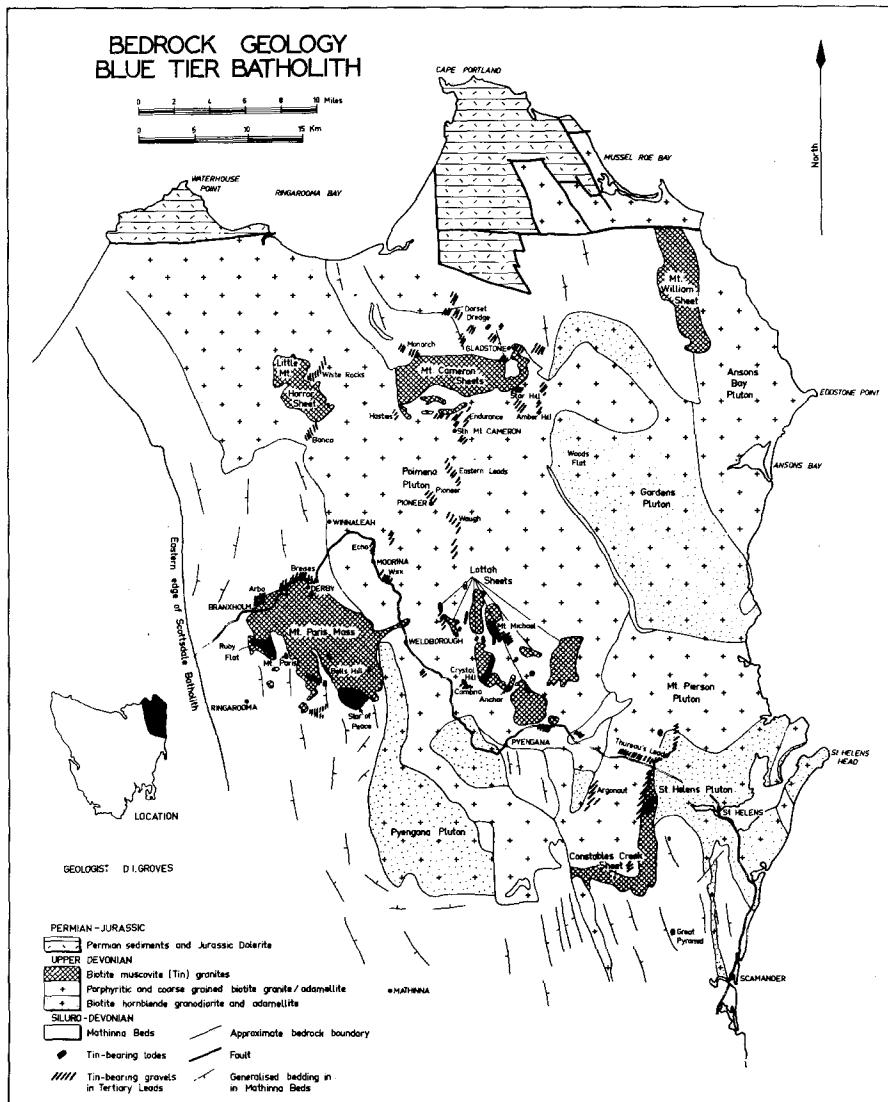


Fig. 1. Geology of the Blue Tier Batholith, showing occurrences of primary and alluvial tin.

Groves, 1972a; supported by gravity data; Leaman and Symonds, 1975).

Cassiterite, with lesser wolframite and molybdenite, mineralization is spatially related to structural irregularities in the roof zones of the muscovite-biotite granite sheets: there are no deposits associated with the base of the sheets. The most important types of mineralization

are flat-lying, greisenized granite sheets occurring below a thin pegmatite layer at the top of the granite sheets and below a roof zone of porphyritic biotite adamellite (e.g. Anchor Mine; Groves and Taylor, 1973). Other deposits include steeply-dipping greisen veins or cassiterite stockworks in the metasedimentary roof rocks immediately above the granite

sheets (e.g. Groves, 1972b). There is strong evidence that the sheet-like bodies of greisenized granite and at least some vein-like greisens are autometasomatic rather than post-magmatic in origin (Groves and Taylor, 1973).

Geochemistry

The bulk composition of the major components in the batholith (Table 1) varies from granodioritic to granitic but in general the variation amongst major and minor elements is relatively small (Groves, 1977). In contrast, the concentration of certain trace elements exhibit extreme variation in abundance within the rocks of the batholith: for example both Sr and Ba vary by factors of ca one hundred. There are well-defined inter-element correlations between Ba, Sr and Rb (Fig. 2), and these define two distinct compositional types: viz. the hornblende-bearing rocks and the biotite and biotite ± muscovite bearing rocks. McCarthy and Groves (in prep.) have shown that these interelement relationships are the result of in situ fractional crystallization, and that the two groups reflect a major change in liquidus mineralogy. The first group represents fractionation of hornblende, biotite, plagioclase, and minor quartz and the second crystallization of biotite, plagioclase, K-feldspar and quartz (± muscovite). The biotite-muscovite granites, with which mineralization is almost exclusively associated, represent crystallization of the final 2 percent or so of liquid following earlier fractional crystallization (Fig. 2): this is in excellent agreement with field estimates of the volume of these granites in the batholith (Groves, 1972a).

Throughout the geochemical study of the granitoids there has been no investigation of the spatial variation in geochemistry of any of the major components of the batholith that may indicate the overall spatial pattern of cooling and fractional crystallization of the batholith (cf. Wolhuter, 1973). However, the occurrence at the basal zone of a muscovite-biotite granite sheet at Mount William and the upper zones of similar gra-

nites at Mount Paris and Lottah does provide an indication of any vertical fractionation within the sheets, provided the sheets in each area represent crystallisation from similar residual liquids.

The major element compositions of the three granites are virtually identical (Table 1), but those from the roof zone are significantly enriched in Li, Sn, Rb and F and slightly depleted in Sr and Ba relative to those from the basal contact (Table 2): despite the relatively thin nature of the sheets, the variation is considerable. The zones of greisenized granite or greisens within the upper part of the Lottah granite sheets are even further enriched in Li, Sn, Rb and F and depleted in Ba (Groves and Taylor, 1973). These geochemical data, together with the occurrence of miarolytic cavities, are consistent with in situ crystallization of the sheets from the roof and floor with consequent water saturation and separation of a discrete water phase which rose and was impounded beneath the granite roof. The resultant geochemical pattern could have resulted from predominantly floor crystallization together with continuous reaction between intercumulus fluid phase and earlier-formed crystals and resulted in crystallization of accessory topaz, fluorite, tourmaline and cassiterite, greisenization of granite (with preservation of igneous textures) and even formation of greisens (with complete destruction of igneous textures) in places (c.f. Groves and Taylor, 1973).

BUSHVELD IGNEOUS COMPLEX Regional Setting

The emplacement of the Bushveld Igneous Complex occurred during the terminal stages of major Proterozoic intra-cratonic sedimentation represented by the Transvaal Supergroup. Although previously considered to be a lopolith, more recent gravity interpretations suggests that the Complex consists of a number of separate lobes (e.g. Willemse, 1969, Hunter, 1976). Nevertheless, the

Table 1. Mean composition of muscovite biotite granites and greisenized granites compared to granodiorites and adamellites of the Blue Tier Batholith

%	1	2	3	4	5	6	7	8	9
SiO ₂	69.4	73.5	77.8	75.7	76.75	77.4	75.7	75.0	75.0
TiO ₂	0.46	0.16	0.04	0.14	0.03	0.03	0.03	0.01	0.00
Al ₂ O ₃	13.85	14.0	12.1	13.1	12.95	13.0	13.5	13.6	13.5
Fe ₂ O ₃	0.8	0.3	0.3	0.25	0.4	0.35	0.25	0.33	0.42
FeO	3.4	1.4	0.6	0.9	0.5	0.6	1.1	1.05	1.75
MnO	0.07	0.02	0.02	0.04	0.02	0.03	0.05	0.04	0.07
MgO	1.55	0.4	0.06	0.16	0.01	0.01	0.05	0.06	0.09
CaO	3.36	1.0	0.6	0.8	0.35	0.5	0.4	0.4	0.55
Na ₂ O	2.4	3.0	2.9	3.1	3.85	3.5	3.3	3.3	2.7
K ₂ O	3.0	4.7	4.55	5.1	4.4	4.2	4.3	4.6	2.9
Li ₂ O	0.01	0.02	0.01	0.02	0.01	0.03	0.05	0.08	0.07
H ₂ O ⁺	1.2	1.3	0.3	0.3	0.4	0.45	0.65	0.88	1.7
H ₂ O ⁻	0.11	0.0	0.03	0.05	0.03	0.06	0.04	0.00	0.0
P ₂ O ₅	0.12	0.13	0.01	0.08	0.00	0.00	0.09	0.22	0.26
CO ₂	-	-	-	-	-	-	-	-	0.42
F	0.09	0.14	0.06	0.07	0.06	0.29	0.49	1.02	0.85
TOTAL	99.74	99.96	99.34	99.76	99.70	100.19	99.60	99.73	99.72
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ppm									
Ba	580		70	290	35	12	28		
Li	34	95	35	85	45	138	220	360	320
Rb	204	365	440	470	435	958	962	1035	1166
Sn	6	9	6	26	15	47	33	50	576
Sr	246	75	<5	36	7	6	4	5	6
W	<4	n.d.	<4	4	8	23	25	n.d.	n.d.
Zr	168	n.d.	75	100	75	70	50	n.d.	n.d.

1. Hornblende- and biotite-bearing granodiorites (6 analyses)
2. Porphyritic biotite adamellites of Poimena Pluton (5)
3. Constables Creek Sheet (2)
4. Little Mount Horror Sheet (2)
5. Mount William Sheet (6)
6. Mount Cameron Sheets (3)
7. Mount Paris Mass (2)
8. Lottah Sheets (6)
9. Greisenized granite of Lottah Sheets (4)

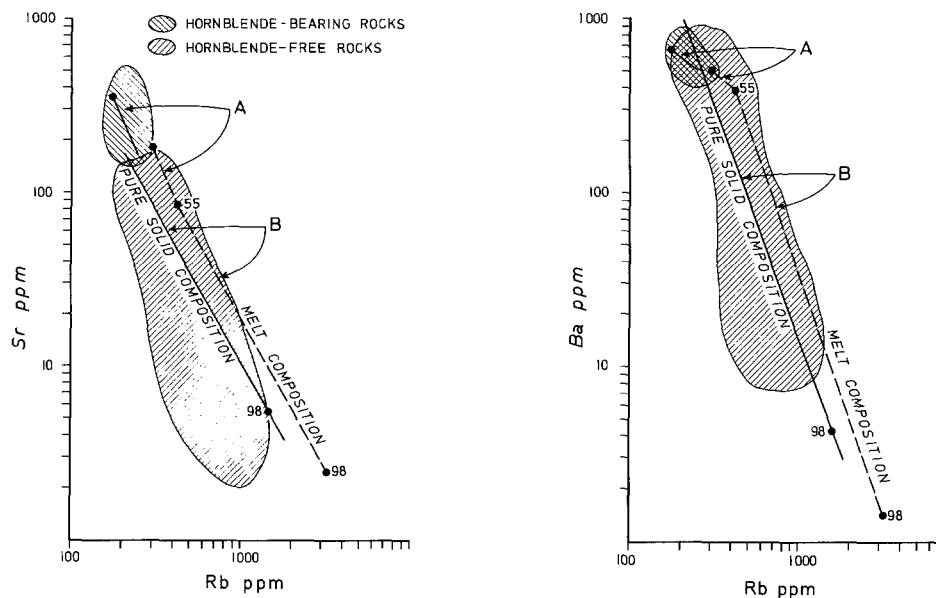


Fig. 2. Trace element geochemistry of the Blue Tier Batholith (fields of hornblende-bearing and hornblende-free cumulates shown) compared to theoretical fractionation trends based on Rayleigh law (McCarthy and Hasty, 1976; McCarthy and Groves, in prep.). Numbers on theoretical trends indicate percentage crystallization. A represents compositions formed by hornblende, biotite, plagioclase and quartz fractionation, and B biotite, plagioclase, K-feldspar, quartz ± muscovite fractionation

Table 2. Contrast in the concentration of certain elements in muscovite-biotite granites from floor and roof zones, Blue Tier Batholith

%	<u>Roof contact</u>	<u>Floor contact</u>
H ₂ O	0.5 - 1.0 (0.85)	0.3 - 0.5 (0.4)
F	0.1 - 1.4 (0.88)	0.01 - 0.09 (0.06)
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ppm		
ba	20 - 35 (28)	10 - 65 (35)
Li	100 - 500 (300)	20 - 65 (45)
Rb	755 - 1225 (1005)	395 - 475 (435)
Sn	25 - 66 (45)	8 - 25 (15)
Sr	4 - 7 (5)	<5 - 12 (<7)
K/Rb	30 - 58 (38)	74 - 95 (84)

rocks of the Complex are broadly conformable with the sedimentary sequences of the Transvaal Supergroup (Cousins, 1959) with only minor transgressive contacts (Willemse, 1969; Button, 1976; van der Merwe, 1976).

The precise emplacement processes for the Complex are not well understood (e.g. Vermaak, 1976; Hunter, 1976), although the extent of pre-emplacement deformation was confined to minor warping (e.g. Liebenberg, 1970; Vermaak, 1970). The Complex was emplaced in four distinct phases (Willemse, 1964): the sill phase, during which dolerite sheets were injected into the Upper Transvaal Supergroup; the epicrustal phase, represented by Rooiberg felsites, granophyre and interbedded sedimentary rocks; the main plutonic phase which gave rise to the layered sequence of mafic and ultramafic rocks; and finally, the late plutonic phase, represented mainly by the Bushveld granitoids discussed below.

Bushveld Granites and Associated Mineralization

The acid phase of the Complex (epicrustal and late plutonic phase rocks) lie stratigraphically above the layered mafic rocks and appears to be broadly sheet-like in form (Hunter, 1976). Isotopic evidence suggests that the late plutonic granitoids are of crustal origin (Davies et al., 1970).

In the Zaaiplaats area, immediately north of Potgietersrus, the granitoids form a thick, westerly-dipping sheet overlying the mafic rocks of the Complex, the two being separated by a sequence of metasedimentary rocks (Fig. 3). Within the granitoids, a variety of textural types are developed: a coarse-grained, mesocratic granite is developed at the base grading upwards into leucocratic granite, granophyric granite and granophyre, known collectively as the Main Granite, and finally into (Rooiberg) felsite (e.g. Strauss and Truter, 1944; de Waal, 1972). The Main Granite has been

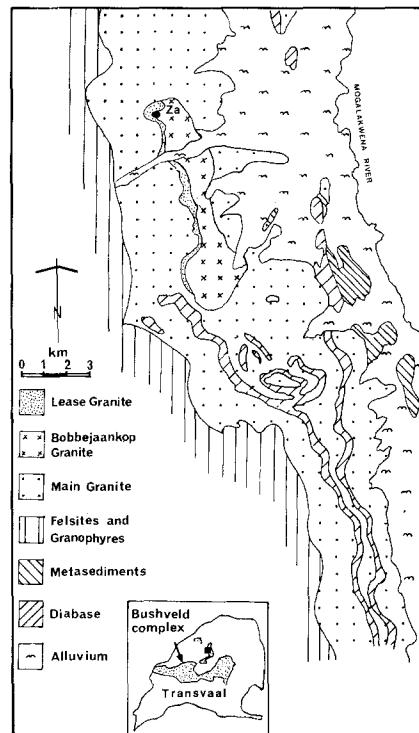


Fig. 3. The geology of the Bushveld Granites in the Zaaiplaats area (after Strauss and Truter, 1944)

intruded by a coarse-grained, red, miarolytic granite called the Bobbejaankop Granite whose upper contact is marked by an aplitic sheet (Lease Granite) overlain by a pegmatite zone. Both the Lease Granite and the pegmatite dip at ca 20°W conformably with the pseudostratification in the Main Granite (Fig. 3).

Tin has been mined in six areas of the Complex and the most economically important deposits are the exocontact deposits in sedimentary rocks of the Transvaal Supergroup although endocontact deposits are also important (Hunter, 1976). In the Zaaiplaats area, a variety of tin mineralization types occur in the upper portion of the Bobbejaankop Granite and in the Lease Granite (e.g. Strauss, 1954). Disseminated mineralization is developed in the Lease Granite, and especially in the Bobbejaankop Granite in which a zone

of low-grade ore occurs some 80 m below and conformable with the basal contact of the Lease Granite. Cassiterite occurs interstitially to quartz and feldspar and is associated with fluorite, commonly in vugs in chloritized and sericitized granite. More steeply-inclined, mineralogically zoned, pipe-like ore bodies are also developed in the upper portions of the Bobbejaankop Granite and in the Lease Granite, which are locally bright red in colour. The pipes appear to have been closed at both ends and are considered to have been formed by replacement processes including silicification, greisenization and tourmalinization, while the magma was still in a semi-solidified state as granite textures are locally preserved (Strauss, 1954). Where the pipes approach the upper contact of the Lease Granite against pegmatite, they tend to widen out into flat, lenticular bodies. The pegmatite is barren and evidently acted as a barrier to ore deposition. The granophytic granites above the pegmatite are only mineralized in areas where throughgoing fractures tapped the

fluids in the Lease and Bobbejaankop granites (Strauss, 1954).

Geochemistry

The Main and Bobbejaankop Granites are similar and extremely homogeneous in composition (Table 3), and lie close to the thermal minimum in the granite system (Rhodes, 1974). However, Ba and Sr exhibit considerable ranges in concentration, the former varying by a factor of 30 and the latter by a factor of 100 (Fig. 4). Furthermore, good interelement correlations exist between Ba, Sr and Rb (Fig. 4).

McCarthy and Hasty (1976) interpreted these Ba, Sr and Rb relationships as indicative of cumulate chemistry, with the Bobbejaankop Granite forming from the same magma as the Main Granite, but towards the end stages of crystallization. From the spatial distribution of Ba in these granites, McCarthy (1977) inferred that crystallization of the parent commenced along the roof and floor and proceeded inwards. Water became increas-

Table 3. Mean composition of the Main and Bobbejaankop granites, Bushveld Complex (from Fourie, 1969)

	Main Granite	Bobbejaankop Granite
SiO ₂	75.9	76.3
Al ₂ O ₃	11.8	12.0
Fe ₂ O ₃	2.0	1.5
FeO	1.1	1.1
MgO	0.2	0.1
CaO	0.6	1.1
Na ₂ O	2.8	3.0
N ₂ O	4.9	4.9
TiO ₂	0.1	0.0

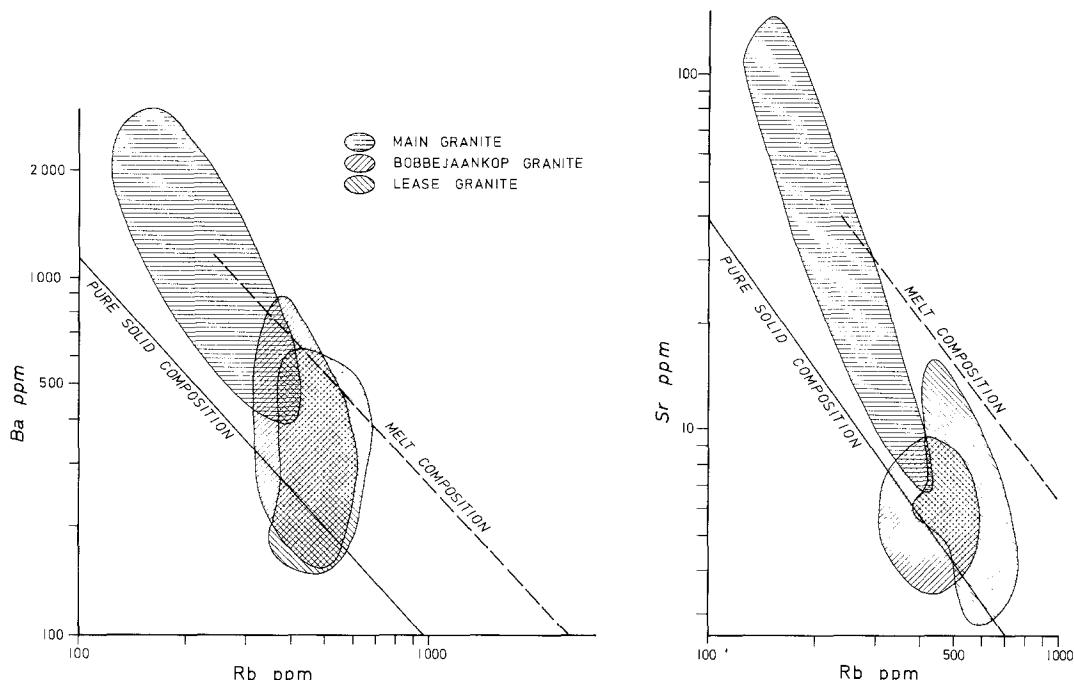


Fig. 4. Trace element geochemistry of the Bushveld Granites in the Zaaiplaats area (after McCarthy and Hasty, 1976)

ingly concentrated in the melt and at a very advanced stage of crystallization (in excess of 85 %), water saturation was attained and a discrete vapour phase separated and collected beneath the solidified roof. Only bottom crystallization occurred from this point onwards and the solids took on a red colour and possessed a miarolytic character as a result of entrapped intercumulus vapour. The very late-stage melt underwent quenching, forming the Lease Granite, and the vapour phase crystallized as a pegmatite sheet (McCarthy, 1977; McCarthy and Hasty, 1976).

GENERAL GENETIC MODEL FOR TIN GRANITES AND MINERALIZATION

The Blue Tier Batholith and Bushveld Granites occur in vastly different tectonic settings, and the nature of their mineralization differs in detail. However, they have several extremely significant features in common which enable

the formulation of a general genetic model for tin mineralization in specific granites.

Early Crystallization Mode

In both cases *in situ* fractionation has resulted in significant concentration of trace elements in residual liquids. The precise starting composition of the melt, and whether biotite (Bushveld) or hornblende and biotite (Blue Tier) are the early mafic cumulate phase(s) appear unimportant. The most significant single factor appears to be the degree of separation of residual liquid from cumulate phases which is reflected in the range in Rb and especially in Ba and Sr concentrations as discussed by McCarthy and Hasty (1976). It is noteworthy that both the Bushveld Granites and the Blue Tier Batholith show extreme ranges in concentration of these elements for major, mappable units of granitic rocks (Figs. 2 and 4): i.e. the highly fractionated end of the trace element range is not

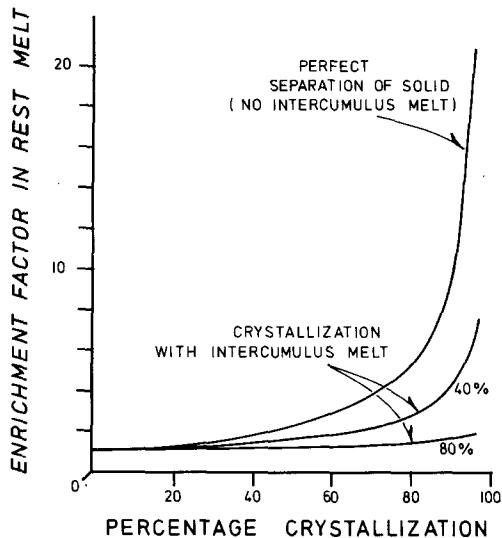


Fig. 5. The enrichment of a perfect incompatible element during fractional crystallization, as a function of the proportion of trapped intercumulus melt (after McCarthy and Hasty, 1976)

simply provided by volumetrically insignificant aplite or leucogranite dykes or sills. The dependence of enrichment of incompatible trace elements, including those of economic significance, on the purity of the cumulates is clearly shown in Figure 5. To obtain viable mineralisation during fractional crystallization it is necessary to have a low intercumulus content throughout the crystallization history.

The geochemical and textural evidence for the cumulate nature of the granitoids is discussed elsewhere (McCarthy and Hasty, 1976; McCarthy and Groves, in prep). Normally there is only fragmentary confirmatory field evidence of the cumulate character of these granites (e.g. cumulate layering in the Blue Tier Batholith). On the contrary, field evidence often suggests multiple intrusions of magma reflecting varying degrees of fractionation (e.g. Blue Tier Batholith and intrusive relationship between the Bobbejaankop and Main Granites of the Bushveld Complex), but the geochemical

evidence for a cumulate origin is so strong as to suggest that the correct interpretation is disruption of early cumulate granitoids by rest liquid. The phenomenon is evidently more extreme in orogenic settings (e.g. Blue Tier) than in anorogenic settings (e.g. Bushveld).

The tin deposits are associated with granites whose composition dictates that they crystallized from the last few percent of the rest liquid of fractional crystallization. A comparison of available published data on such granites elsewhere (Tischendorf, 1977) supports their limited geochemical range and highly fractionated nature. These granites invariably occur towards the roof zone of the batholith and therefore provide an important constraint on the spatial pattern of crystallization of the batholith. In the Bushveld granitoids, for example, McCarthy (1977) has shown that crystallization was initiated along the roof and floor and proceeded inwards. To obtain the present configuration of the granitoid mass, crystallization evidently proceeded more rapidly from the floor and the cumulates on the floor were purer (i.e. less intercumulus liquid) than those on the roof. This may have been due either to crystal settling or convective overturn which modified the heat distribution within the crystallizing batholith.

Crystallization of Final Rest Liquids

As a result of this crystallization mode, the last rest liquids tend to occur as sheets within crystallized granitoids. Prior to this phase, the magma was clearly water undersaturated as it had intruded to high levels (e.g. Harris, Kennedy and Scarf, 1970). Prolonged fractional crystallization leads to concentration of H_2O , and other volatiles in the rest liquid, and ultimately to water saturation in the ultimate sheets of rest liquid. At this point the nature of crystallization changes: a discrete H_2O -rich vapour separates (cf. Jahns and Burnham, 1969) and collects under the solidified roof as bottom crystallisation proceeds. The cumulates that form are commonly red or yellow-brown in

colour and are miarolitic due to entrapped intercumulus water vapour. They may contain cumulus cassiterite, topaz, fluorite or tourmaline at this stage. During this late stage of crystallization, large changes in concentration of incompatible elements occur in both melt and solid, for relatively small amounts of crystallization, producing a very pronounced vertical chemical zonation (see Fig. 5). The earliest vapours are apparently Sn-poor as in both the Bushveld and Blue Tier examples there are essentially barren pegmatites immediately overlying mineralization (see also Baily, 1977). However, there is a strong partition of Rb, Sn, Li, F and B into later vapour phase which rises to concentrate beneath the more pronounced structural highs in the crystallized or partly crystallized roof. Inward crystallization of intercumulus melt may confine the vapour phase into isolated pockets. The vapour phase reacts with partly consolidated cumulates to produce greisenized granite sheets and pipes which are interspersed with unaltered granitic rocks which still, however, contain a high concentration of the enriched incompatible elements. There is compelling evidence, at least at Blue Tier (Groves, 1977), that the Sn was derived directly from the vapour phase, not from leaching of crystallized biotite. The reaction between partial solid and the vapour phase may enhance the vertical geochemical zonation within the granite sheets, as shown by the Blue Tier examples: similar geochemical variation has also been recorded by Stempok and Sulcek (1969) from Cínovec and subhorizontal geochemical variation on a broad scale has also been demonstrated by Klominsky and Absolonova (1974) for the Karlovy Vary Massif, Czechoslovakia.

The two distinct vapour compositions which are formed during fractional crystallization appear to be separated in time by aplite formation, probably representing kinetic quenching of the last melt (McCarthy, 1977). This is evident from field relationships in the case of the Zaaiplaats deposit (see Fig. 3), and by the lack of extreme depletion in Sr and enrichment in Rb in the case of the

aplates from the Blue Tier Batholith (Groves, 1977). The Sn-poor, early vapour solidifies as a pegmatite sheet before the later, incompatible-enriched vapour starts to separate. Although detailed information pertaining to the final concentration stage of Sn-bearing vapour is not available, it seems probable that ultimate concentration of Sn into the vapour occurs during progressive, inward crystallization of the intercumulus melt, when vapour enriched sheets and pipe-like zones form, leading to extensive alteration of the enclosed cumulus minerals. Because of the presence of F and other incompatible elements, these late vapours solidify at considerably lower temperatures than the earlier, incompatible-poor vapour (Baily, 1977).

Form of Mineralization

The precise form of mineralization, both endocontact and exocontact, adjacent to the roof depends on the permeability of the immediate roof zone and in particular whether fractures are developed in the crystallizing magma at an early stage. In the Blue Tier examples, the vapour phase appears to be impounded beneath an impermeable roof zone to form subhorizontal greisenized granite and lesser greisen sheets where the tin-bearing granite is enclosed within earlier crystallized granitoids, and this appears also to be the case at Zaaiplaats: presumably the crystallized granitoids above the roof zone of the mineralized granite are still relatively hot so that discrete cooling joints are less likely to occur within the latter.

Where the mineralized granite locally forms the immediate roof of the batholith as for the Mount Paris Mass of the Blue Tier Batholith (Fig. 1), fractures were developed at an earlier stage due to heat transfer to the cooler country rocks and discrete sub-vertical greisen bodies occur along prominent fracture directions: where such bodies are relatively small, distinct lateral mineralogical zonations occur about a centrally disposed fracture. Where discrete fractures occur in the

country rocks, the vapour phase may escape from the crystallizing magma to produce exocontact mineralized quartz veins or pipes: much of the quartz and possibly sulphides which occur commonly in these veins may be derived from leaching of country rocks by the hot, upward-rising vapour phase. Less commonly the vapours may react with carbonate horizons to produce important replacement deposits as at Mt. Bischoff, Renison Bell and Cleveland in western Tasmania (e. g. Groves et al., 1972).

Summary Model

The essentials of this model are summarized in Figure 6. An initially water-undersaturated granitoid magma intrudes into a high structural level, and fractional crystallization ensues: initial liquidus mineralogy is unimportant. The rest magma is continually disrupting earlier cumulates producing an apparent sequence of emplacement of generally progressively more fractionated cumulate rocks. The efficiency of concentration of incompatible elements is controlled by the purity of the cumulates throughout the crystallization history. Fractional crystallization ultimately produces a water-saturated melt which invariably is located towards the roof zone of the intrusive. While crystallization continues on the floor, vapour bubbles collect in structural high points under the roof. Cumulates which form are now generally reddish in colour, and distinctly miarolitic. At a very late stage of crystallization, quenching of rest melt occurs, producing aplites, and the vapour solidifies as barren pegmatite. Continued crystallization of intercumulus liquid concentrates incompatibles further, and Sn and other elements become more strongly partitioned into the vapour, causing marked depression of solidification temperature. This vapour ultimately loses equilibrium with higher temperature cumulus minerals, and greisenization commences. Dependent on the permeability of the roof, various styles of mineralization may develop.

CONCLUSIONS

It is evident from this comparison of the Blue Tier Batholith and Bushveld Granite that tectonic setting is not of major consequence for the generation of Sn mineralization: of overwhelming importance is the style of crystallization of the host magma, as exemplified by Figure 6. Leaching of Sn from ferromagnesian minerals is not responsible for Sn mineralization in the examples studied. Localization of the mineralization is a continuation of the crystallization style and the permeability of the roof zone. The probability of large, low-grade deposits is strongly enhanced by an impermeable roof, best formed by earlier crystallized, relatively hot, cumulates.

The fact that some greisenized granite and greisen sheets and pipes represent closed systems within the host granite, with no significant alteration of overlying earlier crystallized granitoids, places constraints on the nature of transport of Sn in the vapour. Because the vapour is confined to the space between cumulus minerals, it is clear that the concentration of Sn in the vapour is at least a factor of 2 greater than the grade of the disseminated on which it produced. The fact that F and B are not associated with all Sn deposits suggests that complexes of these elements with Sn are not universally important in the transport of Sn. We suggest that the most likely universal form of transport is as anionic oxy- or hydroxycomplexes. The lack of chlorine-bearing minerals, and the absence of cavities which could be ascribed to halite and sylvite, would argue against the importance of concentrated chloride brines as transportation media in these particular instances.

Available Sr isotopic data for granitoid masses of which the tin-bearing granites form an integral part from the Bushveld (Davies et al., 1970) and eastern Australia (e. g. McDougall and Leggo, 1965; Brooks and Compston, 1965; Brooks, 1966; Richards et al., 1966; Webb and McDougall, 1968) suggest that the host granites are crustally derived, perhaps

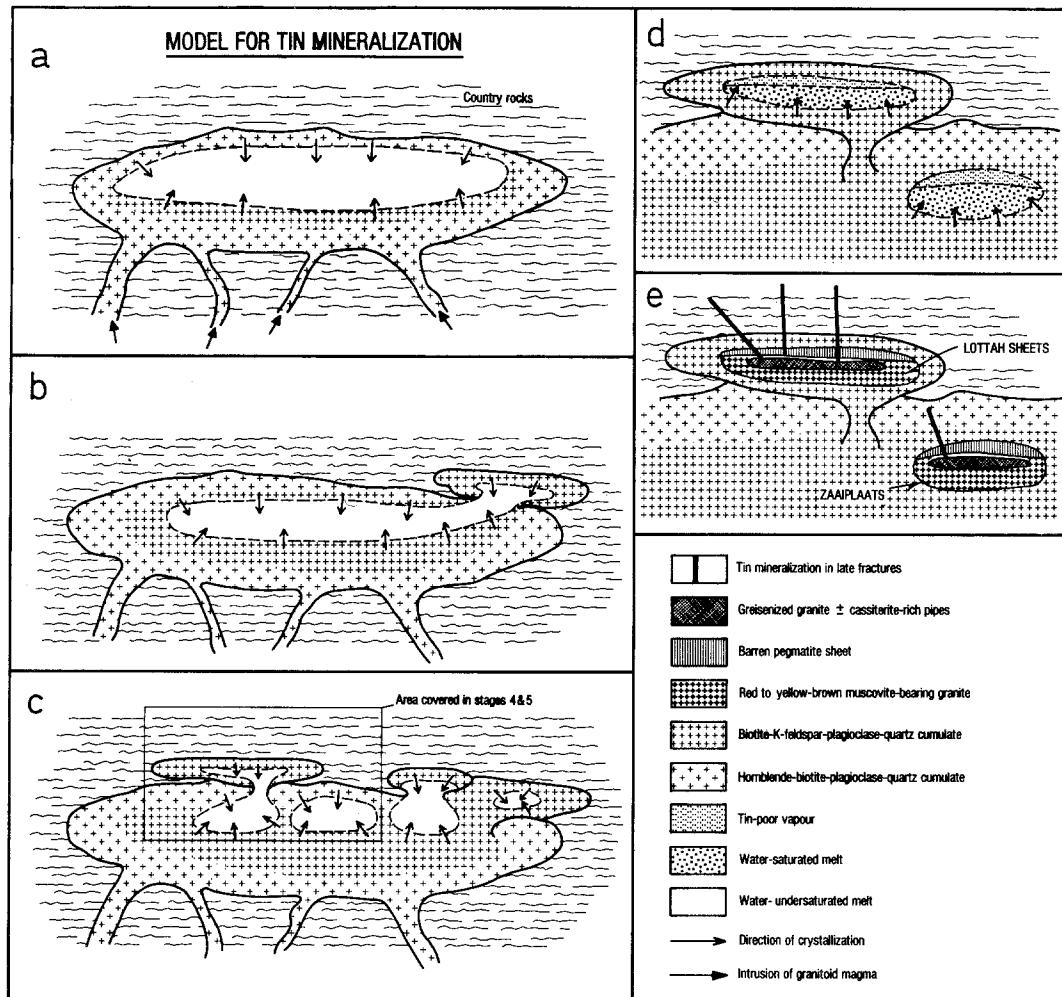


Fig. 6a-e. Proposed model for tin mineralization: vertical scale exaggerated.

a. Initial crystallization of granitoid magma intruded with hornblende, biotite, plagioclase and quartz on liquidus (need not be developed). b. Continued crystallization of magma with progressive change in cumulus mineralogy; disruption of early crystallizates by rest melt forming hornblende-free body. c. Further disruption of early crystallizates during continuing crystallization. d. Continuing crystallization with rest melt attaining water saturation; no roof crystallization but early separation of a tin-poor vapour that collects in structural highs under crystallized roof and forms barren pegmatite sheet at d/e. e. Water-saturated melt crystallizes red or yellow-brown, miarolitic muscovite-bearing granite; Sn-, F- and/or B-rich vapours in late intercumulus liquid cause greisenization beneath roof zone; cassiterite-rich pipes may develop adjacent to greisenized granite; mineralization may occur along late fractures

implying that a two stage model is required for the genesis of Sn deposits. Initial Sn concentrations in magmas derived directly from the mantle, or from melting of oceanic crust are probably not sufficiently high to allow generation of Sn deposits by the mechanisms proposed here. Rather, melting of pre-existing sialic crust, in which Sn is already enriched (e.g. Hamaguchi and Kuroda, 1970) evidently provides the source for Sn deposits. This is supported by the location of Sn deposits well into the continental portion of an active plate margin (e.g. Mitchell and Garson, 1972); in the eastern Australian examples, the zones of Sn mineralization occur adjacent to continental crustal blocks, or inferred blocks (Solomon et al., 1972).

The model proposed here for the genesis of Sn deposits suggests some useful exploration indicators. The economic potential of a granitoid can readily be assessed by examining the range in concentration of Sr and Ba in coarser grained varieties (excluding aplites): the greater the concentration range, the more likely is the presence of economic tin deposits. Furthermore, Sr and Ba are potentially useful pathfinder elements in the search for areas where Sn is likely to have been concentrated, since these areas will be characterized by low Sr and Ba contents. Confirmation of such an area as being a region where late stage crystallization has occurred can be obtained by examining concentrations changes of incompatible elements, such as Li and Rb, with height in the granite.

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