CHAPTER VI

INTERPRETATION AND DISCUSSION

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Mineralogically, the Takua Pit Thong cassiterite-sulfide (magnetite) deposit can be classified as a complex tin ore (Hosking, 1970) similar to the cassiterite-sulfide mineralization at Renison Bell in Western Tasmania, Australia in terms of geologic setting and mineralogy (Patterson et al., 1981). Because of its complex mineral assemblage, it is possible to deduce the fugacities of oxygen and sulfur during the ore formation if the temperature of formation of those minerals and thermodynamic data of the relevant minerals and aqueous species are accurately known.

Fugacities of Oxygen and Hydrogen Sulfide

It is rather unfortunate that no temperature data on the mineralization at the Takua Pit Thong or nearby deposits are available or have been carried out in this study (i.e., fluid inclusion work on filling and freezing temperatures). Nevertheless, some temperature information are available from several tin deposits in Thailand or Southeast Asian tin belt and from some similar types of tin deposits in other parts of the world. For example, Jackson and Helgeson (1985a) have estimated the temperature and depth of Southeast Asian tin deposits to be at ∼ 350 ^{o}C and 500 bars in the presence of an electrolyte solution containing ∼ 1 molal (m) total chloride from the fluid inclusion filling and freezing temperature on quartz-cassiterite deposit at the Samoeng mine in Chiangmai, the Saha Kit mine in Phuket, the Thung Kamin mine in Songkhla and Tong Seng mine of the Kinta Valley in Malaysia. Somewhat wider range of 500 - 250 ^{o}C and 0 - 1 molal NaCl fluid has been estimated from fluid inclusion study of minerals in cassiterite-wolframite-quartz-hematite veins at the Yod Nam mine in Nakorn Sri Thamarat (Norman and Trangcotchasan, 1982). Similarly, the temperatures of about 350 ^{o}C to 300 ^{o}C have been inferred for the ore-forming fluids from fluid inclusion, mineralogical and stable isotope studies on the cassiterite-sulfide mineralization at Renison Bell, Western Tasmania (Patterson et al., 1981). Therefore, as a rough approximation, it might be assumed that the temperature of formation of minerals at the Takua Pit Thong mine could have been below approximately 350 ^{o}C with 1 molal NaCl solution.

Phase relations in the systems $SnO - H_2S - H_2O - O_2$ and $FeO - H_2S - H_2O - O_2$ plotted as a function of fugacity of O_2 and H_2S are available (Jackson and Helgeson, 1985a) and could be directly applicable to the mineral assemblages found in the Takua Pit Thong mine (Figure 56). Based on textural evidence, because magnetite precipitated slightly preceding and/or contemporaneous with cassiterite in the sheared (feeder) zone at the No.3 orebody as well as the No.1 orebody. The f_{O_2} and f_{H_2S} condition of the ore solution responsible for precipitation of this mineral assemblage should be within the magnetite stability field (Figure 56). However, owing to the fact that some magnetite crystals contain small blebs of pyrite and pyrrhotite (suggesting that pyrite and pyrrhotite were precipitated slightly preceding and/or contemporaneous with magnetite), the condition of the ore solution in the phase diagram was likely to be very

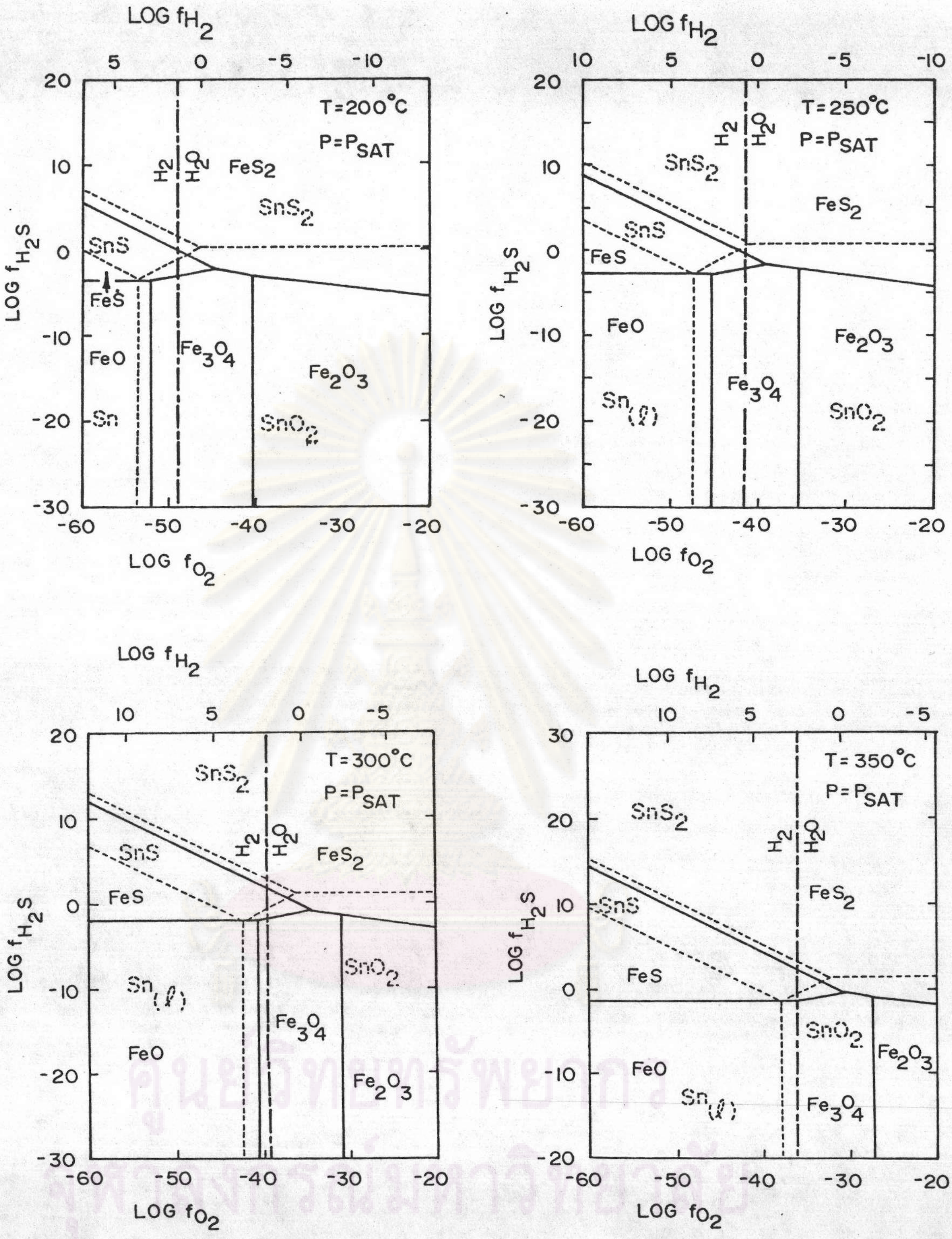


Figure 56 Superimposed phase relations in the system $SnO\text{-}H_2S\text{-}H_2O\text{-}O_2$ and $FeO\text{-}H_2S\text{-}H_2O\text{-}O_2$ as the function of the fugacities of O_2 and H_2S for unit activity of H_2O at 500 bars and 200°, 250°, 300° and 350 °C. The solid curves designate boundaries between the stability fields of minerals in the system $FeO\text{-}H_2S\text{-}O_2$ and the dashed curves represent boundaries between the stability field of minerals in the system $SnO\text{-}H_2S\text{-}O_2$. The long-dashed curve corresponds to the lower boundary of the stability field of H_2O (after Jackson and Helgeson, 1985).

close to magnetite-pyrite-pyrrhotite triple point. As the ore solution ascended to the granite/country rocks contact to precipiteate cassiterite with being followed by pyrrhotite, pyrite and chalcopyrite, the condition of ore solution might have been shifted from the magnetite-pyrite-pyrrhotite triple point to pyrrhotite-pyrite boundary but still remained within cassiterite stability field (Figure 56). In other words, it may be assumed that the overall condition of the ore deposition at the Takua Pit Thong mine was approximately close to the magnetite-pyrite-pyrrhotite buffer. This assumption is also supported by the fact that cassiterite and magnetite crystals do not show any sign of corrosion or resorption that would suggest the non-equilibrium condition during the pyrrhotite, pyrite and chalcopyrite precipitation. The estimation of the Takua Pit Thong ore-forming condition at the magnetite-pyrite-pyrrhotite buffer is exactly the same as those prevailed during cassiterite-sulfide mineralization at Renison Bell, Australia (Patterson et al., 1981) and those responsible for the Southeast Asian tin mineralization (Jackson and Helgeson, 1985 a). As a consequence, the f_{O_2} condition of the Takua Pit Thong ore solution could have been $\sim 10^{-31}$ at 350 ^{o}C, $\sim 10^{-35}$ at 300 ^{o}C, $\sim 10^{-39}$ at 250 ^{o}C or $\sim 10^{-45}$ at 200 ^{o}C with the f_{H_2S} haveing been 10^{0} at 350 ^{o}C, 10^{-1} at 300 ^{o}C, 10^{-2} at 250 ^{o}C or 10^{-3} at 200 ^{o}C.

Color Variation of Biotite, Phlogopite and Phengite

Color variation of biotite has long been known relating to its composition (Grout 1924; Tilley, 1926; Harker, 1932; Hall, 1941 and Hayama, 1959). According to Hayama (1959), the color of biotite depended largely on its TiO_2 content and $Fe_2O_3/(Fe_2O_3 + FeO)$ ratio.

Principally, if the $Fe_2O_3/(Fe_2O_3 + FeO)$ was less than about 0.2, the color of biotite would be controlled by its TiO_2 content. That was the color was becoming progressively darker in shades of reddish-brown as increasing TiO_2 content. In conrast, biotite with relatively low TiO_2 content (< 2 wt%) would display yellowish to greenish-brown to green as the $Fe_2O_3/(Fe_2O_3 + FeO)$ increasing from ∼ 0.2 to 0.6. However, biotite with intermediate contents of both TiO_2 and $Fe_2O_3/(Fe_2O_3 + FeO)$ ratio had a tendency to show its color intermediarily. Nevertheless, in general it appeared that in many cases TiO_2 and Fe_2O_3 would not be both enriched in the same biotite sample.

Guidotti (1984) argued that the Hayama's suggestion could be valid mainly on the biotite that occurred in common pelitic metamorphic rocks. He further suggested that the biotite colors could be used as an indicator of what sort of opaque minerals to be expected as investigation the biotite containing pelitic rocks petrographically. He added that shades of red in the biotite might indicate low Fe_2O_3 contents and ilmenite rather than magnetite or hematite was likely to occur, if any, as opaque oxide. In contrast, shades of greenish-brown was more likely to encounter magnetite and/or hematite and, of course, Fe^{3+} might form a significant fraction of the total iron in biotite. It is unfortunate, however, that no literature is available concerning the color variation of phlogopite and phengite.

The color variation is observed in biotite, phlogopite and phengite from the Takua Pit Thong deposit. As a consequence, it would be interesting to know whether the variation in their colors are really related to the difference in their chemical constituents (i.e., Fe_2O_3, FeO, TiO_2, MgO etc.) as being outlined above by direct chemical analysis, eventhough the occurrence of those micas was probably connected to hydrothermal processes. The chemical analysis would also reveal that the so called "biotite, phlogopite and phengite" are chemically correct or not. Unfortunately, however, such the chemical analysis has not been carried out in present study yet. Nevertheless, using the mineral assemblages outlined earlier, it is possible, at least as a starter, to anticipate the chemical composition of those micas and their relation to the chemistry of the ore solution.

Green coloration of biotite and decoloration of phlogopite and/or biotite to phengite were likely to relate to the ore-forming processes. This argument can be supported by the fact that green biotite and colorless to pale green phengite are the most common gangue minerals found associated within and nearby the orebodies. Also by the fact that green coloration or decoloration was developed inward in a single biotite or phengite crystals and relatively more intense toward an open vug which might have been channelways of the ore-forming fluid.

Of particular note is that magnetite usually associates with green biotite in open vug. The formation of green biotite in the orebodies probably took place shortly before and/or contemporaneous with magnetite and cassiterite. The green coloration of biotite also appears to be more intense in the No.3 orebody where magnetite and cassiterite are the major ore minerals with sulfides as being only trace minerals. The No.3 orebody is believed to represent a feeder zone of the ore solution. All textural evidences seem to point to the same conclusion that the green coloration of biotite should have a direct relation to the occurrence of magnetite. In other words, prior to and/or contemporaneous with the precipitation of magnetite, the ascending ore solution should have a rather high Fe^{3+}/Fe^{2+} ratio because magnetite precipitation requires two Fe^{3+} and one Fe^{2+} ions (i.e.,physicochemical condition within magnetite stability field). The Fe^{3+} in the ore solution might have been introduced into the original brown biotite in the wall rocks and eventually converted its color to greenish brown and to green as the Fe^{3+} contents in the biotite increased.

The mechanism of how the Fe^{3+} ion could enter into the biotite structure is unknown at the hydrothermal condition. Experimental studies on the substitution of Fe^{3+} in biotite are available mainly at high temperature and pressure (e.g.,400 - 800 °C and 1 - 2 kilobars). For example, Wone and Eugster (1965) suggested on the basis of analyses of natural biotites and their experimental results, that much of the Fe^{3+} in biotite could be accounted for by the ferriannite ($Al^{iv} = Fe^{iv + 3}$) and oxyannite ($Fe^{2+} + OH^{-} = Fe^{3+} + O^{2-}$) substitutions and that the Fe^{3+} contents (i.e., oxyannite content) of annite increased with f_{O_2}. Ferrow and Annersten (1984) and Partin et al., (1983) similarly found a relation between Fe^{3+} and f_{O_2} by which at conditions more oxidizing the Fe^{3+} content increased proportionally. Ferrow and Annersten (1984) further concluded that the Fe^{3+} substitution in annite occured by the oxyannite mechanism and by the creation of interlayer vacancies. Eventhough these experimental data were performed at high temperature and pressure conditions, they seem to agree fairly well with present observation that the green coloration of biotite, apparently reflecting a relatively high contents of Fe^{3+}, took place in a high f_{O_2} condition in which magnetite was stable. The substitution of Fe^{3+} in the brown biotite to form the green biotite could be by the mechanism similar to any one of them or combination of the three (oxyannite, ferriannite and interlayer vacancies).

The mechanism of decoloration of phlogopite or biotite to form phengite in the orebodies is less known. The ore solution as it ascended closer to the marble and calcsilicate hornfels contact might have been depleted in Fe^{3+}/Fe^{2+} ratio (comparatively more reducing condition) because large quantity of Fe^{3+} could have been taken up by the formation of green biotite and magnetite especially in the feeder zone. The Fe^{2+} as well as other neccessary constituents in the solution might enter into the original pale brown phlogopite and/or brown biotite with a proper substitution or rearrangement and could possibly be accounted for the transformation to colorless or pale green phengite.

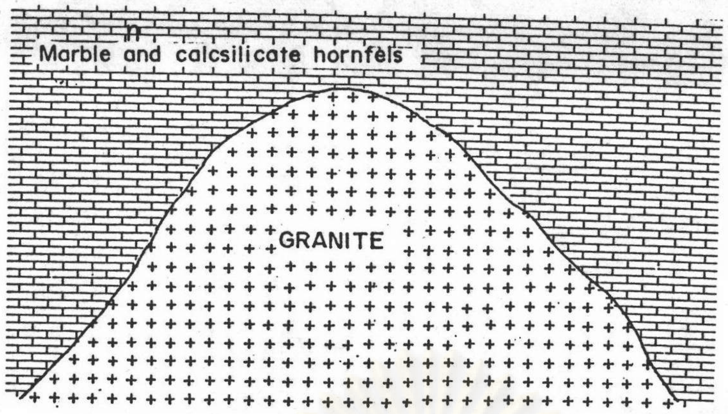
If this was indeed happened it could be explained why the green biotite is abundant in the orebodies toward the granitic contact especially the No.3 orebody which contains only the green variety of biotite whereas the colorless or pale green phengite is enriched toward the marble and calcsilicate hornfels contact.

Generalized Model for the Mineralization at the Takua Pit Thong mine

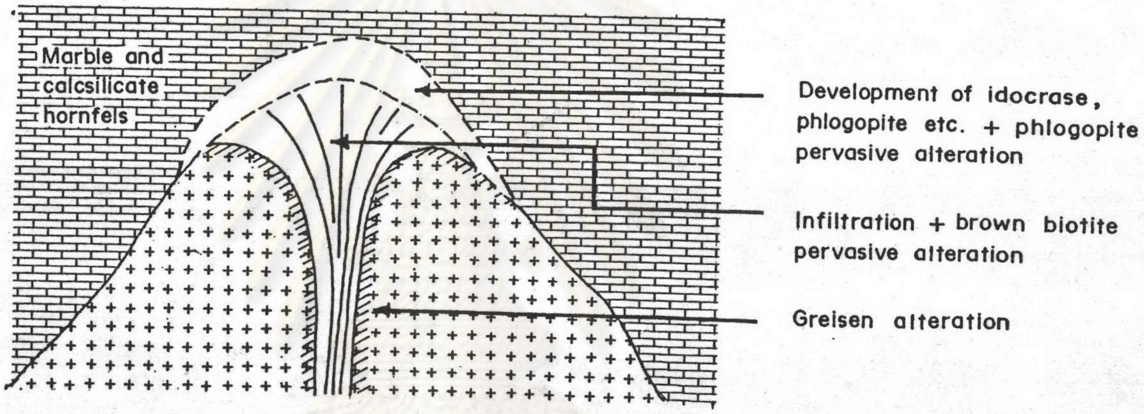
The data in the preceding sections can be combined into a generalized model for the mineralization at the Takua Pit Thong mine and depicted in Figure 57 and Table 3.

In Upper Cretaceous, various phases of strongly differentiated Takua Pit Thong-Khao Daen granite, pegmatite and aplite intruded into presumably Ordovician dolomitic limestone and clastic sequences of the Khao Noi Sethi Formation in the Thung Song Group and thermally metamorphosed the sediments into marble interbedded with calcsilicate hornfels and metamorphosed sandstone of hornblende hornfels facies. Following the granitic intrusions and crystallization, aqueous phase of probably the same magmatic origin might have been migrated along a certain weak zone to the upper parts of the granitic pluton and caused intensely pervasive greisen alteration within the granitic body. Contemporaneous with or subsequent to the greisenization, at or near the contact zone, by addition of Ca and Mg from the country rocks, the ascending fluid might have caused the development of metasomatic infiltration both in country rocks and granitic rocks as well as subsequent intense brown biotite and phlogopite pervasive alteration. The brown biotite and phlogopite rather than muscovite were formed at or near the granitic/wall rocks contact because of high contents of Mg and possibly Fe in the aqueous phase. Furthermore, the brown biotite formed during the metasomatic infiltration and pervasive alteration was likely to be of Mg-rich type. Phlogopite tended to be enriched toward the marble and calcsilicate hornfels side because of the high activity of Mg^{2+} in the solution toward its source.

I Intrusion and contact metamorphism



II Metasomatic infiltration



III Mineralization

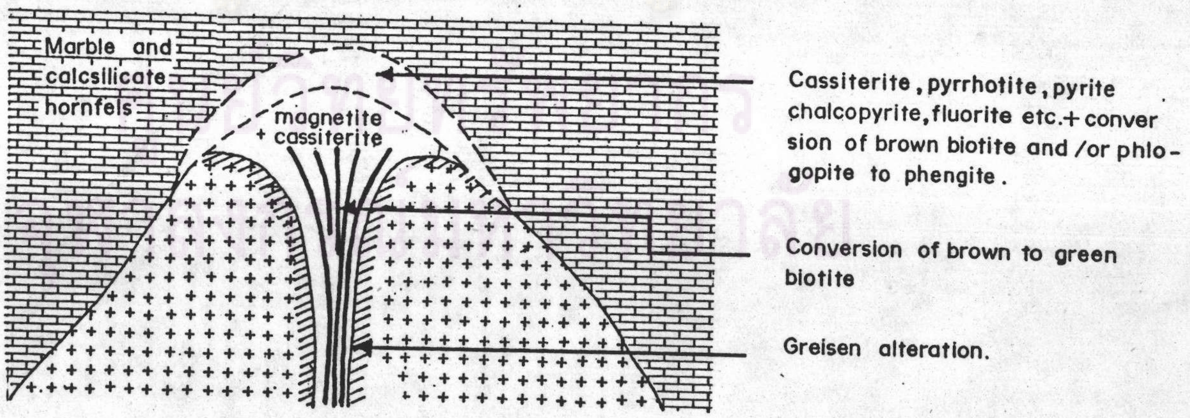


Figure 57 Schematic diagrams showing the sequences of events taken place at the Takua Pit Thong mine.

Table 3 Summary of the minerals occurred during the various events at the Takua Pit Thong mine.

Early Late

I Contact metamorphism	II Metasomatic infiltration	III Mineralization
calcite (recrystallized)	phlogopite	green biotite
phlogopite	brucite	phengite
diopside	idocrase	magnetite
tremolite-actinolite	wollastonite	cassiterite
clinozoisite	scapolite	garnet
sphene	calcite	pyrrhotite
plagioclase	epidote	chalcopyrite
	tremolite-actinolite	pyrite
	sphene	arsenopyrite
	fluorite	sphalerite
	tourmaline	epidote
	muscovite	fluorite
	quartz	quartz
	plagioclase	
	K-feldspar	
	brown biotite	
	green hornblende	

Tectonic movement was followed in approximately north-south direction and provided as additional channelways for subsequent ore-forming fluid which was probably high in Fe^{3+}/Fe^{2+} ratio to ascend. This ore-forming solution as ascending through the feeder zone, might be responsible for the color conversion from brown to green in the biotite by Fe^{3+} substitution and, at about the same time or slightly later, precipitated magnetite and cassiterite while the physicochemical condition was in the magnetite stability field (relatively more oxidizing condition). The condition of the same fluid, as leaving the feeder zone into the granite/country rock contact, might have been progressively changed through green biotite wall rock alteration and magnetite precipitation into pyrrhotite-pyrite stability boundary (relatively more reducing condition). This evolved ore solution which relatively depleted in Fe^{3+}/Fe^{2+} ratio might be responsible for transformation of phlogopite and/or brown biotite to colorless or pale green phengite. Contemporaneous with and/or subsequent to the phengite transformation, the same ore fluid, with addition of Ca from the country rocks, would precipitate cassiterite with being followed by pyrrhotite, pyrite, chalcopyrite, arsenopyrite, sphalerite (rather than magnetite) and fluorite.

This could explain why magnetite and cassiterite tended to precipitate earlier and enriched toward the feeder zone as well as associated with only green biotite whereas most of the sulfide minerals (plus cassiterite) and, in particular, fluorite were progressively enriched toward the granite/country rocks contact and associated with both green biotite and phengite. The depositional mechanism of these minerals especially cassiterite could be the increasing pH resulting from reaction with marble or wall rock alteration or decreasing temperture or combination of the two (Jackson and Helgeson, 1985 a, b).

Another alternative explanation for the observed sequence of mineralization is that the ore solution responsible for precipitation of magnetite, cassiterite and quartz was entirely different both in time and its physicochemical condition from those responsible for precipitation of sulfides and fluorite. This alternative is however less likely because if the ore solutions came at different time, it should be expected that magnetite and pyrrhotite to be deposited more evenly in terms of their quantities both in feeder zone and at the granite/country rock contact.

Ore Beneficiation

Most of cassiterite forms as large euhedral to subhedral crystal with the average grain size of 2 millimeters. Inclusions of foreign component in cassiterite is rarely found. Intergrown of cassiterite with magnetite and quartz is normally coarse crystalline. These textural evidences reveal that to separate cassiterite from other associated minerals does not require an overgrinding in order to save time and money. However, abundant micas in the ore may pose some problems during the grinding because of their platy property. Furthermore, mineralogy outlined earlier should also serve as a helpful data in mineral dressing.