

Contrasting origins of Sn-W mineralization in western Thailand

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December, 1991

**A thesis submitted to the Faculty of Graduate Studies and Research in partial
fulfillment of the requirements of the degree of Doctor of Philosophy**

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ABSTRACT

Investigations of pegmatite, stockwork and replacement deposits, using geological, whole-rock and mineral chemistry, fluid inclusion, and stable isotope data, were undertaken to establish some of the controls of mineralization in the western Thai Sn-W belt.

Cassiterite was saturated in the late-stage melt in the Nong Sua granitic pegmatite and subsequently with wolframite in orthomagmatic fluids. The latter mixed with external aqueous and aqueous-carbonic fluids of evolved meteoric or metamorphic origin at $\leq 650^{\circ}\text{C}$ and ≤ 3.8 kbar. Hydrothermal cassiterite and wolframite crystallized as a result of increasing $f\text{O}_2$, fluid mixing, and cooling. At the Pilok stockwork deposits tourmalinization was followed by successive wolframite-K-feldspar deposition and cassiterite-sulphide greisen mineralization. Tin was remobilized by reduced metamorphic or evolved magmatic-meteoric aqueous-carbonic fluids. Cassiterite was deposited at $300\text{--}450^{\circ}\text{C}$ in response to increasing $f\text{O}_2$, pH, or decreasing temperature. Limited data on the Takua Pit Thong replacement deposit suggest that cassiterite was deposited from aqueous fluids at $\geq 500^{\circ}\text{C}$, but metamorphic aqueous-carbonic fluids were also involved.

The depth of emplacement was one of the major controls on mineralization style. Aqueous-carbonic fluids in the western Thai, and other Sn-W belts, are a consequence of a moderately deep level of emplacement; hypersaline orthomagmatic fluids are associated with sub-volcanic Sn-W mineralization.

RESUME

L'étude de gisements de type pegmatitique, de type stockwerk et de type remplacement a été réalisée, à l'aide de données géologiques, d'analyses chimiques de roche totale et de minéraux, d'inclusions fluides et d'isotopes stables, en vue de déterminer certains processus qui contrôlent la minéralisation dans la partie occidentale de la ceinture de gisements Sn-W de Thaïlande.

La cassitérite a atteint la saturation à un stade tardif de l'évolution du magma granitique à la pegmatite Nong Sua et, subséquemment, avec la wolframite dans les fluides orthomagmatiques. Ces fluides se sont mélangés avec des fluides externes aqueux et carbonique-aqueux d'origine météorique ou métamorphique à des températures $\leq 650^{\circ}\text{C}$ et à une pression $\leq 3.8\text{kbar}$. La cassitérite et la wolframite hydrothermales ont cristallisé suite à une augmentation de la $f\text{O}_2$, au mélange de fluides et au refroidissement. Dans les stockwerks de Pilok la tourmalinisation a été suivie par la précipitation de wolframite-félspath potassique et ensuite par la minéralisation de cassitérite-sulfure de type greisen. L'étain a été remobilisé soit par des fluides réduits d'origine métamorphique, soit par des fluides évolués de type carboniques-aqueux et d'origine magmatique-météorique. La cassitérite a été précipitée à une température de $300\text{-}450^{\circ}\text{C}$ suite à une augmentation de la $f\text{O}_2$, du pH ou à une baisse de la température. Les données restreintes sur le gisement de type remplacement de Takua Pit Thong suggèrent que la cassitérite a été précipitée

à partir de fluides aqueux à une température $\geq 500^{\circ}\text{C}$, mais des fluides métamorphiques carboniques-aqueux ont également été impliqués.

La profondeur d'intrusion a été un des contrôles majeurs de ces types de minéralisation. Les fluides carboniques-aqueux dans la ceinture de l'ouest de Thaïlande et dans les autres ceintures de Sn-W sont la conséquence d'intrusions à une profondeur moyenne. Les fluides orthomagmatiques hypersalins sont associés à la minéralisation sub-volcanique de Sn-W.

ACKNOWLEDGEMENTS

I would like to thank Willy (A.E. Williams-Jones) for his enthusiasm and effort during the course of this study. Field work in Thailand would not have been possible without the support of numerous people in Thailand. I would especially like to thank S. Bunopas, N. Jungyusuk, S. Khositant, C. Mahawat, P. Putthapiban and S. Umavichanee. The hospitality shown to me by the many mine owners I met is greatly appreciated, particularly at the Sombat (Pilk), Som Porn and Takua Pit Thong mines.

Sincere thanks are extended to D. Baker, T. Barrett, M. Boily, J. Clark, G. Cutter, D. Francis, A. Kiddie, R. Martin, F. Nastel, T. Skulski, R. Wares and S. Wood for many stimulating discussions. I also appreciate the technical assistance provided by G. Poirier on the electron microprobe, J. Mungall scanning electron microscope, R. Rossi Raman spectrometer, T. Ahmedali and G. Keating X-ray fluorescence and D. Pezderic and K. Kyser (Univ. Sask.) for stable isotope analyses. Numerous computer drafting tricks were learned with the help of A. Bertrand and S. McCauley. A translation of the abstract was kindly provided by G. Nassif and R. Wares.

Financial support was provided by the International Development Research Centre, Ottawa. I would like to thank C. Pride for his encouragement and help, particularly at the early stages of the project. Last, but not least, I thank my parents for their constant support.

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PREFACE

The thesis contains eight chapters, four of which are in manuscript format. Chapter 2 was submitted to *The Canadian Mineralogist* (July, 1991) and Chapters 3, 4 and 5 have yet to be submitted. In each of these manuscripts the thesis advisor, A.E. Williams-Jones, is a second author. His role in the preparation of these manuscripts consisted of critical evaluation and discussion of the data and my interpretations presented within, and of the organization and writing of the text. I collected all the rock samples used in this thesis and the geological maps, except where noted, are based on my field work, completed from 1986 to 1988. The petrographic, fluid inclusion (microthermometry, Raman spectroscopy and scanning electron microscopy) and electron microprobe data were obtained by me. The whole-rock major and trace element abundances were determined by T. Ahmedali at McGill University and by Acme Analytical Laboratories Ltd., Vancouver, British Columbia, and the oxygen and hydrogen isotope analyses were provided by K. Kyser at the University of Saskatchewan, Saskatoon. The following is an excerpt from the "Guidelines Concerning Thesis Preparation" as required by the Faculty of Graduate Studies and Research at McGill University:

The candidate has the option, subject to the approval of their Department, of including as part of the thesis the text, or duplicated published text, of an original paper or papers.

- Manuscript-style theses must still conform to all other requirements explained in the Guidelines Concerning Thesis Preparation.

- Additional material (procedural and design data as well as descriptions of equipment) must be provided in sufficient detail (eg. in appendices) to allow clear and precise judgement to be made of the importance and originality of the research reported.

- The thesis should be more than a mere collections of manuscripts published or to be published. It must include a general abstract, a full introduction and literature review and final overall conclusion. Connecting texts which provide logical bridges between different manuscripts are usually desirable in the interest of cohesion. It is acceptable for theses to include, as chapters, authentic copies of papers already published, provided these are duplicated clearly and bound as an integral part of the thesis. In such instances, connecting texts are mandatory and supplementary explanatory material is always necessary.

- Photographs or other materials which do not duplicate well must be included in their original form.

- While the inclusion of manuscripts co-authored by the candidate and others is acceptable, the candidate is required to make an explicit statement in the thesis of who contributed to such work and to what extent, and supervisors must attest to the accuracy of the claims at the Ph.D. Oral Defence. Since the task of the examiners is made more difficult in these cases, it is in the candidate's interest to make the responsibilities of authors perfectly clear.

CHAPTER 1

INTRODUCTION

One of the challenges in economic geology is to explain why different styles of mineralization occur. The variety in granite-related deposits is a consequence of a number of factors. These can include the composition of the granitic rocks and the country rocks into which they intruded, the evolution of the granites, the depth of granite emplacement and the timing of vapour saturation, the evolution of the orthomagmatic fluids, the involvement of non-magmatic fluids, e.g., meteoric, metamorphic, or formational waters, and fluid-wallrock interaction. The role of the compositions of the granites and country rocks and the depth of emplacement are more constrained if deposits in a single granite belt are studied. Identifying fluid sources is also more likely to be successful if fluid compositions from different deposits in the same metallogenic belt are compared. My interest is the metallogeny of tin and tungsten and, with the above in mind, I have selected the Sn-W deposits in the western Thai granite belt for detailed study.

Most tin-tungsten deposits are associated with granitic rocks and the latter are generally interpreted to be the source of mineralization (cf. Lehmann, 1990). Three basic models have been proposed for Sn-W ore deposition: 1) **Magmatic** - direct crystallization of Sn-W minerals from granitic melt; 2) **Orthomagmatic** - partitioning of Sn-W into the vapour phase exsolving from the melt, followed by

deposition of Sn-W minerals, and; 3) **Hydrothermal redistribution** - leaching of Sn-W from granite and/or country rocks and re-deposition as Sn-W minerals. The magmatic model (e.g., Gouanvic and Gagny, 1983) is the most controversial. Many authors view the direct crystallization of cassiterite from melt with scepticism (e.g., Taylor, 1979; Štemprok, 1990) and some authors (e.g., Coetzee and Twist, 1989) reject this possibility completely. The most likely hosts for magmatic cassiterite are highly fractionated granitic rocks such as pegmatites. However, this environment of tin mineralization has received comparatively limited scientific scrutiny and as a result, the magmatic model is largely untested. By contrast, the orthomagmatic (e.g., Tischendorf and Föster, 1990) and hydrothermal re-distribution (e.g., Lehmann and Harmanto, 1990) models are both supported by well documented examples and are consequently accepted by most researchers.

The western Thai granite belt and associated Sn-W metallogenic province is one of the world's more important sources of tin and contains an unusual diversity of deposit types. These include pegmatite, disseminated, vein, stockwork, skarn, sulphide-replacement and breccia Sn-W deposits (cf. Tantisukrit, 1978). The western Thai granite belt is particularly well-known for an abundance of pegmatite-hosted deposits (Garson et al. 1975; Manning, 1986). A pegmatite deposit, at Nong Sua, and a stockwork-vein deposit, at Pilok, were therefore selected as the focus of a detailed investigation of the factors that control Sn-W mineralization. Preliminary data were also gathered on the Takua Pit Thong deposit, representative of the

sulphide replacement class. These deposits form the basis of a general evaluation of different styles of Sn-W mineralization within the context of the models discussed above.

A REVIEW OF TIN-TUNGSTEN METALLOGENY

A synopsis of the genesis of tin-bearing granites and the development of Sn-W mineralization are presented below. These subjects are also extensively reviewed by Taylor (1979), Plimer (1987), Hosking (1988) and Lehmann (1990). The final portion of this section summarizes what is currently understood about the Sn-W metallogeny of western Thailand.

Early magmatic evolution

The association between Sn-W mineralization and "specialized" granites has long been recognized (Lehmann, 1990 credits Humbolt, 1823 for this distinction). These granites are typically enriched in Sn, W, F, B, and Li, depleted in Mg, Ca and Sr, and are peraluminous and reduced. Some tin deposits are also associated with peralkaline granites, e.g. in Nigeria (Bowden, 1982), but there are no examples of this style of mineralization in Thailand and they will not be considered further. Tin granites are highly fractionated, e.g., Groves and McCarthy (1978) and tin is concentrated via fractionation. This is probably related to the oxidation state of the melt (Burnham and Ohmoto, 1980; Ishihara, 1981). Tin can occur either in the +2

or +4 valence state (cf. Durasova et al., 1986; Taylor, 1988). At high fO_2 tin behaves as a compatible element and is partitioned into minerals such as magnetite or titanite, hence, differentiation does not concentrate tin (Lehmann, 1990). In reduced melts tin behaves as an incompatible element and differentiation results in tin enrichment. The tungsten contents of Sn-W granites have received much less attention, but there is some evidence that tungsten is also enriched by differentiation (Candela and Boulton, 1990; Lehmann and Harmanto 1990).

It is still a matter of debate whether tin and the associated elements are enriched in the source regions, i.e. at the sites of granitic magma genesis. For example, Plimer (1980) and Schwartz and Askury (1989) argued for tin enrichment in the source (geochemical heritage), whereas Lehmann (1982) and Coetzee and Twist (1989) proposed that tin enrichment in granites can be explained entirely by differentiation. It has also been suggested that the high concentrations of the associated elements, e.g., F and B, may be related to enrichment of these elements in the source region (cf. Manning, 1986). In this study the question of the source of the granites will not be addressed. The western Thai granite belt is recognized as a boron-rich tin province (Manning, 1986). This is used as a starting point, and the late magmatic and hydrothermal processes that are responsible for the formation of economic mineralization are examined.

Late magmatic evolution

Highly evolved leucocratic granites are commonly associated with tin deposits. These granites may contain one or more of muscovite or Li-micas, garnet, fluorite, topaz and tourmaline. The genesis of the leucocratic granites is commonly explained by fractional crystallization followed by high temperature metasomatism, e.g., Charoy (1986), Tischendorf and Förster (1990). In some belts, e.g., the Main Range province in Malaysia, the granites can be classified texturally into "primary" and "two-phase" variants. The development of the two-phase variant is also believed to be related to "volatile fluxing" (cf. Cobbing et al., 1986; Pitfield et al., 1990). Tin-bearing pegmatites are rare in most tin granite provinces, e.g., pegmatites in the Cornubian province are rarely mineralized (Jackson et al., 1989) and no pegmatites have been reported in the Bolivian Sn-W province (Kelly and Turneaure, 1970). The western Thai granite belt, where pegmatite deposits are one of the most important sources of primary tin mineralization (cf. Garson et al., 1975), is an important exception. Most of the previous work on Thai pegmatites, and pegmatites elsewhere, has focused on the internal structure and mineralogy of these rocks, e.g., Gocht and Pluhar (1982), Gouanvic and Gagny (1987) and Foord and Cook (1989), but the genesis of tin-bearing pegmatites remains very poorly understood. This topic is therefore the focus of the first portion of the thesis (Chapters 2, 3 & 4).

Hydrothermal evolution

Orthomagmatic and hydrothermal redistribution are two models that are widely employed to explain the formation of tin mineralization (cf. Taylor, 1979). The orthomagmatic model requires that tin be partitioned into aqueous fluids that are exsolved from the melt. There are several factors that affect the aqueous fluid-melt partition coefficient of tin, the most important of which is probably the chlorinity of the fluid (cf. Taylor, 1988). These factors will be more extensively discussed in Chapter 2. Depending on the environment, the orthomagmatic fluids are interpreted to have initially been passive, as an interstitial "cumulate" (Groves and McCarthy, 1978; Eadington, 1983; Pollard and Taylor, 1986), or to have caused hydrofracturing (Plimer, 1987; Pollard *et al.*, 1987) as is the case in the model proposed by Burnham (1979) for porphyry copper systems. In the redistribution model, fluids leach tin from minerals such as biotite during alteration and later deposit cassiterite. Other elements such as Li, F and B may also be remobilized, cf. Manning and Exley (1984), Štemprok (1987).

Sn-W METALLOGENY OF WESTERN THAILAND

Southeast Asia consists of a number of tectonic terranes including the Shan-Thai (or Sibumasu), Indochina, East Malaysia, and South China blocks (cf. Bunopas and Vella, 1983; Metcalfe, 1988). The details of the tectonic history of these

blocks are not completely resolved, but there is consensus that these terranes had amalgamated and collided with Laurasia by the late Triassic. Four major granite belts are recognized in southeast Asia (Hutchison, 1983; Cobbing et al., 1986). The Main Range province consists of S-type Triassic granites in Malaysia and southern Thailand; the eastern province are I-type Permo-Triassic granites in Malaysia and eastern Thailand; the northern Thai province is characterized by Triassic S-type granites; and, the western Thai province consists of Cretaceous to Tertiary S-type granites with minor I-type intrusions in Thailand, Burma and Indonesia (Sumatra). The ore deposits investigated in this study are all associated with the western Thai granite province. The pre-Cretaceous tectonic evolution will therefore not be considered further. The reader is referred to Şengör et al. (1988) and Hutchison (1989) for a comprehensive treatment of the tectonic evolution of southeast Asia and its relationship to global tectonism.

The western Thai granite province is the most important source of tin in Thailand (Suensilpong et al., 1983) and Burma (Zaw, 1990). Geochronological studies of these granites have given conflicting interpretation of their ages. Rb-Sr isochron ages range from 78 to 124 Ma, Rb-Sr mineral ages from 55 to 72 Ma and K-Ar ages from 54 to 74 Ma (Beckinsale et al., 1979; Nakapadungrat, 1983; Putthapiban, 1984). $^{40}\text{Ar}/^{39}\text{Ar}$ age dates from Charusiri (1989) provide additional constraints, but there are still problems with some $^{40}\text{Ar}/^{39}\text{Ar}$ ages, e.g., at Pilok the mineralization (plateau age) is 2 to 5 Ma older than that of the granite (total fusion

age; 72 Ma). Owing to the dating problems, the age of the western Thai province is broadly constrained as being Cretaceous-Tertiary. The granites in the western Thai belt were first proposed by Mitchell (1977) as having been emplaced in a back-arc setting, related to the subduction of oceanic crust beneath Laurasia (the closing of Tethys prior to the collision of India). Mitchell (1977) also proposed that the S-type granites originated by partial melting of continental crust, but that the dip of the Benioff zone may have changed, thereby explaining the occurrence of minor I-type plutons in this belt. Subsequent tectonic models, e.g., Beckinsale *et al.* (1979), Hutchison (1983), Suensilpong *et al.* (1983), Mitchell (1985) and Zaw (1990), have represented variations of this theme. The Tertiary tectonic evolution of southeast Asia was dominated by major strike-slip faulting (e.g. in Thailand the Three Pagodas and Ranong faults) and rifting that developed in response to the Indian collision (cf. Tapponier *et al.*, 1986). The emplacement of some of the western Thai granites may have been controlled by the strike-slip faults (Charusiri, 1989). The uplift that exposed the western Thai granites was probably related to isostatic rebound in the back-arc (Mitchell, 1985).

The Sn-W deposits in the western Thai metallogenic province are poorly understood. Very few descriptions of the deposits have been published and, where existing, the descriptions are not detailed, e.g., Brown *et al.* (1951), Garson *et al.* (1975). Apart from reconnaissance fluid inclusion investigations by Jackson and Helgeson (1985) and Charusiri (1989), this is the first fluid inclusion study of

deposits in the Thai portion of the western Thai province. Some fluid inclusion analyses of the Burmese portion of this belt are reported by Zaw and Thet (1983) and Zaw (1984). Fluid inclusion studies of Sn-W deposits in other granite belts in Thailand are also rare, e.g., Norman and Trangcotchasan (1982), Khositantont (1990). Similarly, the only previous stable isotope data on any western Thai deposit is that of Lehmann et al. (1991). A limited number of oxygen isotope determinations on barren western Thai granites were reported by Kerrich and Beckinsale (1988).

THESIS ORGANIZATION

Chapters 2, 3 and 4 discuss the role of magmatic processes in the formation of Sn-W mineralization using the Nong Sua aplite-pegmatite as an example. Chapter 2 examines detailed geological relationships, petrographic observations, feldspar chemical data and major and trace element analyses of whole-rock aplite and pegmatite samples. These data provide the basis for proposing a model for the magmatic evolution of the aplite-pegmatite and an opportunity to evaluate the importance of this evolution to Sn-W mineralization. A complete list of the major and trace element data is provided in Appendix I and the feldspar analyses in Appendix II. Chapter 3 reports the results of a chemical investigation of tourmaline, muscovite, garnet and oxide minerals. Textural relationships indicate that these minerals formed in magmatic and/or hydrothermal environments. The

timing of the crystallization of these minerals relative to that of cassiterite and wolframite is investigated in the context of the magmatic-hydrothermal transition. All mineral chemical data presented in this chapter are listed in Appendix II. Fluid inclusion and stable isotope data are presented in Chapter 4 and are used to estimate the pressure, temperature, composition and origin of the hydrothermal fluids. These data are integrated with those from Chapters 2 and 3 to propose a genetic model that explains the Sn-W mineralization at Nong Sua. The microthermometric analyses of fluid inclusions are given in Appendix III. A study of the stockwork deposits at Pilok, selected as being representative of a hydrothermal-dominated system, is presented in Chapter 5. A genetic model for these deposits is proposed, based on geological mapping, mineral chemistry, trace element analyses, fluid inclusion and stable isotope data. A complete list of the mineral chemical analyses is presented in Appendix IV and microthermometric data in Appendix V. Preliminary data were also collected on a sulphide replacement deposit at Takua Pit Thong. The geology of this deposit and tentative interpretations of fluid inclusion and mineral chemistry data are presented in Chapter 6. A list of the mineral chemical analyses is presented in Appendix VI and microthermometric data in Appendix VII. Chapter 7 contrasts the fluid evolution of the Nong Sua and Pilok deposits, and evaluates the magmatic and hydrothermal controls of mineralization. Nong Sua and Pilok are compared to other deposits in western Thailand (including Takua Pit Thong and several deposits that are briefly

described in Appendix VIII) and the western Thai Sn-W province is then compared to other major provinces to establish whether or not the conclusions of this study have more global application.

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CHAPTER 2

EVIDENCE OF MAGMATIC CASSITERITE MINERALIZATION AT THE NONG SUA APLITE-PEGMATITE COMPLEX, THAILAND

by

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ABSTRACT

The Nong Sua intrusion is a rhythmically layered aplite-pegmatite complex which hosts Sn-W mineralization in peninsular Thailand. Both aplite and pegmatite are boron-rich, and tourmaline abundance partly defines layering. Aplite and fine-grained pegmatite have whole-rock compositions similar to that of the granite minimum. Coarse-grained plagioclase-tourmaline-rich and K-feldspar-rich pegmatites have compositions that trend away from, and towards the Or apex, respectively, in the system Q-Ab-Or. These trends are interpreted to be magmatic in origin. K-feldspar is late in the paragenesis, and the K-feldspar-rich pegmatite has the most alkaline composition, suggesting that pegmatite evolution was from peraluminous to more alkaline compositions. Cassiterite disseminated in pegmatite typically has regular or embayed contacts with plagioclase ($An_{6\text{ to }13}$) suggesting that it crystallized directly from the melt. Embayment of cassiterite by myrmekite is interpreted to indicate that cassiterite crystallized prior to or at vapour saturation. Tin abundance correlates positively with K-feldspar (and alkalinity) and is inversely proportional to primary muscovite and tourmaline contents. K-feldspar-rich pegmatite contains up to 42 ppm Sn and, using the available partition coefficients, is estimated to have been in equilibrium with a melt that contained on the order of 700 ppm Sn. This value is close to experimentally determined saturation levels for cassiterite in granitic melts and is consistent with the textural evidence for direct

cassiterite crystallization. Vapour saturation gave rise to cassiterite and wolframite mineralization hosted by veins in aplite that terminate abruptly at aplite-pegmatite contacts. This relationship suggests that the veins formed while the pegmatite was at solidus conditions.

INTRODUCTION

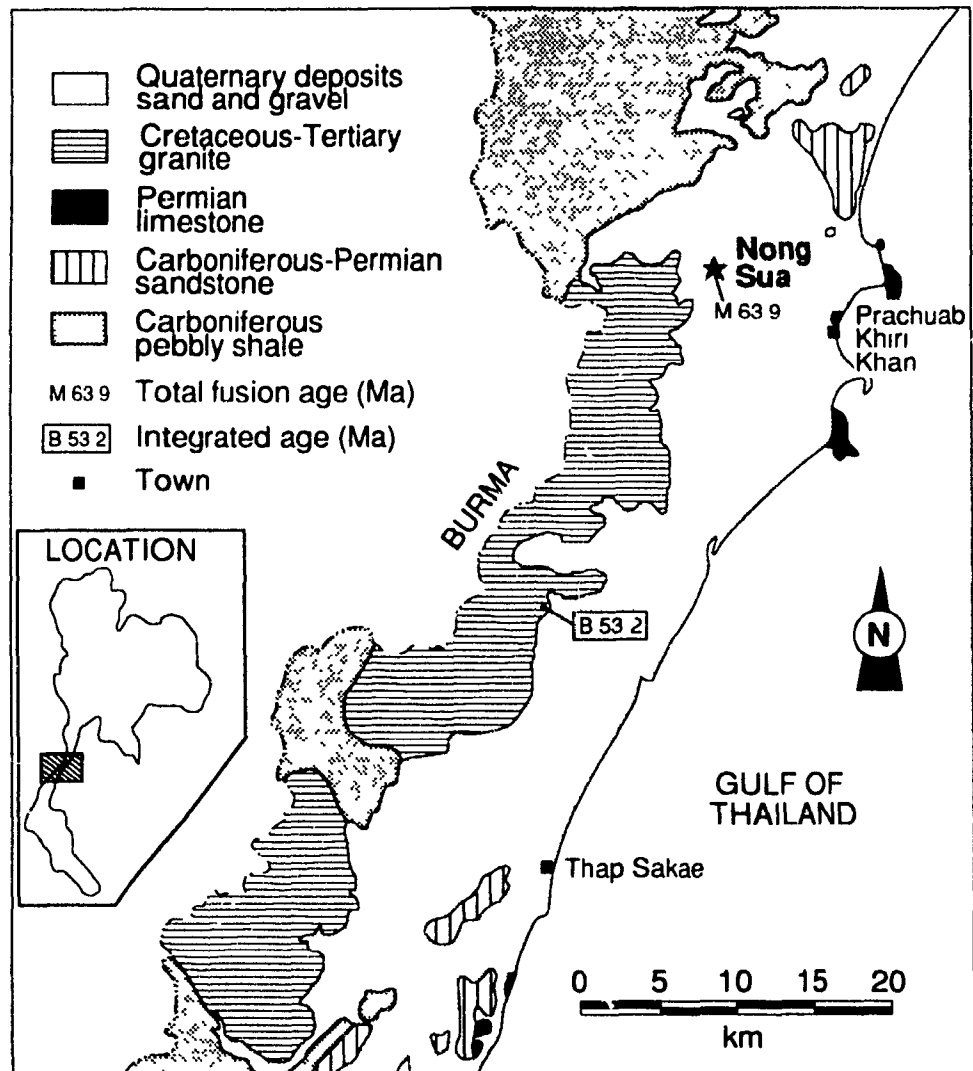
Although tin mineralization is typically associated with granitic rocks, the occurrence of cassiterite as a magmatic mineral has generally been viewed with caution (*cf.* Taylor 1979). Such caution appears to have been justified by a recent compilation of experimental work in which it was shown that tin is relatively soluble in granitic melts (Štemprok 1990). This led Štemprok (1990) to suggest that cassiterite saturation in granitic melts is unlikely, but he did not exclude this possibility for highly evolved melts. In peninsular Thailand cassiterite mineralization is commonly hosted by boron-rich pegmatites (Charusiri 1989). Some authors have linked the mineralization in these pegmatites to magmatic processes (for example, Manning 1986 suggests that the cassiterite may have crystallized from orthomagmatic fluids), however, a thorough investigation of the genesis of the mineralization has yet to be carried out. These pegmatites therefore offer a rare opportunity to evaluate the potential for crystallization of magmatic cassiterite. Unfortunately most pegmatites in Thailand are strongly weathered. An exception is the aplite-pegmatite complex at Nong Sua, which hosts cassiterite and wolframite mineralization and was therefore selected for an investigation of the behaviour of tin in highly evolved granitic systems. Detailed geologic and petrographic relationships at this deposit strongly suggest that some cassiterite formed at magmatic conditions. These observations are supported by major and

trace element data which indicate that magmatic processes dominated the evolution of the pegmatite and that cassiterite crystallization from melt is compatible with this evolution.

GEOLOGICAL SETTING

The Nong Sua tin-tungsten deposit is hosted by a rhythmically layered pegmatite-aplite complex located 10 km west of Prachuab Khiri Khan, approximately 180 km south of Bangkok (Fig. 1). The aplite-pegmatite forms a small hill surrounded by a plain of Quaternary sediments. The deposit lies <5 km east of biotite granite that forms part of the Cretaceous-Tertiary western Thai granite belt (*cf.* Cobbing *et al.* 1986) and roughly the same distance south of metamorphosed pebbly mudstone and sandstone (Fig. 1). Minor hornblende granite is also exposed southwest of Nong Sua. Tabular blocks of metasedimentary rocks are contained within the Nong Sua aplite-pegmatite complex but a lack of surrounding outcrop precludes establishing whether the aplite-pegmatite is a large dyke cutting metasedimentary rocks or whether the metasedimentary rocks are roof pendants. A total fusion $^{40}\text{Ar}/^{39}\text{Ar}$ date of 63.9 ± 0.6 Ma was obtained on a muscovite sample from the Nong Sua pegmatite and an $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 53.2 ± 0.7 Ma on biotite was obtained from a biotite granite located approximately 35 km southwest of Nong Sua (Fig. 1; Charusiri 1989).

Figure 1. Geology of the Prachuab Khiri Khan area, modified after Silpalit *et al.* (1985). $^{40}\text{Ar}/^{39}\text{Ar}$ ages of muscovite (M) and biotite (B) are from Charusiri (1989).



DETAILED GEOLOGY

The Nong Sua complex consists of alternating, laterally continuous, horizons of pegmatite, aplite and metasedimentary rocks (Fig. 2). Aplites are weakly layered to massive and, based on mineral composition, are divided into garnet and tourmaline aplite units. Much of the pegmatite is strongly layered and thus has been termed banded pegmatite (Fig. 2). A second pegmatite unit forms massive tourmaline-rich horizons that are conformable to layering in the banded pegmatite, and this unit has been distinguished as tourmaline pegmatite (Fig. 2). The attitude of the layering in the aplite and pegmatite is consistent, with a strike of $215^{\circ} \pm 10^{\circ}$ and a dip of $30^{\circ} \pm 5^{\circ}$ to the northwest. Owing to the shallow dip of the aplite-pegmatite, the plan view is of limited use in showing the distribution of the different units in the complex. Nine vertical sections were therefore measured along the length of the complex and a longitudinal section was constructed to show this distribution (Fig. 2).

APLITE

The garnet and tourmaline aplites are distinguished by the respective presence or absence of garnet. Both units are generally massive, with minor pegmatite horizons <10 cm thick that impart a weak layering to the rock. Aplite horizons are laterally continuous, in some cases for at least 100 m along strike (Fig.

2), and are conformable to layering in the complex. The aplites are composed of quartz, K-feldspar and plagioclase, $\pm$ muscovite, $\pm$ tourmaline, $\pm$ garnet $\pm$ apatite. The grain size is variable, (typically <0.5 to 2 mm) with some larger anhedral K-feldspar grains being up to 2 cm across. Quartz and K-feldspar, and rarely quartz and plagioclase, are present as dendritic micrographic intergrowths. Three varieties of K-feldspar are common: perthite, tartan twinned, and untwinned. Plagioclase is either twinned or occurs as myrmekitic intergrowths with quartz. Garnet forms equant subhedral grains that are generally concentrated in laminae parallel to, or at a shallow angle to, the principal layering. Disseminated garnet also occurs as anhedral grain clusters up to 0.5 cm across. Tourmaline is present as disseminated grains up to 5 mm across that are typically subhedral and poikilitic with rounded quartz inclusions.

PEGMATITE

The banded and tourmaline pegmatites are similar mineralogically but are texturally distinct. The tourmaline pegmatite is massive, whereas the banded pegmatite is strongly layered. The division between these units is drawn somewhat arbitrarily on Figure 2 as there are tourmaline pegmatite layers within the banded pegmatite unit that are too thin to be represented on this diagram. Both units are composed of quartz, plagioclase, K-feldspar, tourmaline, muscovite, and trace apatite; minor garnet is locally present in the banded pegmatite. Layering in the

Figure 2. A plan view and longitudinal section showing the geology of the Nong Sua aplite-pegmatite.

banded pegmatite is recognized by changes in the grain size and/or modal proportions of minerals. In particular, the abundance of tourmaline delineates light and dark horizons. Individual horizons commonly contain tourmaline and feldspar crystals that are oriented perpendicular to layering. These features indicate that the layering is magmatic in origin. This conclusion is similar to that of Manning (1983) who suggested that tourmaline at Nong Sua is magmatic in origin, from compositional and textural relationships.

The banded pegmatite consists of the following five lithologies and gradations between them: 1) fine-grained tourmaline-rich pegmatite; 2) fine-grained aplite; 3) coarse-grained K-feldspar-rich; 4) coarse-grained graphic quartz-K-feldspar; and 5) coarse-grained tourmaline-rich pegmatite. Thin (<10 cm) quartz-perthite-plagioclase-tourmaline $\pm$ muscovite dykes also locally crosscut the pegmatite and aplite units. The grain size of the banded pegmatite varies from 2 to 5 mm in the fine-grained tourmaline-rich pegmatite and aplite layers to greater than 10 cm in the coarse-grained K-feldspar-rich and graphic pegmatite layers. The coarse-grained tourmaline-rich pegmatite typically contains crystals 2 to 5 cm and rarely up to 15 cm long. Fine-grained tourmaline-rich layers are volumetrically the most important lithology in the banded pegmatite. They are composed of coarse-grained anhedral perthite (up to 5 cm across) set in a finer-grained matrix of plagioclase, quartz, tourmaline, muscovite and K-feldspar (perthite and microcline).

The K-feldspar to plagioclase ratios of the different units are highly variable, but much of the fine-grained tourmaline-rich unit has a granitic composition (discussed below). Coarse-grained tourmaline-rich pegmatite typically contains very little K-feldspar, but subhedral to euhedral primary muscovite grains, up to 3 cm across, are common. K-feldspar-rich pegmatite is composed of either graphic quartz-perthite intergrowths, best developed immediately above the garnet aplite contact (near section 7, Fig. 2), or perthite-quartz layers in which K-feldspar crystals are euhedral. Minor K-feldspar along fractures also occurs as a replacement of plagioclase.

Deformation features are variably developed in the pegmatite. These include: broken tourmaline crystals (perpendicular to the c-axis); kinked muscovite; quartz, plagioclase and K-feldspar with undulose extinction; and plagioclase with displaced twins. Quartz and feldspar are not strained in some samples and highly deformed in others.

METASEDIMENTARY ROCKS

The metasedimentary xenoliths/roof pendants are equivalents of the Carboniferous to Permian Kaeng Krachan Group pebbly mudstone and sandstone, and occur as tabular blocks within the aplite-pegmatite (Fig. 2). They are typically massive, and are composed of quartz, biotite, muscovite, rutile, and locally hornblende. Within 10 cm of the contacts with aplite or pegmatite the

metasedimentary rocks are generally altered to a tourmaline-quartz assemblage. The contacts are parallel or subparallel to the layering of the pegmatite, and pegmatite dykes rarely crosscut individual metasedimentary blocks. In the area surrounding Nong Sua near the western Thai granite belt the Kaeng Krachan Group sedimentary rocks have been metamorphosed to biotite facies. The grade of regional metamorphism decreases away from the granites to sub-greenschist facies near the Kaeng Krachan dam.

MINERALIZATION

Vein, disseminated and contact styles of mineralization are recognized at the Nong Sua deposit. Cassiterite is present in all three styles, whereas wolframite is almost exclusively vein-hosted. Disseminated cassiterite also contains significant concentrations of Nb-Ta oxides. The timing of each style of mineralization is constrained by textural and structural relationships with the host aplite and pegmatite. Thus cassiterite and wolframite crystallization can be related to the evolution of the pegmatite. The relative economic importance of the different styles of mineralization cannot be ascertained, as the mine is not currently in production and records of past production are not available.

Vein Mineralization

Mineralized veins are <1 m thick, <10 m long, and generally lack alteration

haloes. Contacts between veins and aplite are usually very sharp with the mineralization being restricted to the vein. Rarely, vein contacts are diffuse, reflecting quartz flooding of the aplite. At such contacts the mineralization may be disseminated in the wallrock immediately adjacent to the vein. An important feature of the veins is that they are contained within the aplite units and terminate abruptly at the contacts between aplite and banded pegmatite (Fig. 3). This suggests that the aplite was at subsolidus conditions but that the pegmatite had not cooled sufficiently to sustain brittle fracture, *i.e.*, the pegmatite was at or near solidus conditions during vein formation. Two additional features support this interpretation. Near section 9 (Fig. 2), a tourmaline layer in aplite continues through a quartz vein. This feature is similar to the "phantom layering" through perthite crystals described by Jahns & Tuttle (1963). At Nong Sua, we interpret that the aplite was locally at solidus conditions during vein formation, and that the tourmaline layer and quartz vein formed synchronously. In the same area a late pegmatite dyke crosscuts a vein, indicating that there was at least some magmatic activity after vein formation. Veins are therefore concluded to have formed from aqueous fluids while the pegmatite was at solidus conditions.

Veins which host mineralization are composed largely of quartz and tourmaline, with accessory K-feldspar, plagioclase, muscovite, cassiterite and wolframite, trace amounts of scheelite and apatite, and rare chalcopyrite, arsenopyrite, and pyrite. Where sulphides are observed, they occur along fractures

Figure 3. Quartz-tourmaline veins cutting garnet aplite that terminate abruptly at the aplite-pegmatite contact. The abbreviations are: A - aplite, M - metasedimentary rocks, P - pegmatite, V - vein. The field of view is approximately 3 m across.



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in quartz and thus post-date vein formation. The feldspars are generally untwinned although twinned plagioclase and perthitic K-feldspar are not uncommon. Muscovite occurs as incipient alteration of feldspars and does not constitute a primary vein mineral. Vein tourmaline crystals range up to 4 cm long, and are typically broken perpendicular to the c-axis and infilled by quartz $\pm$ feldspar. Strain is variable; quartz is unstrained in some veins and highly deformed in others. In the latter veins, K-feldspar and plagioclase have thin arcuate shapes which are probably a result of pressure dissolution. Wolframite, cassiterite, tourmaline and, to a lesser extent, feldspar are intergrown in the veins. Wolframite occurs as coarse-grained euhedral blades up to 4 cm long. Minor scheelite $\pm$ muscovite has replaced wolframite along fractures. Cassiterite forms dark red, subhedral to anhedral crystals that are typically 1 cm across and are locally up to 4 cm across.

Disseminated Mineralization

Disseminated cassiterite in the banded pegmatite unit, and less commonly in the garnet aplite unit, occurs in small patches (<0.5 m across) that have no discernible orientation. Mineralized pegmatites and aplites contain the highest tin grades encountered in this study, but are otherwise identical to those barren of mineralization. The cassiterite is dark red, subhedral to euhedral, typically <1 cm across, and is intergrown with quartz, K-feldspar, plagioclase, muscovite, and in particular with tourmaline. Opaque inclusions are common in disseminated

cassiterite but are rare in vein-hosted cassiterite. The inclusions have been identified from electron microprobe analyses as Nb-Ta rutile, ilmenite, columbite and ixiolite. These minerals are also common to other pegmatite-hosted tin deposits in Thailand (Praditwan 1988).

Contact Mineralization

A very small proportion of the mineralization is classified as contact-style, and occurs as cassiterite disseminated in pegmatite at the contacts with metasedimentary rocks where it is in close spatial association with coarse-grained secondary muscovite and beryl. The coarse-grained muscovite has replaced feldspars, is intergrown with cassiterite, and defines funnel-shaped zones which spread out along the contacts, but do not extend into the metasedimentary rocks. This is interpreted to reflect a hydrothermal process in which the metasedimentary rocks acted as an impermeable boundary and forced fluids to flow laterally.

FELDSPAR MINERALOGY

Feldspar compositions and textural relationships imply a history of cassiterite mineralization that involved a complex interplay of magmatic and hydrothermal processes. In the aplites and pegmatites, plagioclase and K-feldspar have several distinct modes of occurrence and, in the case of plagioclase there are at least two

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compositional varieties. The feldspar mineralogy of the veins is somewhat simpler, consisting solely of albite and two types of K-feldspar (untwinned and perthite).

Textural Relationships

Plagioclase generally occurs as subhedral to euhedral grains in aplite and pegmatite. It also is found as exsolved lamellae, inclusions, and as discontinuous veinlets in perthite. The plagioclase inclusions are distinctive in that they are invariably euhedral and commonly exhibit Carlsbad twinning, whereas other plagioclase has polysynthetic twins or, less commonly, is untwinned. In contrast to plagioclase, K-feldspar in the aplite and the finer-grained pegmatite units is anhedral. The coarser-grained K-feldspar in these units is typically perthitic; finer-grained K-feldspar comprises microcline (tartan twinned) and a phase that is untwinned. Exsolved plagioclase in the perthite varies texturally from string-like to coarser-grained and ragged. The latter show crude parallel alignment and are optically continuous. The coarse-grained perthite crystals are irregularly distributed and are surrounded by a finer-grained matrix composed mainly of plagioclase, quartz and microcline. The occurrence of euhedral plagioclase crystals that extend from the matrix into the perthite (Fig. 4a), suggests that perthite post-dated the subhedral to euhedral variety of plagioclase. This conclusion is supported by the presence of euhedral plagioclase inclusions in perthite.

A conspicuous feature of the pegmatites, and to a lesser extent the aplites,

is the occurrence of myrmekite at perthite-plagioclase grain boundaries. Myrmekite commonly forms anhedral masses that display convex contacts into the perthite. Some of these masses resemble bulbous or lobate myrmekite (Fig. 4b; *cf.* Phillips 1974; Hibbard 1979). Less commonly, the contacts are irregular with numerous reentrants in perthite. Myrmekite also occurs as an interface that separates subhedral to euhedral plagioclase from perthite. In such cases the myrmekite forms part of the plagioclase crystal, and albite twins, visible elsewhere in the plagioclase, are destroyed in the domain occupied by the myrmekite. These textures suggest that the myrmekite represents a reaction between perthite and plagioclase. The observation that myrmekite zones locally crosscut single perthite grains supports the replacement origin.

Graphic intergrowths are much less common than myrmekite, and are developed mainly between quartz and perthite in both aplite and pegmatite. Rare graphic intergrowths of plagioclase and quartz are observed in aplite. These grains have a plumose structure with optically continuous quartz radiating outward. Perthite that is graphically intergrown with quartz is otherwise indistinguishable from that described above. It may contain plagioclase inclusions and is commonly rimmed by plagioclase and myrmekite. In some cases the graphic quartz rods intergrown with perthite extend into the myrmekite indicating that the latter replaced pre-existing perthite.

A significant proportion of the perthite, including that intergrown with

quartz, has been converted to microcline. Development of tartan twinning is spatially associated with zones of undulose extinction and is therefore considered to have been stress induced. A similar association between exsolution in perthite and undulose extinction may indicate that exsolution was also stress-related. Much of the tartan-twinned K-feldspar in aplite occurs in the groundmass. The origin of this K-feldspar is uncertain, however, its interstitial nature may indicate that it is late in the paragenetic sequence. Some K-feldspar occurs as a replacement product of subhedral to euhedral plagioclase. This is also seen in the rare crystals of plagioclase that are graphically intergrown with quartz.

The feldspar mineralogy of the veins is comparatively simple. The plagioclase is typically untwinned. Two types of K-feldspar are observed: untwinned, and perthitic. The perthite contains fine-grained string exsolution of plagioclase. Coarse-grained ragged exsolution, which characterizes much of the perthite in aplite and pegmatite, is not observed.

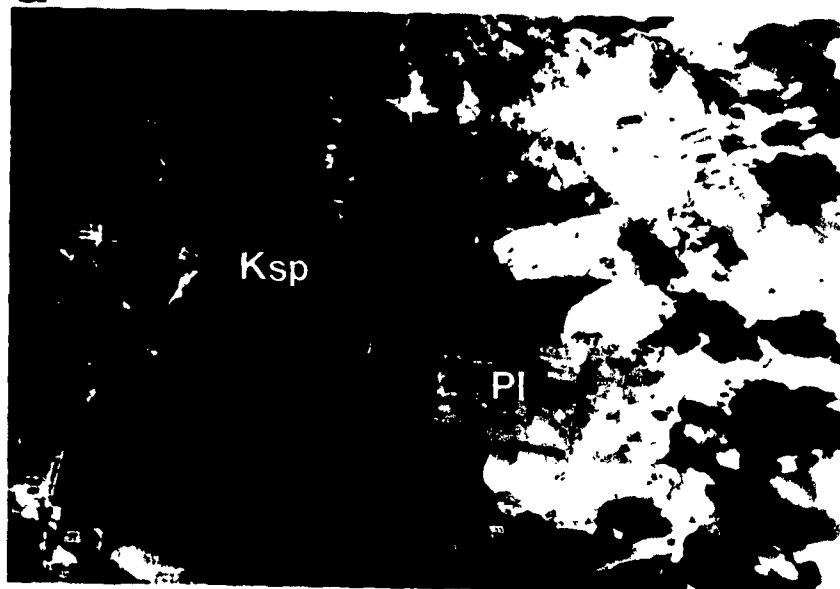
Cassiterite-Feldspar Textures

The relationship between cassiterite and feldspars varies with the style of mineralization. Vein-hosted cassiterite and K-feldspar or cassiterite and albite are intergrown along crystal faces suggesting that the cassiterite and feldspars were contemporaneously deposited from aqueous fluids. Cassiterite-feldspar relationships in disseminated mineralization are more complex. The contacts between

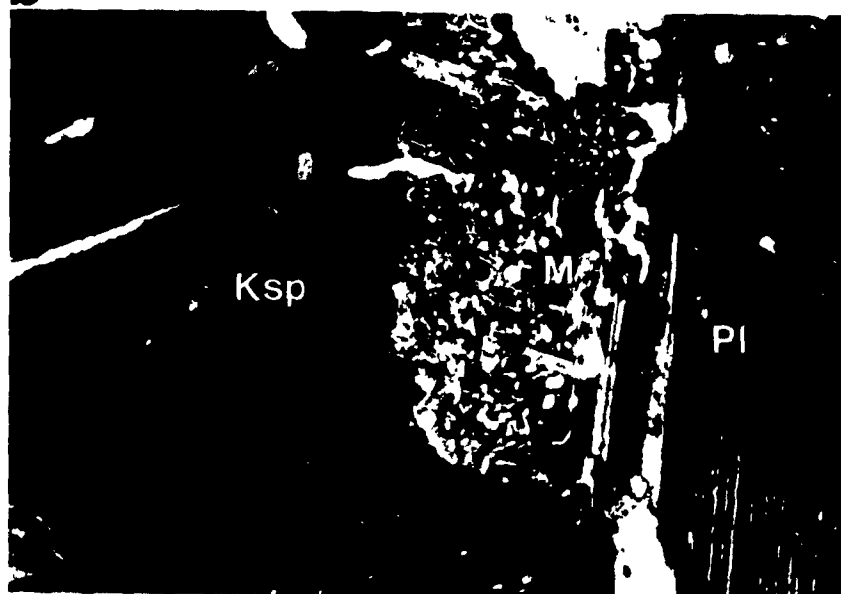
Figure 4. Photomicrographs of feldspar and cassiterite-feldspar textures.

- a) euhedral plagioclase (Pl) enclosed by coarse-grained anhedral perthite (Ksp); field of view 3.3 mm
- b) bulbous myrmekite (M) between K-feldspar (Ksp) and plagioclase (Pl); field of view 3.3 mm
- c) cassiterite (Cas) embayed by An_{10} plagioclase (Pl); field of view 0.85 mm
- d) myrmekite (M) embaying cassiterite (Cas) and perthite (Ksp); tourmaline (T) is also intergrown with cassiterite; field of view 2.6 mm

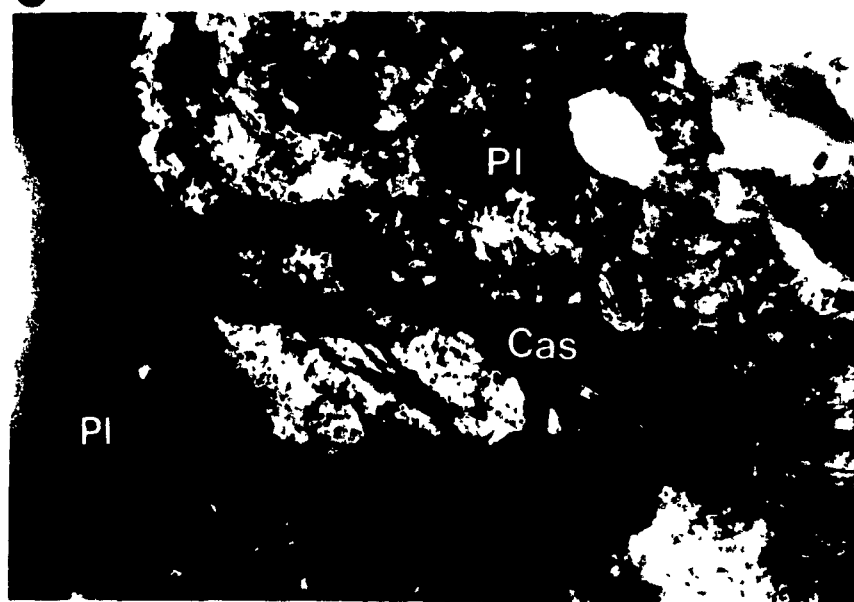
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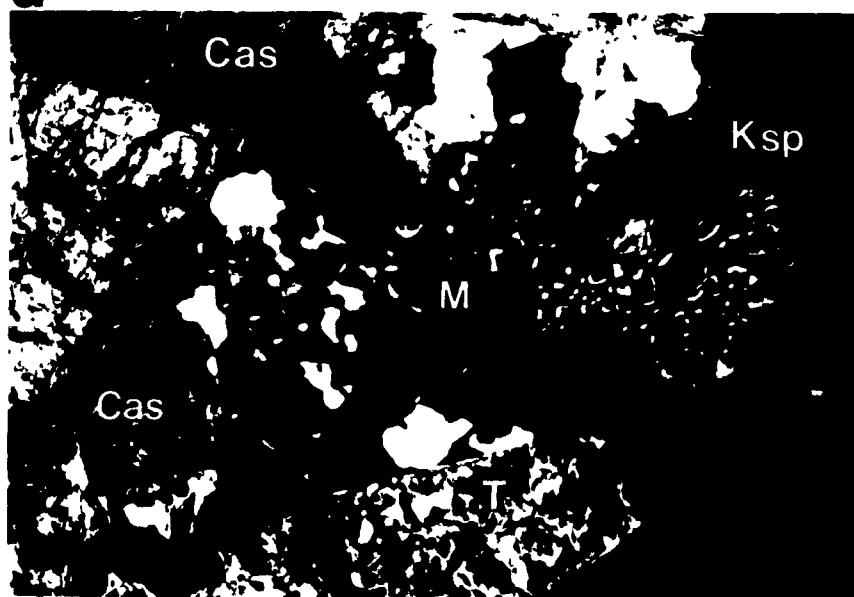
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c



d



disseminated cassiterite and plagioclase are also generally defined by crystal faces. Locally however, cassiterite has been resorbed and is embayed by plagioclase (Fig. 4c). Cassiterite more commonly displays embayed contacts with perthite, microcline and myrmekite (Fig. 4d). The feldspars at embayed contacts contain fine-grained inclusions of cassiterite. These features suggest that cassiterite crystallization was pene-contemporaneous with, but locally preceded, the crystallization of some of the subhedral to euhedral plagioclase, and was followed by the formation of perthite and myrmekite.

Feldspar Chemistry

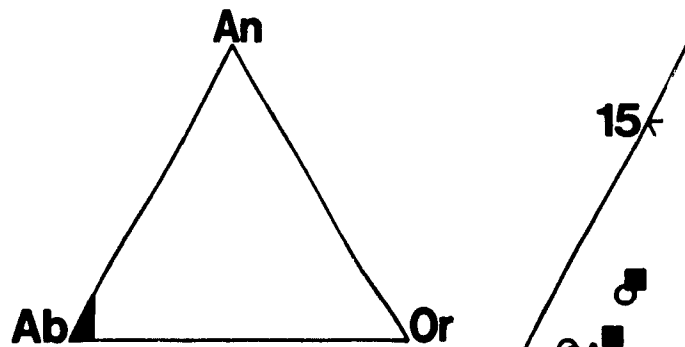
Plagioclase compositions were determined by electron microprobe analyses and can be divided into two groups: An_0 to An_6 and An_6 to An_{13} (Fig. 5). Subhedral to euhedral plagioclase in both the aplite and pegmatite units typically has compositions that fall in the more calcic group. The Or contents of the plagioclase are typically <2 % (Fig. 5) which indicate interaction with subsolidus hydrothermal fluids (*cf.* Parsons & Brown 1984). Within most crystals and amongst crystals in the same sample the plagioclase composition varies by less than An_1 . Plagioclase that is exsolved from, or occurs as inclusions in K-feldspar has compositions that fall in the more albitic group, typically in the range An_1 to An_3 . Plagioclase in myrmekite generally is compositionally indistinguishable from non-myrmekitic plagioclase in the same sample. In both cases the An values range

from 6 to 13. Some myrmekite grains are zoned, being most sodic at K-feldspar contacts and increasingly calcic away from the K-feldspar; the most calcic myrmekite has the same An-Ab-Or contents as non-myrmekitic plagioclase in the same sample. An example of this zoning is illustrated in Figure 6. In quartz veins, the plagioclase compositions are consistently albitic and range from An_0 to An_3 (Fig. 5).

K-feldspar compositions do not differ significantly as a function of either the textural class of the mineral or the nature of the host rock. Most crystals (K-feldspar portions of perthite, tartan twinned, and untwinned) have compositions within the range Or_{90} to Or_{96} and contain a negligible anorthite component.

Fenn (1986), London *et al.* (1989) and MacLellan & Trembath (1991) have interpreted graphic intergrowths of quartz and K-feldspar to reflect non-equilibrium growth (supersaturation) in vapour-undersaturated water-rich conditions. Much of the coarse-grained perthite, including that graphically intergrown with quartz, formed after oligoclase-calcic albite crystallization. Therefore much of the feldspar crystallization probably occurred at vapour-undersaturated conditions. Planar cassiterite-oligoclase contacts suggest pene-contemporaneous growth, and the embayment of cassiterite by oligoclase (Fig. 4c) implies that some cassiterite crystallized before the end of plagioclase crystallization. The textural evidence therefore indicates that disseminated cassiterite crystallized directly from the melt. An alternate explanation that must be examined is whether the disseminated

Figure 5. Plagioclase and myrmekite compositions from barren and mineralized aplite and pegmatite and from mineralized veins. Each point represents the average composition of several crystals in a single sample.



PLAGIOCLASE

- mineralized pegmatite
- barren pegmatite
- mineralized aplite
- barren aplite
- x vein

MYRMEKITE

- △ mineralized
- ▲ barren

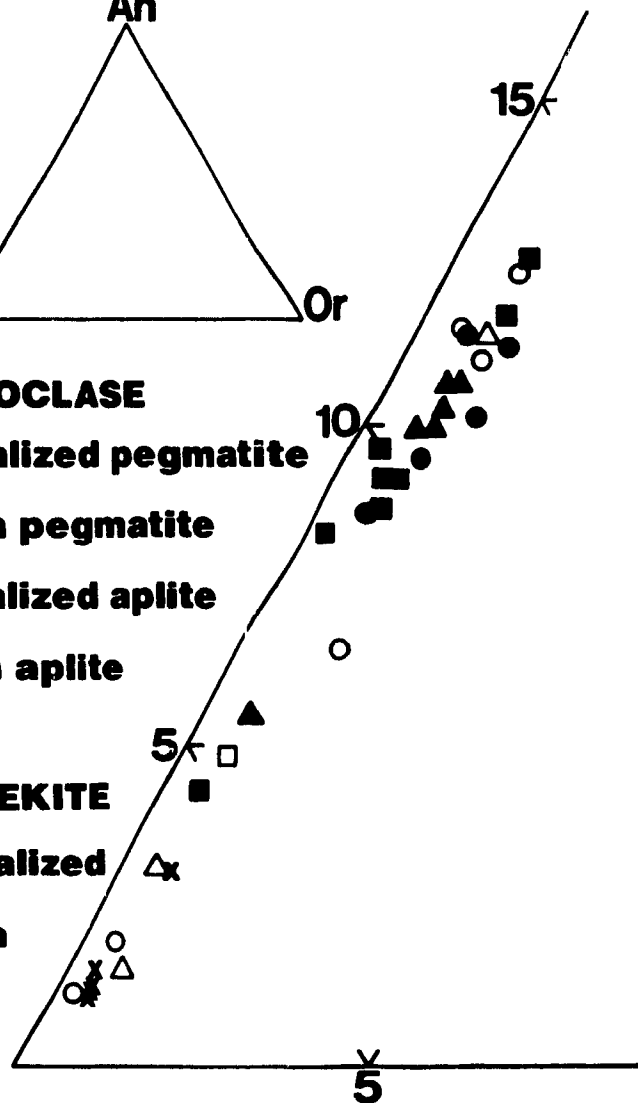
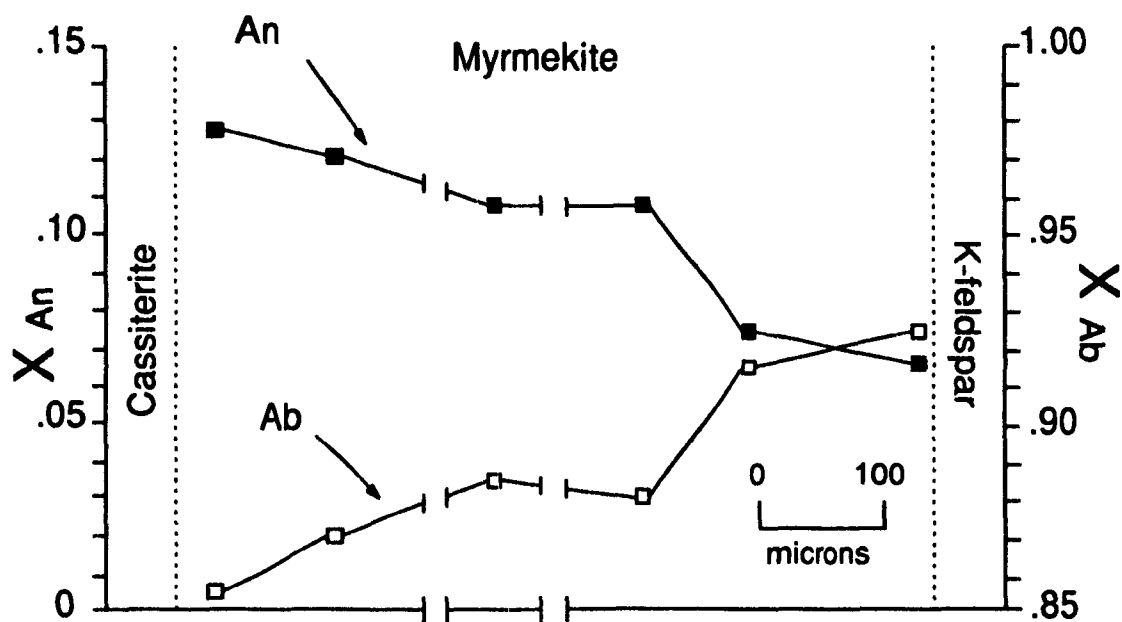


Figure 6. Plagioclase compositions across a zoned myrmekite grain. The myrmekite has embayed contacts with cassiterite and perthite (*cf.* Fig. 4d).



cassiterite could have precipitated from aqueous fluids in a vapour-saturated melt (after the vapour and melt separated, the chemical potentials of tin in melt and fluid were probably different). The occurrence of myrmekite late in the sequence of pegmatite crystallization provides some constraints on cassiterite formation. Myrmekite in similar environments has been ascribed to a late vapour-saturated magmatic event by Hibbard (1979) and Hopson & Ramseier (1990). The compositional similarity between myrmekite and magmatic plagioclase at Nong Sua suggests that both crystallized at comparable conditions. We therefore conclude that myrmekite reflects vapour saturation of the melt, although an early subsolidus origin cannot be entirely discounted. The disseminated cassiterite clearly pre-dates myrmekite (Fig. 4d) and thus probably began to crystallize before vapour saturation, but some disseminated cassiterite may have also crystallized in the presence of melt and vapour.

After the attainment of vapour saturation in the melt, compositions of plagioclase crystallizing from aqueous fluids were dependent on of T, P, and in particular, fluid composition. The primary magmatic fluid of the pegmatite has been interpreted by Linnen & Williams-Jones (1991) to be of low salinity. In general the Ca/Na ratio of fluids coexisting with melt is strongly dependent on chlorinity, and for the low salinity fluids at Nong Sua, the Ca/Na ratio of the fluid would have been much lower than that of the melt (*cf.* Candela, 1989). Similarly, plagioclase that crystallized from the orthomagmatic fluid would have had a very

low Ca/Na ratio (Schliestedt & Johannes, 1990). Therefore the albite (An_0 to An_3) observed in veins that host cassiterite at Nong Sua could have formed from hydrothermal fluid while the pegmatite was at or near solidus conditions.

WHOLE ROCK CHEMISTRY

If, as the textures suggest, cassiterite indeed crystallized from the melt, the chemical evolution of the aplite-pegmatite and the behaviour of tin should reflect this evolution. Four types of samples were analyzed for major and trace elements: 1) aplite; 2) fine-grained banded pegmatite; 3) coarse-grained K-feldspar-rich and graphic pegmatite, and 4) coarse-grained tourmaline-rich pegmatite. Average compositions for each of the sample types are listed in Table 1. Sample sizes ranged from 1 to 2 kg, and are probably representative for the aplite and fine-grained pegmatites. Analyses of the coarse-grained pegmatite units may not be representative; however, the analyses of coarse-grained units represent, at worst, a mixture of lithologies and are still useful in assessing the behaviour of tin during pegmatite evolution and any subsequent alteration.

MAJOR AND TRACE ELEMENT RELATIONSHIPS

The presence of primary muscovite, garnet and tourmaline indicate that the aplites and pegmatites are peraluminous. This is also reflected by normative corundum values which range from 0.79 to 5.08 wt. %. Rubidium contents vary

TABLE 1
Average Whole-Rock Compositions of Aplite and Pegmatite

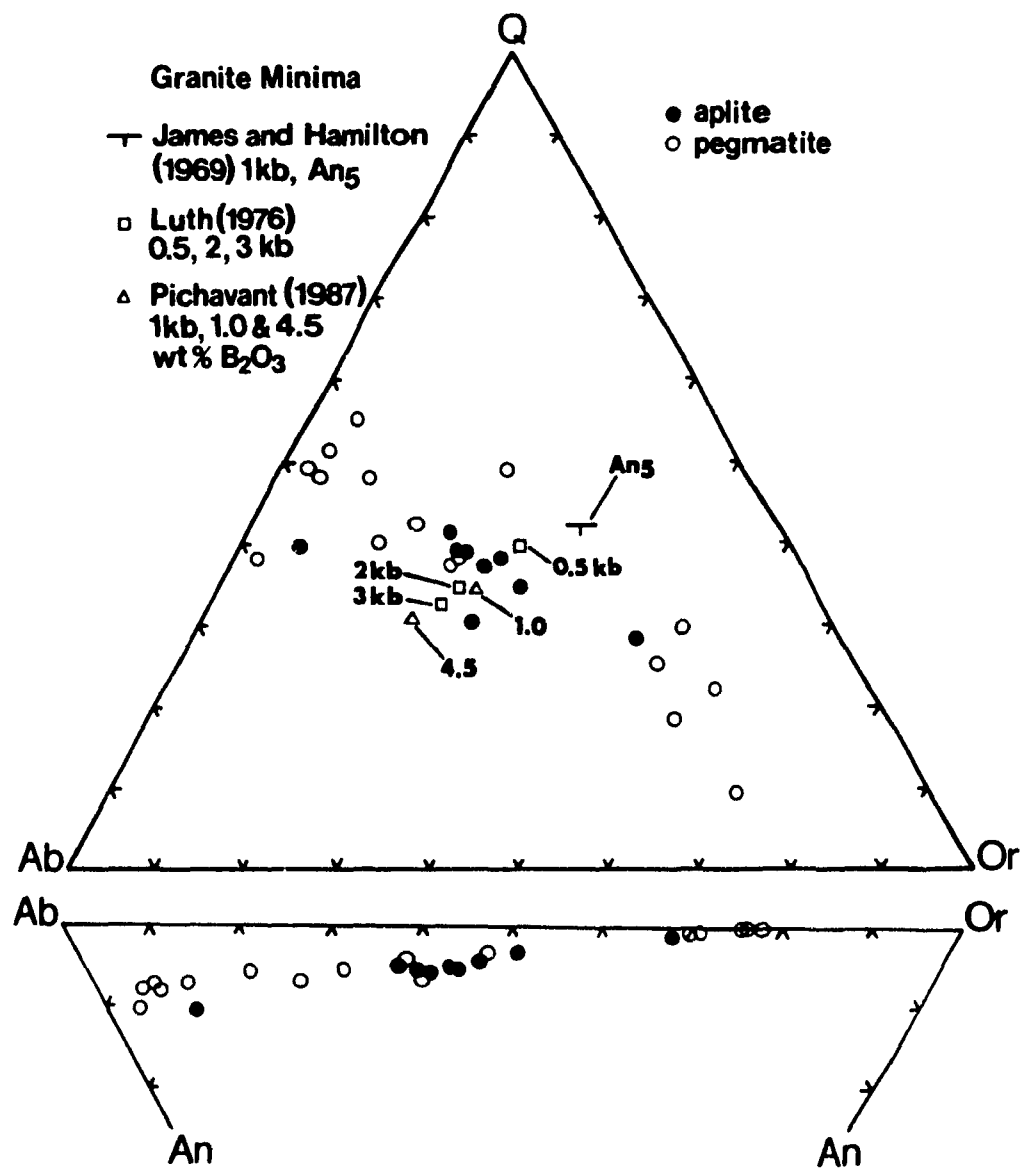
Type	Apl. 9	Ksp. 7	Tour. 4	Apl.BP. 1	BP. 4
SiO ₂	74.95	74.45	77.86	76.70	76.37
TiO ₂	0.04	0.02	0.08	0.02	0.06
Al ₂ O ₃	14.88	14.36	13.74	14.85	14.49
Fe ₂ O ₃	0.50	0.17	1.25	0.27	0.79
MnO	0.06	0.01	0.03	0.01	0.02
MgO	<0.01	<0.01	0.03	<0.01	<0.01
CaO	0.72	0.25	0.76	1.39	0.76
Na ₂ O	3.97	2.85	4.60	6.37	4.42
K ₂ O	4.28	7.60	1.16	0.42	2.87
P ₂ O ₅	0.05	0.05	0.03	0.05	0.04
LOI	0.59	0.25	0.50	0.29	0.45
Total wt%	100.10	100.01	100.05	100.38	100.28
BaO ppm	14	17	17	42	18
Cr ₂ O ₃ ppm	<15	<15	<15	<15	<15
V ppm	<10	<10	<10	<10	<10
Ni ppm	<10	<10	<10	<10	<10
Nb ppm	27	10	52	35	39
Zr ppm	23	3	16	28	13
Y ppm	20	5	14	23	11
Sr ppm	12	11	15	8	15
Rb ppm	399	626	96	376	233
Pb ppm	32	46	16	24	21
Th ppm	8	5	8	8	11
U ppm	<5	<5	<5	<5	<5
F ppm	384	107	483	140	385
Li ppm	22	10	28	14	18
Ta ppm	8	4	13	10	14
B ppm	256	217	4133	369	1951
Sn ppm	17	22	14	12	16
Norm					
q	34.09	28.34	45.06	35.20	38.14
or	25.27	44.94	6.87	2.48	16.96
ab	33.56	24.14	38.91	53.90	37.40
an	3.24	0.99	3.59	6.55	3.50
c	2.54	1.10	3.61	1.52	2.83
il	0.06	0.02	0.07	0.02	0.05
hem	0.43	0.17	1.25	0.27	0.79
ap	0.12	0.10	0.07	0.12	0.09
ru	0.01	0.01	0.04	0.01	0.04
hy	0.04	0.02	0.08	0.02	0.03
mt	0.11				

Apl. represents aplite, Ksp. K-feldspar-rich pegmatite, Tour. tourmaline-rich pegmatite, Apl.BP. aplite contained within banded pegmatite, and BP. fine grained tourmaline-rich banded pegmatite. Major elements, BaO, Cr₂O₃, V and Ni were determined from fused pellets by X-ray fluorescence with a detection limit of 0.01 wt %, 15 ppm for Cr₂O₃, and 10 ppm for BaO, V, and Ni. Total iron is reported as Fe₂O₃. Most samples contain < 0.01 wt % MgO, therefore median values are given. For TiO₂, MnO, and P₂O₅ longer counting times were used to improve the detection limit to 0.001 wt %. Except for Sn, B, F, Ta and Li trace elements were determined from pressed powder pellets by X-ray fluorescence with detection limits of 3 ppm for Nb, Zr, and Sr, and 5 ppm for the rest. Sn was determined by NH₄l fusion and atomic absorption analysis with a detection limit of 1 ppm. Three aplite samples contained visible cassiterite (53, 95, and 106 ppm Sn) and are excluded from the average. B was determined by Na₂O₂ fusion and induced coupled plasma analysis with a detection limit of 2 ppm. F was determined by NaOH fusion and specific ion electrode analysis with a detection limit of 10 ppm. Ta and Li were determined by induced coupled plasma analysis after 0.5 g of sample was digested with 10 ml of HClO₃-HNO₃-HF at 200°C and diluted with 10 ml of aqua regia. The detection limit of Ta and Li is 3 ppm.

from 37 to 995 ppm, Sr from 7 to 18 ppm, and Ba from <10 to 54 ppm. The strong enrichment of Rb relative to Sr and Ba is evidence that the pegmatitic melt was highly evolved (El Bouseily & El Sokkary 1975). Figure 7 shows aplite and pegmatite compositions expressed in terms of normative quartz-orthoclase-albite and anorthite-orthoclase-albite. Most of the aplite and fine-grained tourmaline-rich pegmatite samples are similar in composition to the granite minimum. Other samples are either enriched or depleted in K-feldspar. In this respect the compositions are comparable to those of other pegmatites, *cf.* Jahns & Tuttle (1963), and considering the coarse grain size, the degree of scatter is relatively small. The absence of compositional shifts toward the Q or Ab corners on Figure 7 is consistent with the lack of petrographic evidence for significant greisen and albitic alteration. It is difficult to assess potassic alteration using Figure 7 as pegmatite compositions trend toward the Or apex, but the lack of widespread plagioclase replacement by K-feldspar suggests that potassic alteration is of limited importance in controlling the whole rock chemistry. The restriction of veins to the aplite units indicates that the aplite crystallized prior to the pegmatite. The granitic composition of the aplite therefore probably represents the composition of pegmatite melt during emplacement (excluding elements such as boron which are partitioned into the vapour).

Most of the samples contain significant CaO and B₂O₃, both of which change the composition of the granite minimum; the normative anorthite component is

Figure 7. Compositions of aplite and pegmatite expressed in terms of normative quartz (Q), albite (Ab), orthoclase (Or), and anorthite (An), and granite minima for various pressures and compositions.



typically 3 to 4 wt %, and B_2O_3 contents range up to approximately 1 wt. %. The addition of an anorthite component to the haplogranite system shifts minimum melt compositions toward the quartz-orthoclase tie-line, whereas the addition of boron has the opposite effect, moving compositions toward the albite apex (James & Hamilton 1969; Pichavant, 1987). Although the bulk of the aplite compositions are close to granite minimum compositions at 1 to 2 kbar (Fig. 7) this comparison does not provide quantitative estimates of pressures owing to the opposing effects of Ca and B. However, if tourmaline is present during crystallization, the Q-Ab-Or phase relationships are not significantly affected (Benard *et al.* 1985). Therefore 1 to 2 kbar is probably the minimum pressure of crystallization and higher pressures are more likely.

The dependence of chemical variation of major elements on the K-feldspar to plagioclase ratio is also apparent when the major elements are plotted as a function of SiO_2 . Figure 8 shows that K_2O and Al_2O_3 vary inversely, whereas Na_2O and CaO vary positively with increasing SiO_2 . In terms of the alkali to aluminum ratio the most peraluminous rocks are plagioclase-rich with very little K-feldspar. This is reflected, mineralogically, by the presence of primary muscovite in these rocks. Higher K-feldspar contents yield less peraluminous compositions; this trend continues through to the coarse-grained K-feldspar-rich samples which are meta-aluminous (Fig. 9). The timing of crystallization of the different units is poorly constrained, except that K-feldspar is commonly late. Since graphic quartz-K-

Figure 8. Compositions of aplite (solid circles) and pegmatite (open squares) expressed in terms of weight percent K_2O , Na_2O , Al_2O_3 , and CaO versus SiO_2 .

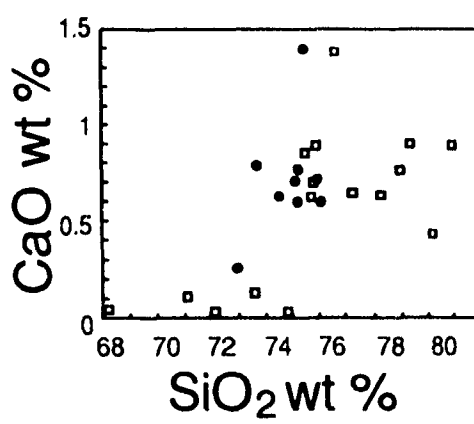
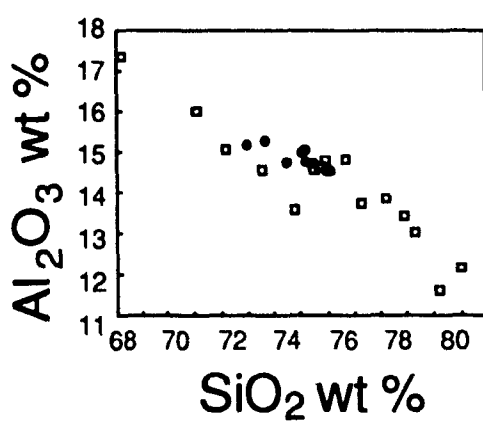
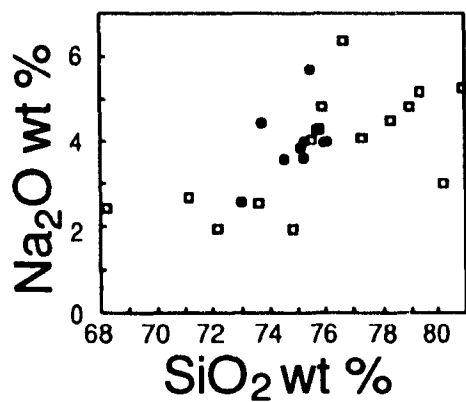
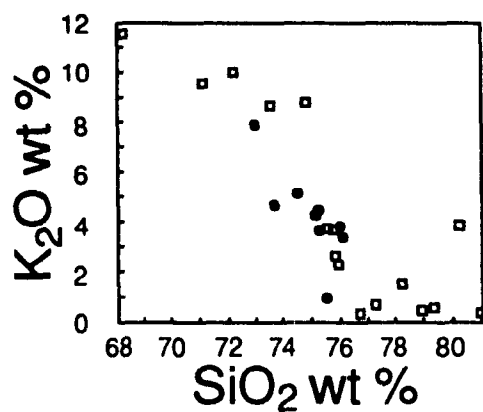
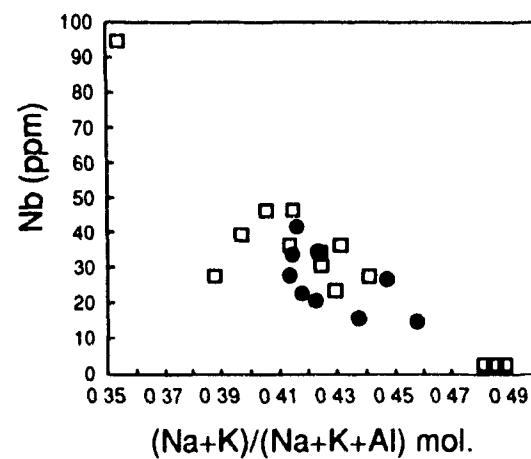
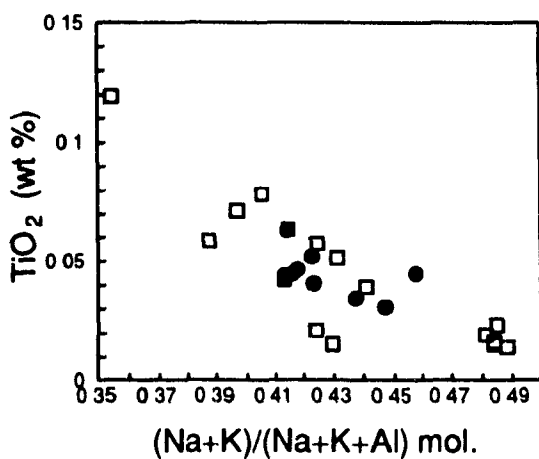
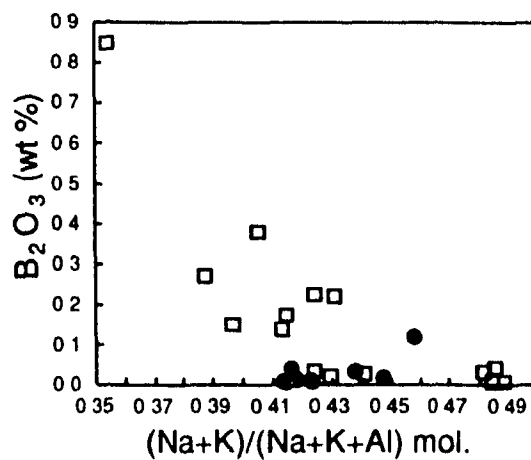
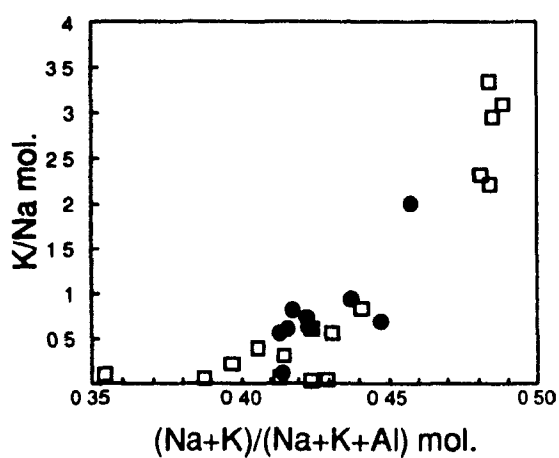


Figure 9. Compositions of aplite (solid circles) and pegmatite (open squares) expressed in terms of molar K/Na, wt % B_2O_3 and TiO_2 , and ppm Nb versus molar $(Na+K)/(Na+K+Al)$.



feldspar and perthite grains within units are late in the paragenetic sequence we conclude that the graphic and coarse-grained K-feldspar units are also late. However, the banded nature of the pegmatite suggests that the crystallization sequence may have been repetitive. Although melt compositions differ from whole rock analyses of pegmatite samples, it is significant that the most alkaline rocks are late. This is similar to the "liquid line of descent" proposed by London *et al.* (1989), whereby pegmatite evolution was modelled by decreasing SiO_2 and increasing alkali elements. However, increasing Na/K ratios after the onset of K-feldspar crystallization in that study are not observed at Nong Sua.

Boron is inversely proportional to the alkali:aluminum ratio (Fig. 9) which confirms the petrographic observation that tourmaline is associated with plagioclase-rich rocks. Figure 9 shows similar trends for TiO_2 and Nb which suggests that these elements are controlled largely by tourmaline abundance. In addition, Fe_2O_3 , F and Li are strongly correlated with B_2O_3 concentration indicating that these elements are also controlled, at least partly by tourmaline (tourmaline pegmatite also contains primary muscovite).

Distribution of Tin

It is possible to evaluate the behaviour of tin during granite evolution and the effects of subsolidus alteration and remobilization of tin by plotting the concentration of Sn versus parameters such as Ti and Rb/Sr, which can in turn be

related to fractionation (Lehmann 1990). We have determined these parameters and compared their distributions to those of Lehmann & Mahawat (1989), who investigated the chemistry of the granites in the western Thai belt. Analyses of three aplite samples that contain trace disseminated cassiterite were excluded.

During plagioclase fractionation most granitic suites contain progressively less Sr and more Rb. At Nong Sua Sr abundances are low, reflecting a high degree of fractionation, and are independent of whole-rock composition. The melt-feldspar partition coefficients for Rb are in favour of K-feldspar over plagioclase (Smith 1974). This is reflected by the strong correlation between Rb and K_2O on Figure 10. The partitioning of Sr between melt and K-feldspar and plagioclase is less consistent but in some studies of rhyolites these partition coefficients have been shown to be similar, *e.g.* Crecraft *et al.* (1981). The Rb-Sr relationships on Figure 10 therefore reflect the crystallization of different plagioclase:K-feldspar ratios in the melt. Similar Rb-Sr relationships are reported for pegmatites elsewhere (*e.g.*, Černý *et al.* 1984; Walker *et al.* 1986; Trumbull, 1991). The strong correlation between Rb and K_2O and between $\log Sr$ and $\log Rb/Sr$ suggests that the Rb-Sr distribution has not been affected by alteration. The $\log Sn$ versus $\log Rb/Sr$ and $\log Ti$ diagrams show positive and negative correlations, respectively, similar to those proposed by Lehmann & Mahawat (1989) but at lower Sn abundances (Fig 10). The small degree of scatter suggests that significant hydrothermal remobilization of tin did not occur. In unaltered S-type granite suites tin behaves

as an incompatible element, and the highest tin concentrations occur in the most fractionated rocks (Lehmann 1990). This relationship may also apply to the internal development of pegmatites, *i.e.*, the most evolved rocks at Nong Sua contain the highest tin concentrations.

Figure 11 shows the distribution of Sn versus different major elements. Tin is positively correlated with Al_2O_3 and K_2O and is negatively correlated with SiO_2 . Similar diagrams were constructed for B_2O_3 and F; these elements do not correlate with tin and thus did not affect tin distribution. The Al_2O_3 - K_2O -rich samples are mainly composed of K-feldspar and quartz and lack tourmaline and muscovite, which occur in samples with granitic or plagioclase-rich compositions. The distribution of Sn in non-mineralized aplite-pegmatite is therefore apparently related to K-feldspar abundance. K-feldspar abundance correlates with the whole rock alkali:aluminum ratio and this ratio also affects the solubility of tin in granitic melts (discussed below). Therefore the molar $(\text{K}+\text{Na})/(\text{K}+\text{Na}+\text{Al})$ ratio was plotted against Sn abundance (Fig. 11). The most peraluminous rocks have the lowest Sn concentrations and the most alkaline (K-rich) have the highest tin abundances. We suggest that tin behaved as an incompatible element in Nong Sua melts and that the concentration of tin in the magma increased during crystallization. The higher tin values in K-feldspar-rich rocks are consistent with magmatic evolution to more alkaline compositions, and with the observation that K-feldspar is paragenetically late. The systematic variation of tin in aplite and pegmatite also indicates that

Figure 10. Compositions of unmineralized aplite (solid circles) and pegmatite (open squares). The reference lines of Lehmann and Mahawat (1989) are regressed from analyses of granites in western Thailand and the tin barren versus tin bearing fields are also from that study.

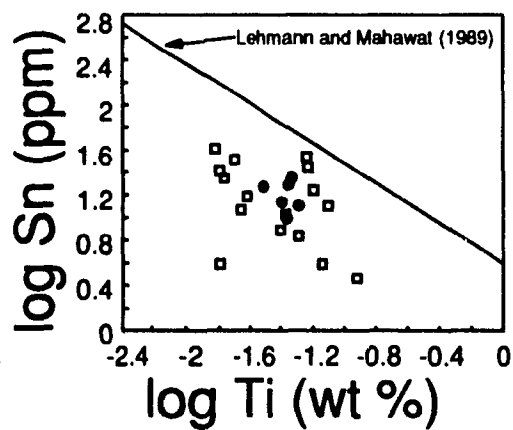
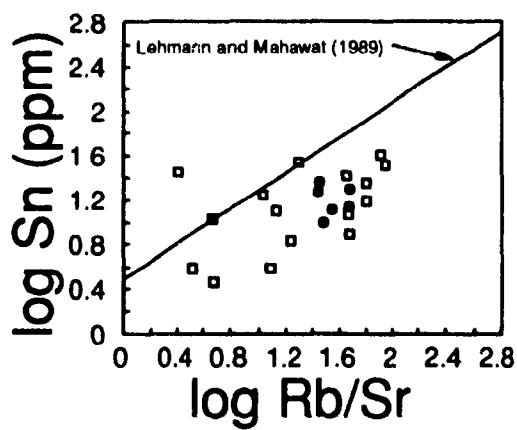
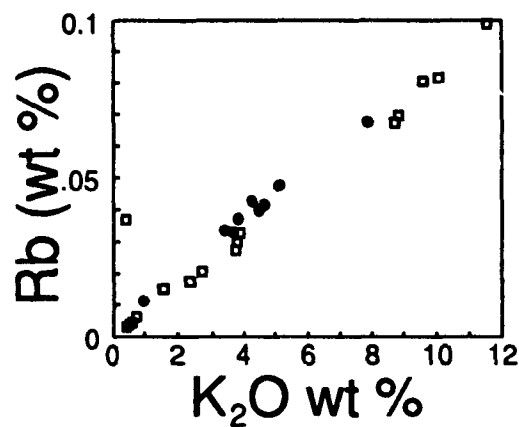
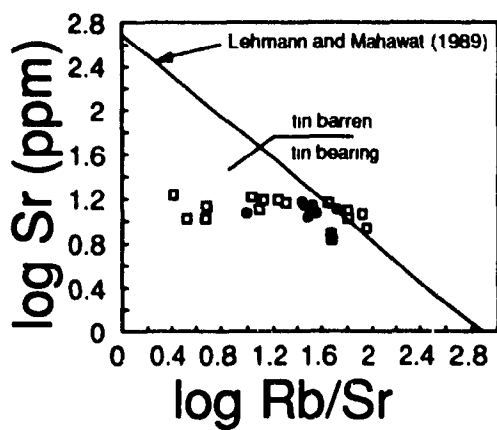
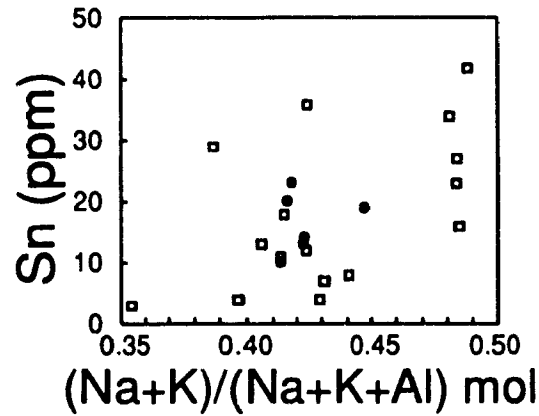
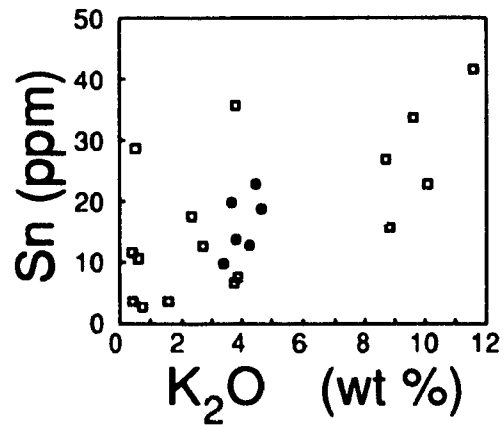
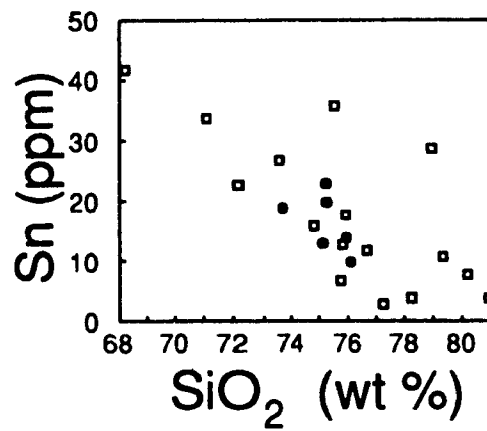
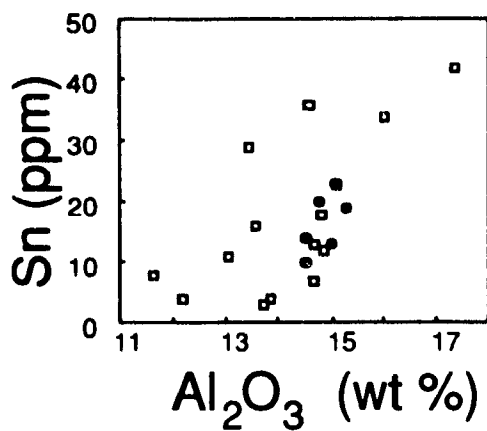


Figure 11. Distribution of tin in unmineralized aplite (solid circles) and pegmatite (open squares).



significant tin was not remobilized as a result of alteration of silicates or oxides. This is supported by the lack of geological or petrographic evidence of pervasive hydrothermal alteration.

Although magmatic processes clearly controlled the distribution of tin in the non-mineralized aplite and pegmatite, it remains to be shown whether or not the tin abundances at Nong Sua are consistent with cassiterite saturation, in light of the available data on tin solubility in granitic magmas. The latter are mainly summarized in Štemprok (1990) who concluded that the dominant control of Sn solubility is temperature. By contrast, Naski & Hess (1985) and Taylor (1988) concluded that the dominant controls of tin solubility in melts are the alkali to aluminum and K/Na ratios. Both studies also suggested alkali-Sn complexing of the type that has been proposed for other high field strength elements such as Ti (Dickinson & Hess 1985). Although the effect of melt structure on mineral-melt partition coefficients is not known, the strong correlation between K_2O and Sn at Nong Sua (high alkali:aluminum) indicates that such complexing may be important.

According to the solubility data of Štemprok (1990), approximately 1000 and 2400 ppm Sn are required to saturate a melt in cassiterite at 600° and 700°C, respectively. The experimental data of Taylor (1988) predict somewhat different cassiterite solubility. At 700° to 800°C, 2 to 3 kbar, and fO_2 at FMQ to FMQ + 1.5 log units, cassiterite saturation occurs at 400 to 2500 ppm Sn. SnO_2 solubility in these experiments increased with increasing (Na+K)/Al and Na/K ratios and with

decreasing fO_2 . All the cassiterite saturation values reported above are much higher than the Sn contents of aplites and pegmatites at Nong Sua (Table 1). However, the whole rock values are dependant on the mineral composition and the mineral-melt partition coefficients for the minerals, *i.e.*, since tin was concentrated by fractionation the bulk distribution coefficient is <1 and the tin concentration in the rock is less than the melt from which it crystallized.

Experimental studies of mineral-melt partition coefficients for tin are unfortunately lacking. However, based on empirical data, Kovalenko *et al.* (1988) proposed temperature dependent partition coefficients. For the minerals observed at Nong Sua, the partition coefficients are $K_{\text{mica}}^{\text{Sn}} > 1 > K_{\text{plagioclase}}^{\text{Sn}} > K_{\text{K-feldspar}}^{\text{Sn}} > K_{\text{quartz}}^{\text{Sn}}$. Using these partition coefficients and whole-rock compositions it is possible to estimate the concentration of tin in the melt at Nong Sua. The pegmatites with the highest tin contents are composed of ≥ 98 % perthite + quartz (Fig. 11) and have an average normative quartz content of 28 wt %. The average tin content of these pegmatites is 22 ppm (Table 1) and individual samples range up to 42 ppm. The tin content of the melt was calculated using normative quartz and perthite (Or+An+Ab) and their respective partition coefficients at 600° and 700°C (Kovalenko *et al.* 1988) assuming that all the tin is contained in quartz and K-feldspar (indicated by the K_2O -Sn correlation). Concentrations of Sn estimated to have been in the melt were 391 and 253 ppm at 600° and 700°C, respectively, for the average K-feldspar-rich pegmatite. These concentrations are below the

cassiterite saturation values of Taylor (1988) and Štemprok (1990), suggesting that during much of the pegmatite evolution cassiterite was not saturated. The sample with the highest tin content (42 ppm; 13 % normative quartz) is estimated to have been in equilibrium with a melt containing 667 ppm Sn at 600°C and 444 ppm at 700°C. These values are in the range of saturation values quoted by Taylor (1988) and close to those of Štemprok (1990), which is consistent with our earlier conclusion that, based on textural evidence, the pegmatite melt was locally saturated with cassiterite.

It is also important to note that the lowest tin abundances are in the most peraluminous rocks containing tourmaline and muscovite. In light of the relative order of the above mineral melt partition coefficients, the peraluminous rocks must have crystallized from a melt with much lower Sn concentrations. This supports the interpretation that peraluminous rocks formed early and were followed by more alkaline rocks. The fractionation required for a tin concentration of 700 ppm in the melt can be estimated using the Rayleigh fractionation equation:

$$C_1 = F^{(D-1)} \times C_0$$

where C_1 is the concentration in the fractionated melt, C_0 is the initial melt concentration, D is the bulk solid-melt distribution coefficient, and F is the proportion of melt remaining. Bulk distribution coefficients of 0.1 to 0.3 are probably reasonable for Nong Sua (*cf.* Coetzee & Twist, 1989). The aplites have an average tin content of 17 ppm (Table 1), but since $D < 1$ this is not the initial

concentration. If we assume the initial melt contained 50 ppm Sn, 95 to 98 % fractionation is required to attain 700 ppm Sn in the melt ($D = 0.1$ to 0.3). Using an initial melt composition of 100 ppm Sn, 88 to 94 % fractionation is required. It therefore follows that the crystallization of cassiterite at Nong Sua probably occurred after >90 % of the crystallization of the aplite-pegmatite was completed.

An additional consideration is the effect of vapour saturation of the melt on the distribution of tin. The occurrence of vein-hosted and contact-style mineralization is evidence that hydrothermal fluids contained significant tin. The lack of alteration of aplite and pegmatite indicates that these fluids did not leach tin from the wallrock, which implies that tin was partitioned into the orthomagmatic fluid at vapour saturation. However, even if $K_{\text{fluid-melt}}^{\text{Sn}}$ is less than 1.0, hundreds of ppm Sn could still be in solution, given that the melt contained in the order of 700 ppm Sn. Unfortunately the partitioning of tin between vapour and melt in granitic systems is poorly understood. Taylor (1988) found that the vapour-melt partition coefficient for tin ($K_{\text{fluid-melt}}^{\text{Sn}}$) derived from experimental studies is proportional to the square of the aqueous chloride molality and that $K_{\text{fluid-melt}}^{\text{Sn}}$ is greater than 1.0 if the chloride concentration is greater than approximately 1 molar. Durasova *et al.* (1986) also found that $K_{\text{fluid-melt}}^{\text{Sn}}$ is proportional to chloride concentration, but estimated that the $K_{\text{fluid-melt}}^{\text{Sn}}$ for a 1 N Cl solution coexisting with haplogranite is 0.2, at 750°C and 1.5 kbar, at the Ni-NiO buffer. The partition of tin into low salinity aqueous fluids thus remains unresolved. At Nong Sua such

low salinity fluids are interpreted to be of orthomagmatic origin (<1 M; Linnen & Williams-Jones, 1991) and therefore tin was probably not strongly partitioned into this fluid. Based on the textural and whole-rock chemical evidence we conclude that cassiterite crystallized directly from melt prior to or at vapour-saturated conditions.

CONCLUSIONS

Crystallization of the Nong Sua complex was commenced with the formation of aplite with a granitic composition. Pegmatite later crystallized to form units with strongly contrasting compositions. K-feldspar-rich units are interpreted to be late, reflecting an evolution toward alkali enrichment and silica depletion of the melt. Tin abundance varies with K_2O suggesting that the tin concentration of the melt increased with fractionation. By using the tin content of K-feldspar-rich samples, experimental tin solubility, and empirical mineral-melt partition data, it has been shown that the melt contained on the order of 700 ppm Sn, which is consistent with the direct crystallization of cassiterite from melt. This supports the textural relationships between cassiterite and feldspar which imply cassiterite crystallization prior to or at vapour-saturated melt conditions; vapour saturation is reflected by myrmekitic plagioclase replacement of K-feldspar, which post-dated disseminated cassiterite mineralization. Mineralization continued after vapour saturation to form vein-hosted cassiterite and wolframite, and contact-style cassiterite. In the case of

the former, the timing of mineralization is constrained by crosscutting relationships to have been at or near solidus conditions of the pegmatite.

ACKNOWLEDGEMENTS

This research was carried out with the aid of a grant from the International Development Research Centre, Ottawa, Canada. We would like to thank S. Bunopas, N. Jungyusuk, S. Khositant, and P. Putthapiban of the Geological Survey Division, Department of Mineral Resources of Thailand for their support and advice during the course of this project. The manuscript was improved by comments from D. Baker and J. Clark.

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CHAPTER 3

MINERALOGICAL CONSTRAINTS ON THE MAGMATIC- HYDROTHERMAL TRANSITION AT THE NONG SUA APLITE- PEGMATITE, THAILAND: IMPLICATIONS FOR TIN-TUNGSTEN MINERALIZATION

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ABSTRACT

The Nong Sua aplite-pegmatite complex is host to disseminated cassiterite and vein-hosted cassiterite-wolframite mineralization. Features such as primary layering in the pegmatite and veins are interpreted in terms of magmatic and hydrothermal processes and, in turn, related to mineralization. Zoned tourmaline crystals in aplite and pegmatite have preserved their primary compositions and overlap with those of hydrothermal origin, suggesting an overlap of P, T, fO_2 and cation activities and a possible continuum between magmatic and hydrothermal crystallization. This in turn suggests that the veins formed while the pegmatite was at or near solidus conditions. Structural relationships support this interpretation. Some hydrothermal tourmaline is enriched in Ca and Mg, indicating that there was late infiltration or diffusion of these elements. Such infiltration-diffusion may have also triggered cassiterite crystallization by increasing the fO_2 of the melt. Textural and compositional data for garnet suggest that it crystallized from melt. Some garnet cores are defined by primary fluid inclusions that are enriched in tin indicating that at vapour saturation, the melt was tin-rich. Columbite, ixiolite and Nb-Ta-rutile are restricted to being aplite-pegmatite-hosted and typically occur as inclusions in disseminated cassiterite, or in the Sn-rich cores in garnet. These oxides are interpreted to have crystallized from melt. Muscovite compositions record a history of post-magmatic re-equilibration down to approximately 400°C.

We conclude that both magmatic (direct crystallization from melt) and orthomagmatic (crystallization from magma-derived aqueous fluids) models of tin-tungsten mineralization are applicable in explaining the origin of pegmatite-hosted deposits. These conclusions have implications for granite-hosted tin-tungsten mineralization elsewhere.

INTRODUCTION

Tin-tungsten mineralization has been linked to the magmatic-hydrothermal transition in some granitic systems, e.g., Eadington (1983), Charoy and Pollard (1989). One of the problems inherent in interpreting this transition is that granites are generally considered to be open systems, i.e., after the host granite has crystallized, it is modified (chemically and texturally) by the fluids of a more differentiated, residual granite, and later interact with fluids of a non-magmatic origin (cf. Marignac and Cuney, 1991). However, the magmatic and hydrothermal evolution of granitic pegmatites more closely approximate closed-system behaviour (Jahns and Burnham, 1969; Taylor *et al.*, 1979). Thus pegmatites are potentially better-suited than granites for an investigation of the relationship between the magmatic-hydrothermal transition and tin-tungsten mineralization. However, tin-bearing pegmatites have received very little scientific study (some exceptions are Gouanvic and Gagny, 1987; Foord and Cook, 1989). The western Thai granite belt is well known for containing pegmatite-hosted cassiterite deposits (Garson *et al.* 1975; Manning, 1986), therefore a study of a cassiterite-wolframite-bearing pegmatite at Nong Sua, in western Thailand, was undertaken to evaluate the importance of the magmatic-hydrothermal transition to tin-tungsten deposition.

The timing of the mineralization has been constrained in our previous work (Linnen and Williams-Jones, submitted; Chapter 2); cassiterite is interpreted to have

crystallized directly from melt and from aqueous fluids. In this study the compositions of tourmaline, muscovite, garnet and cassiterite and other oxides from different stages of the pegmatite development are examined. The origin of features such as layering in the pegmatite and veins at Nong Sua are considered to be magmatic or hydrothermal, respectively. This implies a magmatic origin for the primary pegmatite minerals and the compositions of pegmatite- and vein-hosted minerals can be interpreted in terms of the magmatic and hydrothermal evolution. In addition, textural relationships between silicate minerals and cassiterite-wolframite provide the temporal link between the magmatic-hydrothermal transition and Sn-W mineralization, and thereby provide constraints on the mechanisms of cassiterite and wolframite deposition.

GEOLOGICAL SETTING

Cassiterite and wolframite are hosted by a rhythmically layered aplite-pegmatite complex at Nong Sua, located approximately 180 km south of Bangkok, within the western Thai granite belt (cf. Cobbing *et al.* 1986). The age of the complex is 63.9 ± 0.6 Ma based on a total fusion $^{40}\text{Ar}/^{39}\text{Ar}$ date of muscovite by Charusiri (1989). The complex comprises laterally continuous horizons of garnet aplite, tourmaline aplite, banded pegmatite, and tourmaline pegmatite, consisting of quartz, K-feldspar, and plagioclase $\pm$ muscovite $\pm$ tourmaline $\pm$ garnet $\pm$ apatite. The normative compositions of the aplites are close to that of the granite minimum,

whereas those of the pegmatites range from albite-quartz-rich to orthoclase-rich (Linnen and Williams-Jones, submitted; Chapter 2). Garnet is restricted to the garnet aplite and banded pegmatite and is the principal feature that permits the two aplite units to be distinguished. The banded pegmatite has been subdivided into five lithotypes, based on mineralogy and texture. Tourmaline and muscovite are most concentrated in plagioclase-rich units or in units with a granitic composition, and are rare or absent in K-feldspar-rich units (op. cit.).

The mineralization consists of cassiterite disseminated in pegmatite and to a lesser extent in aplite; vein-hosted wolframite and cassiterite; and cassiterite intergrown with muscovite at the contacts between pegmatite and metasedimentary rocks (contact mineralization). Feldspar-cassiterite textures suggest that the crystallization of disseminated cassiterite locally preceded that of plagioclase, K-feldspar and myrmekite. This is interpreted to reflect crystallization from melt at, or prior to, vapour-saturation (Linnen and Williams-Jones, submitted; Chapter 2). The quartz-tourmaline veins that host cassiterite and wolframite are contained within aplite units and terminate abruptly at aplite-pegmatite contacts, suggesting that the veins formed while the pegmatite was above or near solidus conditions (op. cit.). Contact mineralization is characterized by coarse-grained secondary muscovite, and is also considered to have precipitated from hydrothermal fluids (op. cit.).

ANALYTICAL METHODS

Mineral compositions were determined using a Cameca Camabax electron microprobe operated at 15 kV, 8 to 10 nA, using a beam diameter of 1 to 10 μm , depending on the mineral analyzed. The stoichiometry of tourmaline was calculated assuming 31 (O,OH,F), 3 boron atoms, and total iron as Fe^{+2} . The calculated stoichiometry of Si is 6.0 ± 0.15 . The Z-site was filled with 6.0 cations of Al after the deficiency of Si was taken into account, and the remaining Al was assigned to the Y-site (cf. Gallagher, 1988). The lithium contents of tourmaline were not determined; however, the average whole-rock Li abundance in tourmaline-rich (and muscovite-rich) pegmatite is only 28 ppm (Linnen and Williams-Jones, submitted; Chapter 2). Lithium is therefore unlikely to be a major constituent of the tourmaline or muscovite. The stoichiometry of muscovite was calculated assuming total iron as FeO, and 24 (O,OH,F). The stoichiometry of garnet was calculated assuming total iron as FeO, and 12 oxygens. The stoichiometric Al+Ti in garnet is 2.0 ± 0.05 and $\text{Fe}^{+2}+\text{Mn}+\text{Mg}+\text{Ca}$ is 3.0 ± 0.1 , indicating that the andradite (Fe^{+3}) component is negligible. The stoichiometries of the Ti-Fe-Nb-Ta oxides were calculated using the formulas listed in Foord (1982).

TOURMALINE

Tourmaline in the pegmatite is generally coarse-grained, euhedral, and zoned

with olive cores and brown rims. Blue tourmaline is rarely present either as growth bands in brown tourmaline or as haloes around fractures cutting olive-brown tourmaline. Tourmaline in aplite is typically fine-grained, anhedral to subhedral, commonly poikilitic, and is disseminated either as individual grains or as grain clusters. The colour and zoning of this tourmaline are similar to that in pegmatite, although olive cores are less common. Variations in tourmaline abundance delineate individual layers in the banded pegmatite unit and locally tourmaline is oriented perpendicular to layering. These features indicate a magmatic origin for the pegmatite-hosted tourmaline, implying that the zoning in this mineral reflects physicochemical changes in the melt. The compositions of tourmaline in aplite are similar to those of the tourmaline in pegmatite (discussed below) indicating that the aplite tourmaline is also magmatic in origin.

Tourmaline associated with disseminated cassiterite (mineralized) or contained within quartz $\pm$ cassiterite $\pm$ wolframite veins differs from that hosted by unmineralized aplite and pegmatite by commonly containing fluid inclusions and blue tourmaline growth bands, and by rarely having olive cores. There is a strong spatial and temporal relationship between disseminated and vein-hosted cassiterite, and tourmaline. This is exemplified in a sample of disseminated mineralization by a cassiterite growth zone that is defined by tourmaline inclusions (Fig. 1). The origin of mineralized tourmaline is equivocal. Regular contacts with all other pegmatite minerals suggest a magmatic origin but the abundance of fluid

TABLE 1
REPRESENTATIVE TOURMALINE COMPOSITIONS

	1	2	3	4	5	6	7	8
FeO	12.82	12.28	12.62	12.46	13.92	9.43	12.39	9.86
MnO	0.20	0.15	0.18	0.21	0.29	0.35	0.19	0.07
TiO ₂	0.43	0.84	0.29	0.76	0.42	0.86	0.94	0.03
Na ₂ O	1.79	1.93	1.42	1.83	2.03	2.12	1.81	1.90
Al ₂ O ₃	33.84	34.26	35.02	34.18	32.61	31.88	33.96	33.06
SiO ₂	35.95	35.65	36.97	35.98	36.25	36.67	35.24	37.80
F	0.55	0.48	0.22	0.59	0.88	1.15	0.34	0.26
MgO	1.93	2.26	1.11	1.53	1.77	5.50	2.00	5.03
K ₂ O	0.04	0.07	0.04	0.06	0.06	0.05	0.06	0.01
CaO	0.23	0.30	0.10	0.38	0.26	0.91	0.33	0.20
-O=F	0.23	0.20	0.09	0.25	0.37	0.48	0.14	0.11
Tot.	87.55	88.01	87.89	87.72	88.11	88.46	87.12	88.12
Na	0.577	0.618	0.450	0.589	0.660	0.678	0.585	0.596
Ca	0.041	0.053	0.018	0.068	0.047	0.161	0.059	0.035
K	0.008	0.015	0.008	0.013	0.013	0.011	0.013	0.002
Si	5.980	5.887	6.046	5.968	6.075	6.045	5.872	6.111
Al Z	6.020	6.113	5.954	6.032	5.925	5.955	6.128	5.889
Al Y	0.615	0.554	0.796	0.651	0.515	0.239	0.540	0.410
Fe	1.784	1.696	1.726	1.729	1.951	1.300	1.726	1.333
Mg	0.479	0.556	0.271	0.378	0.442	1.351	0.497	1.212
Mn	0.028	0.021	0.025	0.030	0.041	0.049	0.027	0.010
Ti	0.054	0.104	0.036	0.095	0.053	0.107	0.118	0.004
F	0.289	0.251	0.114	0.310	0.466	0.600	0.179	0.133

1 olive tourmaline core from aplite, 2 brown tourmaline rim (same grain as 1), 3 olive tourmaline core from pegmatite, 4 brown tourmaline rim (same grain as 3), 5 brown (Fe-rich) "mineralized" tourmaline, 6 brown tourmaline (Mg-rich; same grain as 5), 7 brown tourmaline from vein, 9 late blue tourmaline along fracture

Figure 1. Photomicrograph of tourmaline (To) inclusions in cassiterite (Cas) delineating a growth band.



inclusions alludes to an involvement of hydrothermal fluids. Therefore we tentatively interpret these textures to reflect crystallization from a vapour-saturated melt. Vein tourmaline, based on its mode of occurrence, is interpreted to have precipitated from hydrothermal fluids.

Olive cores in aplite- and pegmatite-hosted tourmaline have a restricted range of compositions. The $\text{MgO}/(\text{MgO}+\text{FeO}+\text{MnO})$ weight ratio ranges from 0.12 to 0.21. The cores are richer in FeO and poorer in TiO_2 compared to the brown tourmaline rims (Fig. 2 and Table 1). This contrasts with the compositions of other tourmaline from Thailand (including Nong Sua) in Manning (1982) where TiO_2 and MgO are reported to be enriched and FeO is depleted in the cores. However, the range of magmatic tourmaline compositions determined in this study is similar to that of Manning (1982). Mineralized and vein tourmaline have compositions that range from those of magmatic origin to being enriched in Mg (Fig. 2). The restricted range of composition of magmatic tourmaline and a minor enrichment of Mg and Ca in the rims are evident in a representative profile from the core to the rim of a tourmaline crystal in pegmatite (Fig. 3). Mineralized tourmaline contrasts with other magmatic tourmaline by displaying a much wider range of compositions and more complex zoning (Fig. 4 and Table 1). However, for both simple and complex zoning there is a strong correlation between Ca and Mg and a negative correlation of these elements with excess Al (Al_V) (Figs 3 & 4).

The composition of tourmaline deviates significantly from the schorl-dravite

Figure 2. Compositions of tourmaline in terms of weight percent TiO_2 , FeO + MnO , and MgO . Solid circles represent olive, stars brown, open circle blue growth bands and solid triangles late blue tourmaline along fractures.

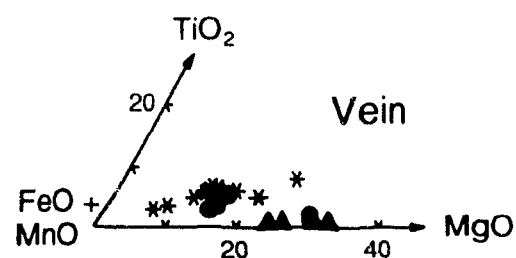
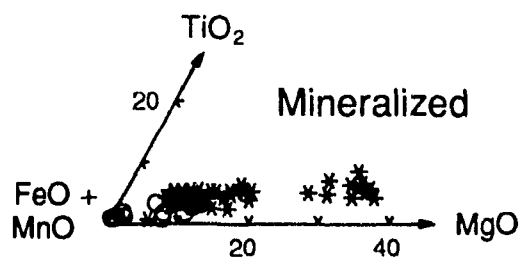
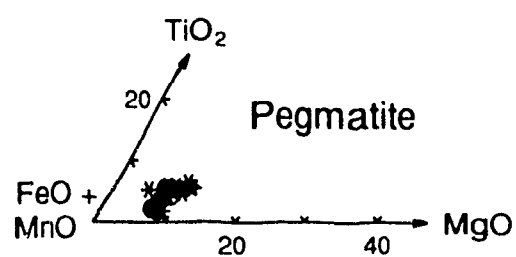
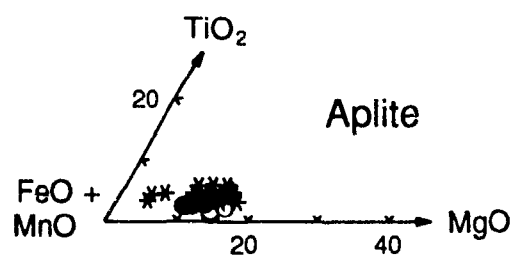
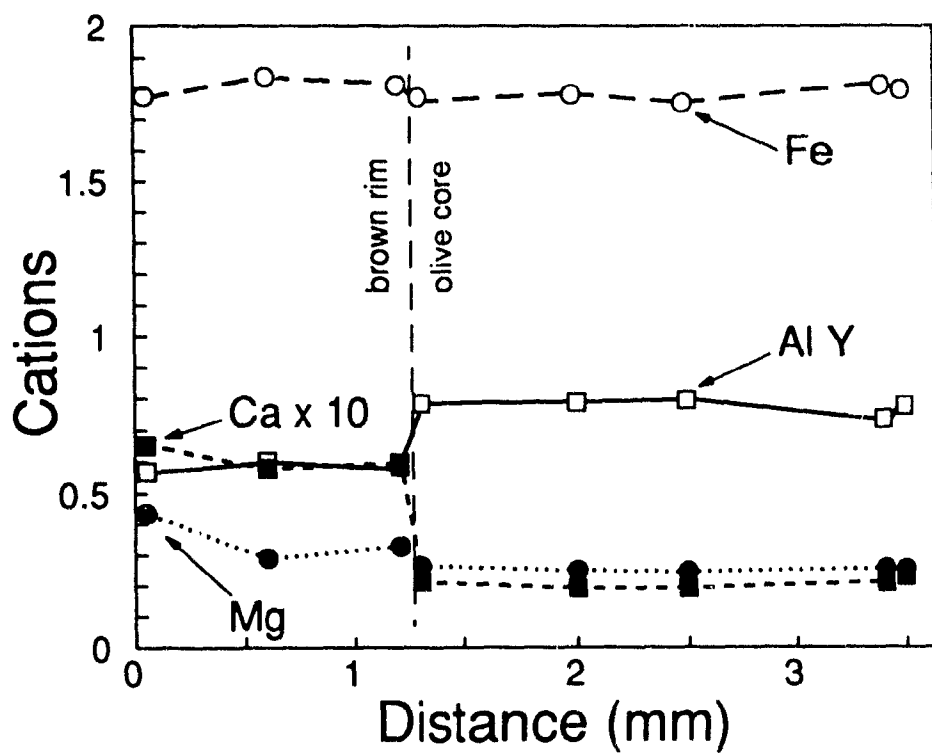


Figure 3. Results of an electron microprobe traverse from the core to rim of a zoned primary tourmaline crystal in pegmatite showing the cation proportions of Fe (open circles), Mg (solid circles), Al_V (open squares) and Cax10 (solid squares).



series. This is made evident by plotting $R^+(Na + 2Ca)$ and $R^{2+}(Mg + Mn + Fe)$ versus $R^{3+}(Al + 4/3 Ti)$ (Fig. 5) where $R^+R^{2+}_3R^{3+}_6(BO_3)_3Si_6O_{18}(OH,F)_4$ represents the formula of the schorl-dravite series. The magmatic tourmaline at Nong Sua has a restricted range of $R^+ + R^{2+}$ values, from approximately 2.5 to 3.2, and R^{3+} values, from approximately 6.5 to 7.0. Mineralized and vein tourmaline range from these values to a maximum $R^+ + R^{2+}$ value of 3.8 and a minimum R^{3+} value of 6.3 (Fig. 5). The deviation from the schorl-dravite series is due primarily to excess Al (Y-site) and deficiency in the total alkali content. This is probably the result of coupled substitutions such as the proton-deficient and alkali-free substitutions proposed by Foit and Rosenberg (1977): $(OH)^- + R^{2+} = R^{3+} + O^{2-}$; $R^+ + R^{2+} = R^{3+} + \square$; where $\square$ represents a site vacancy.

Magmatic tourmaline generally contains higher excess Al, up to 0.8 cations, and lower Na+Ca, as low as 0.45 cations, than the mineralized and vein tourmaline. The high excess Al of magmatic tourmaline is a common feature of pegmatites and in some cases correlates with magmatic evolution, e.g., Jolliff *et al.* (1986). Increases of excess Al in tourmaline have been interpreted by several authors, including Jolliff *et al.* (1986), to reflect decreasing temperature. The observation that excess Al in magmatic tourmaline decreases from core to rim and that younger hydrothermal tourmaline have lower excess Al suggests that the Al contents of the tourmaline at Nong Sua are not related to cooling. It is significant that the tourmaline-rich pegmatite is interpreted to have crystallized from the most

Figure 4. Results of an electron microprobe traverse of a zoned "mineralized" tourmaline crystal in pegmatite showing the cation proportions of Fe (open circles), Mg (solid circles), Al_Y (open squares) and Cax10 (solid squares).

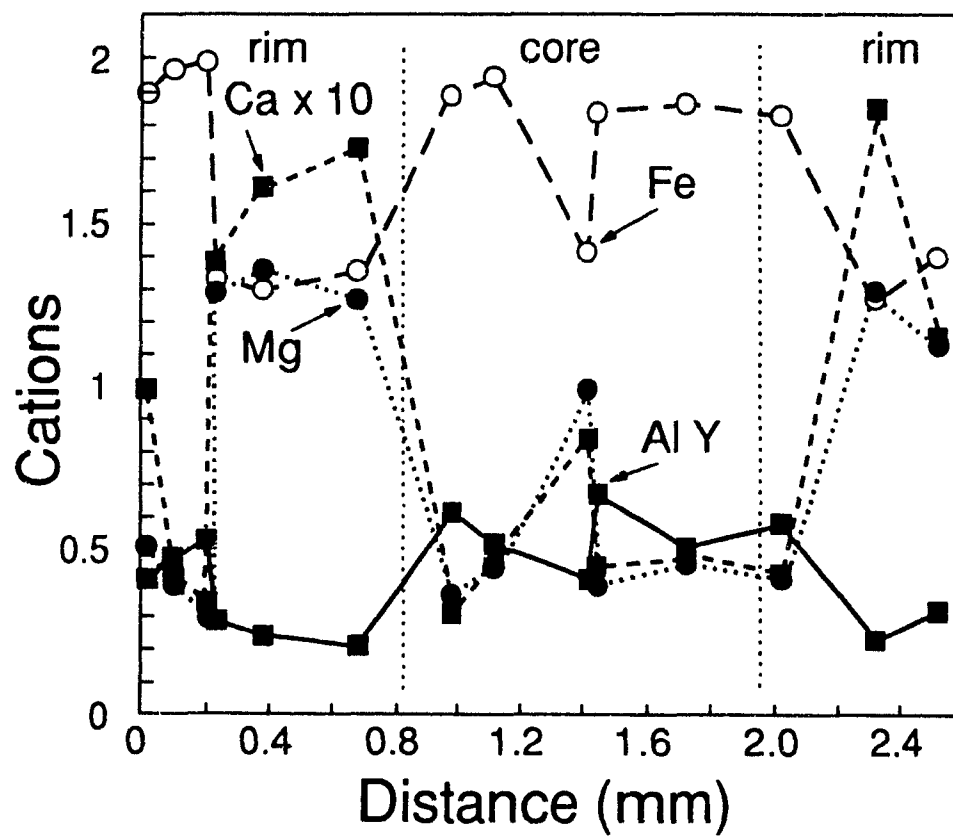
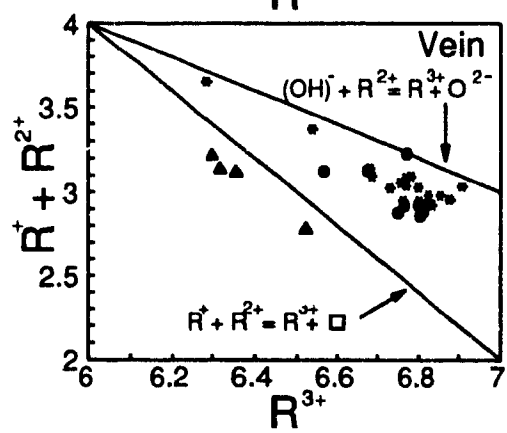
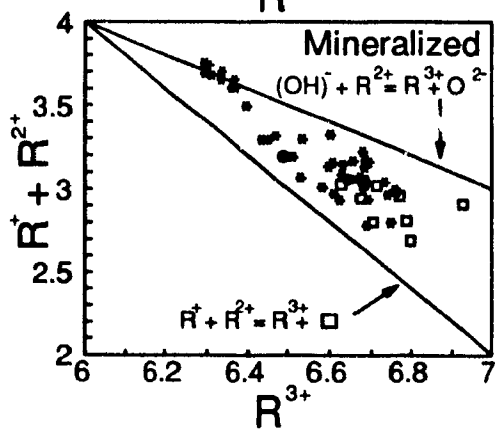
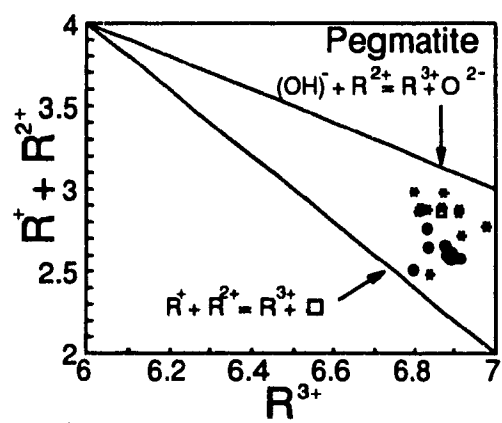
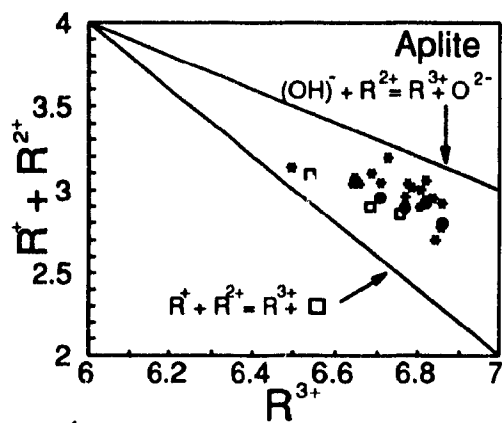


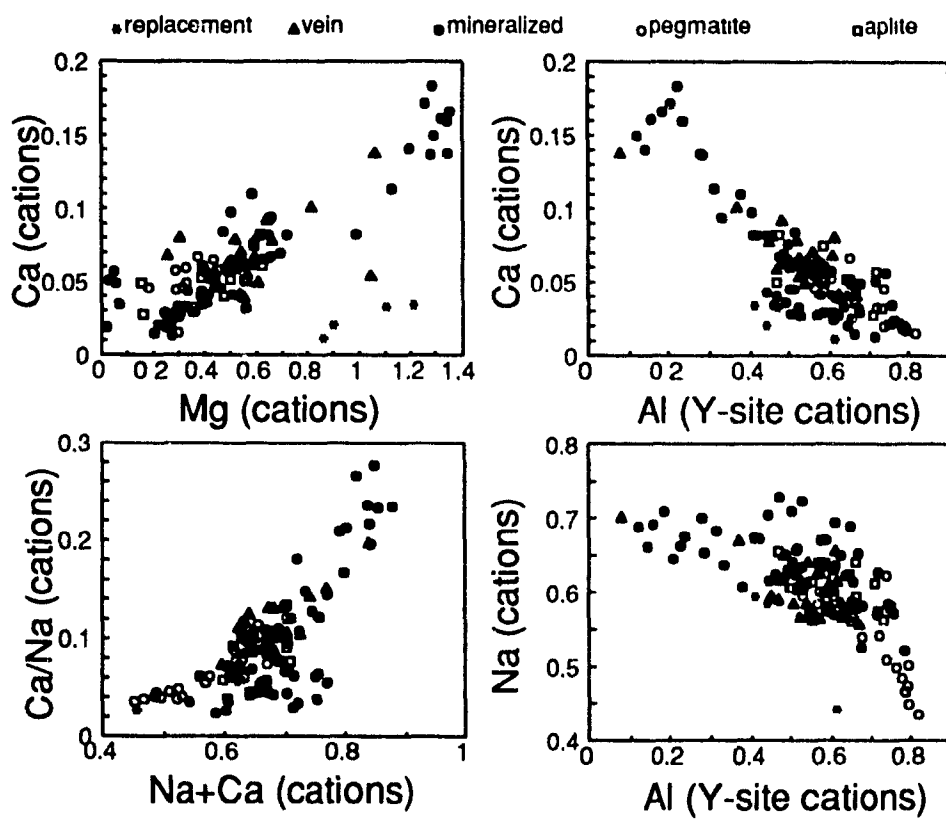
Figure 5. Compositions of tourmaline in terms of R^+ , R^{2+} , and R^{3+} cations where $R^+ = Na + 2Ca$, $R^{2+} = Mg + Fe + Mn$, and $R^{3+} = Al + 4/3 Ti$. The tourmaline stoichiometry was calculated assuming 31 (O,OH,F), and 3 B. Solid circles represent olive tourmaline, stars brown tourmaline, open squares blue growth bands and solid triangles late blue tourmaline along fractures.



peraluminous melt at Nong Sua and that the tourmaline in this unit is also the most peraluminous, whereas aplite and its tourmaline are both less peraluminous (Linnen and Williams-Jones, submitted; Chapter 2). These observations suggest that the compositions of magmatic tourmaline at Nong Sua were potentially controlled by the alkali:aluminum ratio of the melt.

Vein and mineralized tourmaline range from magmatic compositions to those with higher Ca, Na, Ca/Na and Mg and lower Al_Y (Fig. 6). The composition of the mineralized tourmaline inclusions in cassiterite (Fig. 1) is nearly identical to that in the brown rims of magmatic tourmaline from pegmatite and aplite which suggests coeval crystallization of tourmaline and cassiterite from the melt ($\pm$ vapour). The MgO contents of aplite and pegmatite are ≤ 0.03 wt % (Linnen and Williams-Jones, submitted); tourmaline is the only mineral that contains significant MgO and; tourmaline is richer in Mg than the coexisting melt (Benard *et al.* 1985). The Ca-Mg enrichment in the rims of magmatic tourmaline (Fig. 3) are therefore probably not the result of crystallization in a closed system. London (1990) proposed that tourmaline crystallization in pegmatites results from infiltration of bivalent cations from the wallrock into the pegmatite via aqueous or aqueous-carbonic fluids. We suggest that the core to rim increases of Ca and Mg abundances in magmatic and mineralized tourmaline at Nong Sua also reflect Ca-Mg metasomatism from an external fluid source. The much greater compositional variation in mineralized tourmaline compared to primary tourmaline may be

Figure 6. Compositions of tourmaline in terms of molar Ca, Na, Ca/Na, Na+Ca, and Al_y. Tourmaline stoichiometry was calculated assuming 31 (O,OH,F), and 3 B. Open circles represent tourmaline from pegmatite, open squares from aplite, solid circles "mineralized" tourmaline, solid triangles from tourmaline from veins and stars late blue tourmaline along fractures.



reflection of vapour saturation, i.e., much slower diffusion rates of cations in vapour-undersaturated melt relative to that where there is infiltration and fluid mixing in vapour-saturated melt. In the case of veins, the overlap of compositions is evidence that the veins formed from aqueous fluids while the pegmatite was at vapour-saturated conditions. The enrichment of Ca and Mg in vein tourmaline suggests precipitation in the presence of mixed orthomagmatic and external fluids.

The youngest tourmaline is blue and occurs as haloes around fractures in older magmatic and hydrothermal tourmaline. Analyses of this tourmaline are included with the vein data set in Figures 2, 5 and 6. The composition of blue replacement tourmaline contrasts with that of other tourmaline types; it has low Al, Ti, and Ca/(Ca+Na) ratio, and high Mg contents (Table 1). The composition of this tourmaline does not overlap with those of magmatic or early hydrothermal origin. The occurrence of blue tourmaline in haloes and its contrasting compositions indicate a late hydrothermal origin and crystallization from fluids of markedly different temperature, pressure or composition.

There are very few studies in which both magmatic and hydrothermal tourmaline compositions have been determined. Two such studies are of tourmaline associated with the tin granites of Alaska (Swanson et al. 1988 and 1990) and in the Larderello geothermal field (Cavarretta and Puxeddu, 1990). In both these studies the magmatic tourmaline is Fe-Al-rich whereas the hydrothermal tourmaline is richer in Ca-Mg with lower Al concentrations. It is important to note that no

compositional overlap between the magmatic and hydrothermal tourmaline was documented in either study. This lack of overlap suggests that tourmaline compositions can be used to discriminate between magmatic and hydrothermal origins (see also Henry and Guidotti, 1985; Clarke et al., 1989). The similarity of magmatic and hydrothermal tourmaline at Nong Sua may therefore reflect a continuum between crystallization from melt and precipitation from fluids.

MUSCOVITE

Muscovite in the pegmatite ranges up to 5 cm across and, where intergrown with coarse-grained plagioclase, displays sharp contacts with no evidence of replacement. Muscovite in aplite is present as books that are generally 1 to 3 mm long. It is typically intergrown with quartz, but may also have idiomorphic contacts with plagioclase, K-feldspar, garnet, and tourmaline. The muscovite textures in both pegmatite and aplite indicate a primary origin (cf. Miller et al., 1981).

Secondary muscovite most commonly occurs as a fine-grained incipient alteration of plagioclase and K-feldspar, developed locally in aplite, pegmatite and veins. This muscovite post-dated cassiterite-wolframite mineralization and is penecontemporaneous with trace scheelite that replaces wolframite along fractures. Muscovite that is characteristic of contact-style mineralization is coarse-grained, generally 5 mm across, occurs with quartz and beryl, has replaced plagioclase, and is intergrown with cassiterite. Secondary muscovite associated with disseminated

mineralization is texturally and chemically similar to that in contact-style mineralization and thus both are grouped together in Figures 7 and 8 and in the discussion below.

Primary muscovite in aplite and pegmatite is enriched in TiO_2 and Na, and depleted in FeO and MgO relative to secondary muscovite (Fig. 7 and Table 2). In this respect it is similar to primary and secondary muscovite observed in peraluminous granites elsewhere, e.g., Miller *et al.* (1981), Monier *et al.* (1984). The relationships between total M^{+2} octahedral cations (Fe+Mg+Mn), F and the paragonite mole fraction ($\text{Na}/(\text{K}+\text{Na})$) and between total octahedral cations and Si^{IV} and Al^{VI} are shown by Figure 8. The primary muscovite has low M^{+2} totals and is enriched in F relative to secondary muscovite. The Si^{IV} and Al^{VI} proportions of primary muscovite are close to those of end-member muscovite, whereas secondary muscovite contains a significant phengite component. The deviation from 1:1 $\text{M}^{+2}:\text{Si}^{\text{IV}}$ substitution (Fig. 8) suggests that the secondary muscovite contains a minor trioctahedral Fe-Mg component (Monier and Robert, 1986). The low and high phengite components in primary and secondary muscovite, suggest formation at relatively high and low temperatures, respectively (Monier and Robert, 1986).

Estimates of the temperature of primary muscovite are provided by the plagioclase-muscovite geothermometer (Green and Usdansky, 1986). Using the average plagioclase and muscovite compositions from electron microprobe analyses and a pressure of 3 kbar (see Linnen and Williams-Jones, 1991) the calculated

Figure 7. Compositions of primary (magmatic) and secondary muscovite in terms of TiO_2 , FeO , and MgO weight percent and Ti, Mg, and Na cation percent. Solid squares represent muscovite from pegmatite, solid circles from aplite, open squares from contact-style mineralization and open circles secondary muscovite. Each data point represents the average of several analyses.

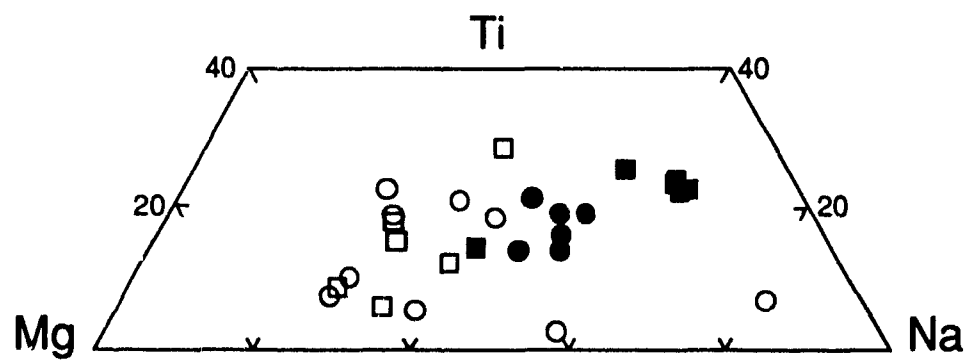
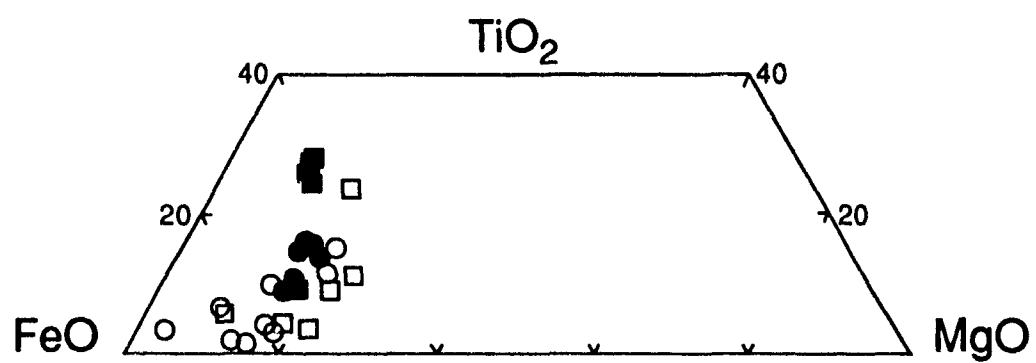
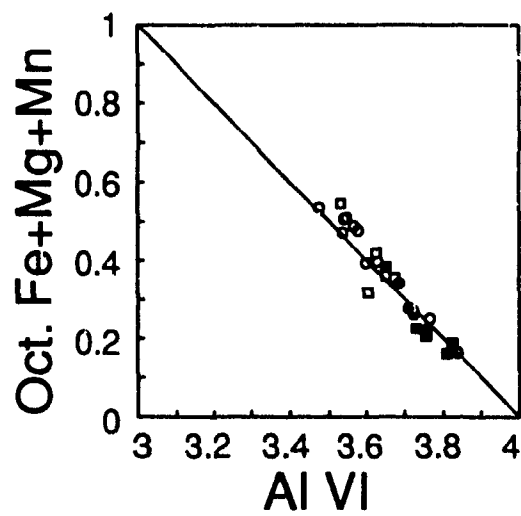
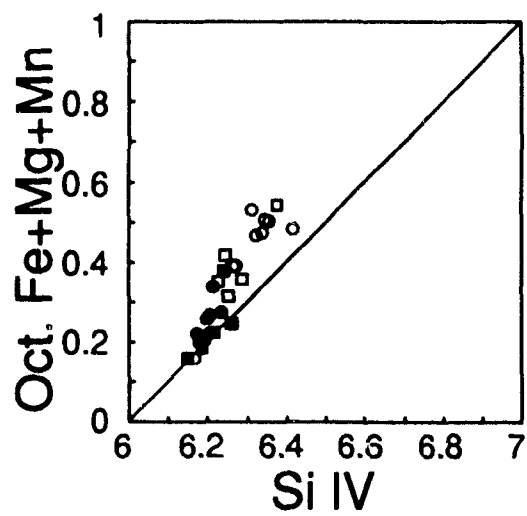
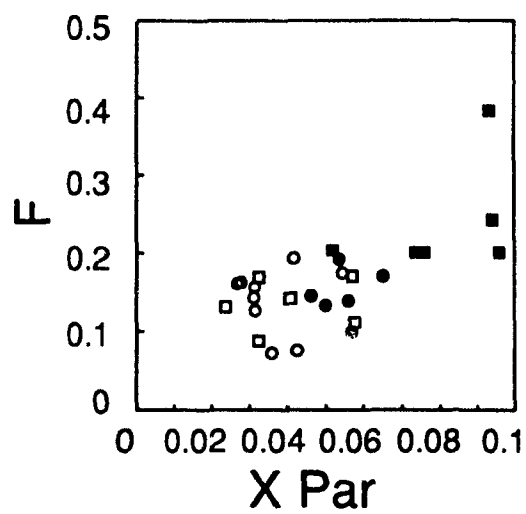
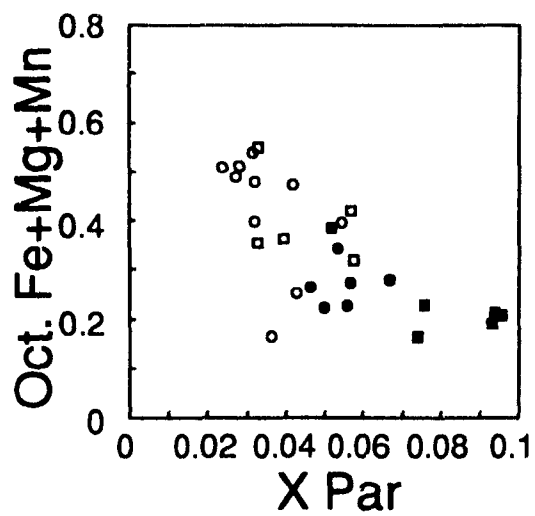


Figure 8. Compositions of primary (magmatic) and secondary muscovite in terms of the paragonite mole fraction ($X_{\text{par}} = \text{Na}/(\text{Na}+\text{K})$) versus the total octahedral cations ($\text{Fe}+\text{Mg}+\text{Mn}$) and stoichiometric F and, the total octahedral cations versus tetrahedral Si and octahedral Al. Solid squares represent muscovite from pegmatite, solid circles from aplite, open squares from contact-style mineralization and open circles secondary muscovite. Each data point represents the average of several analyses.



temperatures are 401° to 489°C (Table 3). These temperatures are below those of the solidi for aplites and pegmatites (Benard et al., 1985) and therefore indicate re-equilibration at lower temperature. The re-equilibration of plagioclase and muscovite involved loss of K and Na, respectively. Possible maximum temperatures were therefore calculated using maximum Or and X_{par} compositions in coexisting plagioclase and muscovite. These temperatures range up to 603°C. We propose that the compositions of primary muscovite were modified by subsolidus exchange with magmatic or external fluids. Secondary muscovite probably also formed from these fluids, but the incipient nature of the replacement of feldspars and the very localized occurrence of contact mineralization suggest low fluid to rock ratios.

GARNET

Garnet is most abundant in the garnet aplite and, less commonly, is present in the banded pegmatite. In the garnet aplite it has two modes of occurrence. It is developed as thin garnet-rich bands, typically <2 cm thick, that are either parallel to layering or cut layering at a shallow angle; the crystals tend to be euhedral and <0.5 mm across. This mode of occurrence is generally developed at aplite-pegmatite contacts. Garnet is also disseminated in the garnet aplite unit where it is present as anhedral to subhedral grain clusters or as individual subhedral to euhedral crystals. Grain clusters are <1 cm across and are made up of individual

TABLE 2
REPRESENTATIVE MUSCOVITE AND GARNET COMPOSITIONS

	1	2	3	4	5	6	7	8
FeO	1.67	1.17	2.50	2.37	15.38	26.47	16.95	24.35
MnO	0.05	0.17	0.12	0.09	26.85	15.90	24.42	17.78
Na ₂ O	0.46	0.86	0.30	0.41	0.04	0.00	0.17	0.00
Al ₂ O ₃	34.60	35.91	33.21	32.96	20.71	20.48	20.47	20.45
SiO ₂	46.87	46.30	46.22	46.65	34.70	37.08	37.04	37.28
MgO	0.45	0.18	0.79	0.76	0.28	0.37	0.07	0.15
K ₂ O	11.30	10.69	11.36	11.13	n.a.	n.a.	n.a.	n.a.
CaO	0.00	0.00	0.00	0.00	0.68	0.54	0.64	0.54
TiO ₂	0.35	0.36	0.43	0.49	0.15	0.00	0.17	0.00
SnO ₂	n.a.	n.a.	n.a.	n.a.	0.17	0.00	0.00	0.00
F	0.44	1.13	0.26	0.72	n.a.	n.a.	n.a.	n.a.
-O=F	0.19	0.48	0.11	0.30				
Tot.	96.00	96.28	95.08	95.27	98.95	100.80	99.76	100.50
K	1.926	1.829	1.962	1.928				
Na	0.119	0.224	0.079	0.108	0.006	0.000	0.000	0.000
Ca	0.000	0.000	0.000	0.000	0.060	0.047	0.055	0.047
Si	6.263	6.210	6.259	6.334	2.908	3.024	3.041	3.043
Al IV	1.737	1.790	1.741	1.666	1.996	1.991	1.958	1.989
Al VI	3.713	3.887	3.560	3.608				
Fe	0.187	0.131	0.283	0.269	1.079	1.805	1.163	1.663
Mg	0.090	0.036	0.159	0.154	0.034	0.043	0.007	0.017
Mn	0.006	0.019	0.014	0.010	1.907	1.099	1.698	1.230
Ti	0.044	0.045	0.055	0.063	0.010	0.000	0.009	0.000
F	0.186	0.479	0.111	0.309				

1 primary muscovite in aplite, 2 primary muscovite in pegmatite, secondary muscovite - contact mineralization, 4 secondary muscovite - incipient alteration of feldspar, 5 garnet core with fluid inclusions, 6 garnet rim (same grain as 5), 7 fluid inclusion-free garnet core, 8 garnet rim (same grain as 7).

TABLE 3
PLAGIOCLASE-MUSCOVITE GEOTHERMOMETRY

Sample	Rock Type	Plag-Musc T _{mean} (°C)	Plag-Musc T _{max} (°C)
PK-17e	pegmatite	476	603
PK-25	aplite	401	451
PK-33	pegmatite	489	553
PK-43a	pegmatite	472	559
PK-58	aplite	413	475

grains that are generally <0.5 mm across, whereas single subhedral to euhedral crystals range up to 8 mm in diameter.

Garnet in the banded pegmatite unit is relatively coarse-grained, typically 4 to 8 mm across, and commonly is euhedral. It is present either as disseminated crystals or in bands parallel to layering. In aplite and pegmatite some crystals of plagioclase and perthite contain euhedral inclusions of garnet and none of the garnet crystals exhibit textures that imply a replacement origin. The occurrence of garnet in layers, as inclusions and its composition (discussed below) are interpreted to reflect a magmatic origin (a similar conclusion was reached by Manning, 1983).

A feature of both aplite- and pegmatite-hosted garnet is that the cores of some crystals are outlined by primary fluid inclusions. Within these cores, solid inclusions of bladed rutile, columbite or ixiolite, and rounded quartz grains may also be present and the rims are free of solid and fluid inclusions. The presence of primary fluid inclusions suggests that this garnet crystallized at vapour-saturated conditions. It could be argued that garnet bands at aplite-pegmatite contacts are a hydrothermal feature. However, the lack of replacement textures suggest that, if the garnet is hydrothermal, it probably crystallized from a fluid that coexisted with melt and crystals.

Garnet in both aplite and pegmatite has compositions of $>.95 X_{\text{Alm}} + X_{\text{Spes}}$ and $<.05 X_{\text{Gros}} + X_{\text{Pyr}}$ where X_{Alm} , X_{Spes} , X_{Gros} , and X_{Pyr} are the almandine, spessartine, grossular, and pyrope mole fractions. The cores are Mn-rich and the rims are Fe-

rich (Fig. 9 and Table 2), are consistent with a magmatic origin (Manning, 1983). Similar compositions are reported in Manning (1983, 1984) for garnet in aplite and pegmatite from Thailand (including Nong Sua). Many, but not all, grains that display Mn-Fe zoning also contain significant tin in their cores (Table 2). The highest tin concentrations occur in cores that are defined by primary fluid inclusions. This is illustrated by data for a representative garnet crystal which contains up to 1500 ppm SnO_2 in its Mn-rich core but in its Fe-rich rim SnO_2 is below the electron microprobe detection limit (approximately 100 ppm) (Fig. 10). The TiO_2 distribution mimics that of SnO_2 , both of which exhibit abrupt decreases outside the core defined by fluid inclusions. Cassiterite mineralization is linked to garnet crystallization by high tin concentrations in garnet cores containing fluid inclusions and by the similarity of the ice melting and homogenization temperatures of these fluid inclusions with those in cassiterite (Linnen and Williams-Jones, 1991). By inference, if garnet is magmatic, then some of the cassiterite is also magmatic.

OXIDE MINERALOGY

Cassiterite and wolframite are the dominant oxide minerals that are present at Nong Sua. Cassiterite occurs as disseminations in pegmatite (and less commonly in aplite) and is also vein-hosted, whereas wolframite is almost exclusively vein-hosted. In addition, rutile, ilmenite, columbite and ixiolite are present in minor to

Figure 9. Compositions of garnet from aplite and pegmatite in terms of cation percent Fe, Mn, and Mg.

Figure 2

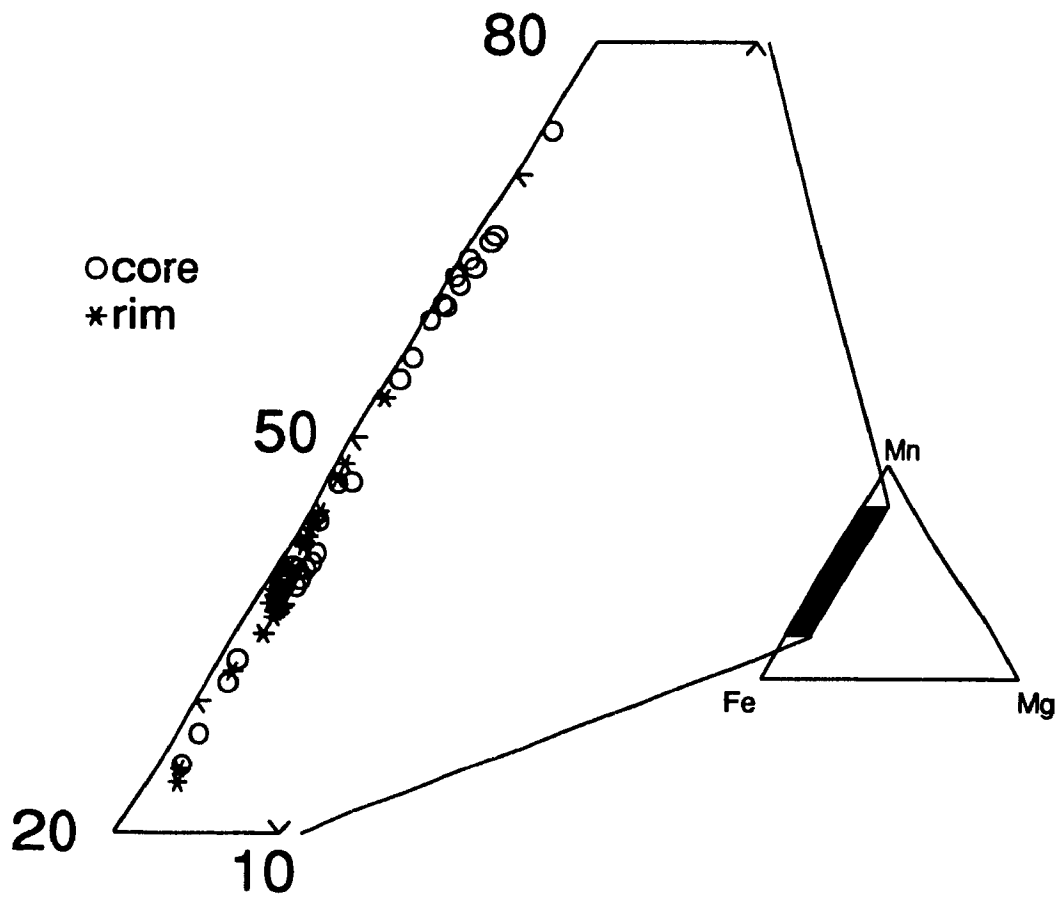
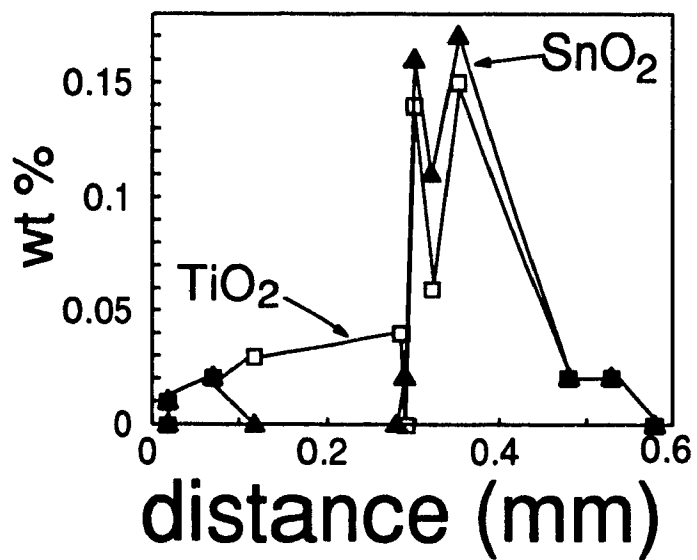
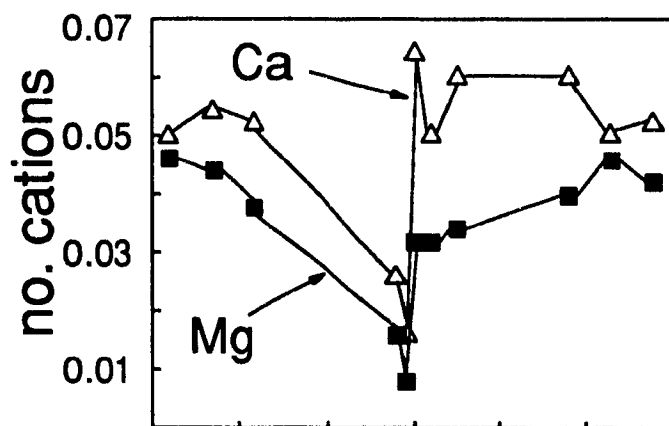
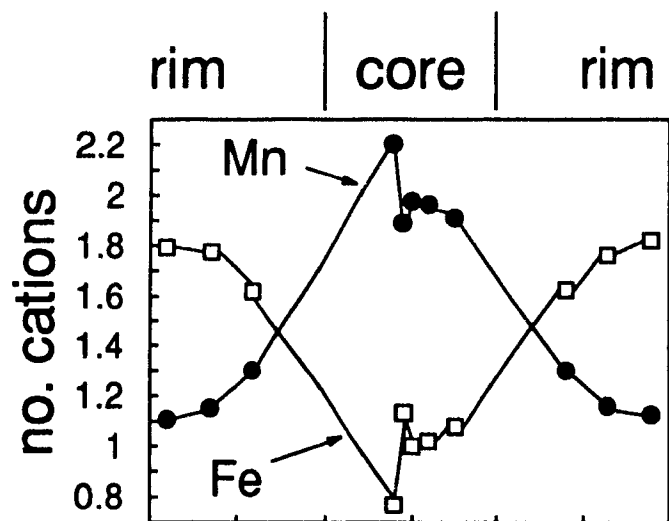


Figure 10. Compositions across a single garnet crystal with a core defined by primary fluid inclusions.



trace amounts; their low concentrations are also indicated by low whole-rock Ti, Nb and Ta abundances (Linnen and Williams-Jones, submitted; Chapter 2). These Ti-Fe-Nb-Ta minerals are common in other Thai pegmatites (Praditwan, 1988) and at Nong Sua occur as inclusions in cassiterite, garnet, tourmaline, muscovite, K-feldspar, plagioclase and quartz. Their occurrence as inclusions is also restricted to aplite-pegmatite hosted phases, i.e., they do not constitute part of the vein mineralogy.

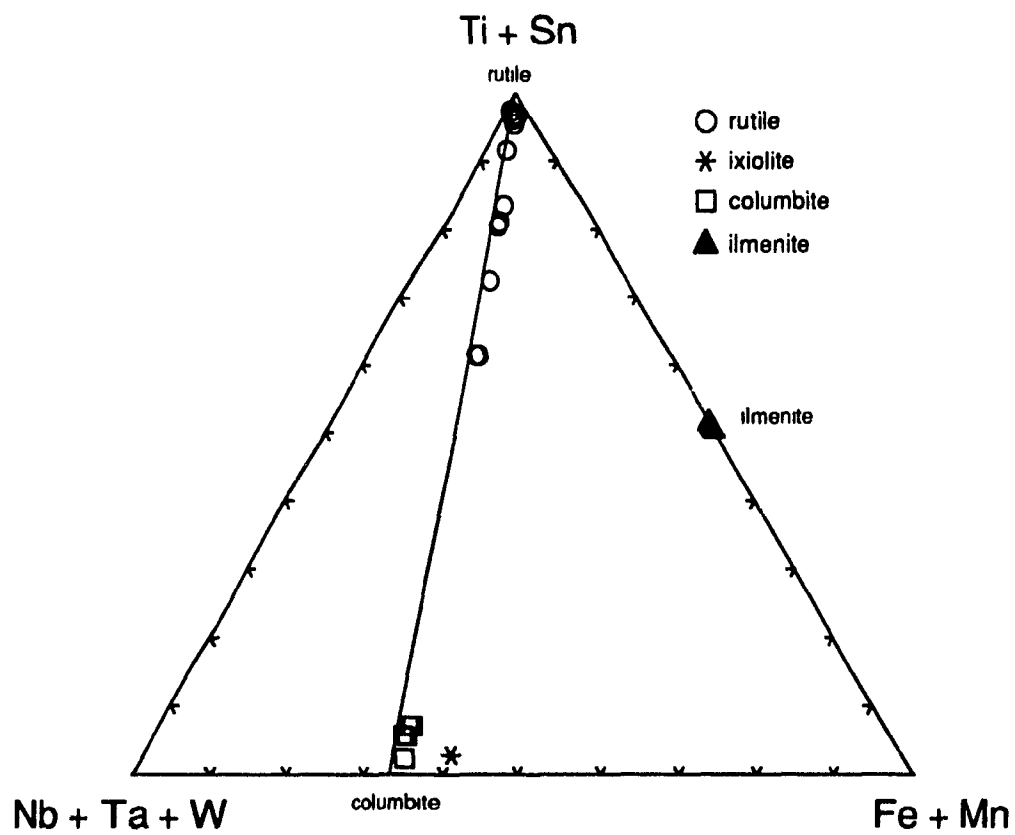
Disseminated and vein cassiterite are red and display only weak colour zoning. Both are chemically homogeneous, consisting of >98 % SnO_2 , with trace amounts of TiO_2 , Ta_2O_5 , Nb_2O_5 and FeO (Table 4). However, as mentioned above, Ti-Fe-Nb-Ta oxide inclusions are common in disseminated cassiterite and are absent in vein-hosted cassiterite, with the exception of very rare rutile inclusions in the case of the latter. The inclusions are generally bleb-like or irregularly shaped, and the percentage of inclusions varies amongst individual cassiterite crystals. This suggests that the inclusions were trapped, rather than originating by exsolution. The trapping of oxide inclusions by cassiterite has two important implications. The first is that, owing to the compositional similarity of the disseminated and vein-hosted cassiterite, both must have formed under similar conditions (T , $f\text{O}_2$, pH , etc.) or re-equilibrated with later fluids. The second implication stems from the fact that the oxide inclusions are common in disseminated cassiterite, but rare elsewhere in the pegmatite. This suggests that cassiterite and rutile-columbite-ixiolite

TABLE 4
REPRESENTATIVE OXIDE COMPOSITIONS

	1	2	3	4	5	6
FeO	9.64	1.67	40.58	16.11	14.24	0.00
MnO	0.08	0.03	6.01	5.90	4.60	0.03
WO ₃	3.62	2.88	0.00	34.33	8.66	0.00
Ta ₂ O ₅	23.11	0.16	0.03	6.18	15.66	0.01
SnO ₂	3.44	1.11	0.46	0.47	0.71	98.23
TiO ₂	45.36	91.67	52.03	1.50	3.50	0.63
Nb ₂ O ₅	14.88	2.04	0.23	36.11	52.36	0.01
CaO	0.02	0.07	0.01	0.02	0.03	0.26
Tot.	100.14	99.88	99.36	100.67	99.75	99.20
Fe	0.141	0.019	0.863	1.161	0.727	0.000
Mn	0.001	0.000	0.129	0.431	0.238	0.001
W	0.016	0.010	0.000	0.767	0.137	0.000
Ta	0.110	0.001	0.000	0.145	0.260	0.000
Sn	0.024	0.006	0.005	0.016	0.017	0.984
Ti	0.596	0.951	0.995	0.097	0.161	0.012
Nb	0.118	0.013	0.003	1.407	1.446	0.000
Ca	0.000	0.001	0.000	0.002	0.002	0.007

1 rutile, 2 rutile, 3 ilmenite, 4 ixiolite, 5 columbite, 6 cassiterite

Figure 11. Compositions of rutile, columbite, ixiolite and ilmenite in terms of cation % Ti, Sn, Ta, Nb, W, Fe, and Mn.



crystallization were linked, i.e. that cassiterite nucleated on pre-existing oxide grains or vice versa. The occurrence of rutile-columbite-ixiolite inclusions in Sn-rich cores of garnet supports this linkage.

The composition of rutile ranges from that of near end-member TiO_2 to having a significant columbite component (Fig. 11). This range of composition is a common feature of pegmatite-hosted rutile (Černý and Ercit, 1985). The substitutions in rutile, in terms of cation proportions, are $\text{Fe} \approx \text{Nb} \approx \text{Ta} \gg \text{Sn} \approx \text{W}$, but the relative importance of different metal substitutions varies with each of the oxides (Table 4). Columbite is strongly enriched in Nb, is Fe-rich, and in W-rich grains $\text{W} \approx \text{Mn}$. The compositional range of ilmenite is more restricted and consists almost entirely of Fe, Ti, and Mn. Ixiolite is rare, occurs as very small grains and is identified from stoichiometry (assuming 8 oxygen) interpreted from electron microprobe analysis. The major cations are Fe, Nb, Mn, Ta and W and the cation totals are ≈ 4.0 .

IMPLICATIONS FOR Sn-W MINERALIZATION

The compositional variation of primary and secondary muscovite from Nong Sua, as discussed above, is similar to that observed elsewhere in peraluminous granite systems. We suggest that the muscovite compositions reflect subsolidus re-equilibration with magmatic and/or external fluids. Temperatures estimated from plagioclase-muscovite geothermometry indicate that the re-equilibration occurred

down to approximately 400°C. Since the timing of the re-equilibration is poorly constrained, it is difficult to link the compositional variation of muscovite with Sn-W mineralization. However, it is probable that primary muscovite continued to interact with subsolidus fluids after cassiterite-wolframite crystallization had ceased, suggesting that disseminated and vein mineralization occurred at temperatures >400°C.

The compositions of tourmaline and garnet appear to reflect the conditions at which they crystallized, i.e., they did not undergo exchange with subsequent fluids. Thus, the compositional overlap of aplite-pegmatite and vein tourmalines suggests that both crystallized at similar conditions. It therefore follows that the veins formed while the pegmatite was at or near solidus conditions. The abrupt termination of veins at aplite-pegmatite contacts (cf. Linnen and Williams-Jones, submitted; Chapter 2) supports this interpretation. Also implicit in this hypothesis is that late-stage pegmatite solidification and vein emplacement occurred over a narrow temperature range, which suggests that decreasing temperature was probably not the dominant control of magmatic or hydrothermal cassiterite-wolframite deposition.

The high concentrations of tin in garnet cores that are defined by fluid inclusions are probably also related to the magmatic-hydrothermal transition. The fact that tin abundances are associated with cores that contain fluid inclusions (other spessartine-rich cores are not Sn-rich) suggests that the tin concentrations are

probably not directly related to manganese contents. We propose that the enrichment of tin in late garnet (fluid inclusions indicating vapour saturation) reflects enrichment of tin at the latest stages of the pegmatite evolution. The restriction of Ti-Nb-Ta oxides to aplite-pegmatite suggests that these elements were not highly mobile at Nong Sua and that rutile, columbite and ixiolite probably crystallized directly from melt. From the common occurrence of these oxides as inclusions in disseminated cassiterite it follows that the disseminated cassiterite crystallized directly from melt. This is consistent with the textural and whole-rock chemical data presented by Linnen and Williams-Jones (submitted; Chapter 2).

It is significant that there is a strong temporal and spatial relationship between cassiterite and tourmaline, and that "mineralized" tourmaline displays the greatest compositional variation. We suggested above that this compositional variation was the result of Ca-Mg metasomatism from external fluids. Diffusion or infiltration of external fluids could potentially affect the solubility of tin in the melt. One of the factors that controls tin solubility in granitic melts is fO_2 (Taylor, 1988). Thus, if more oxidized external fluids interacted with more reduced melt, cassiterite crystallization could have been triggered. Potentially this also explains the association between Ti-Fe-Nb-Ta oxides and disseminated cassiterite. Similarly, fO_2 may have also been important in controlling the hydrothermal mineralization. Unfortunately, the fO_2 cannot be estimated from the minerals present at Nong Sua.

Although an orthomagmatic origin has been proposed for many granite-hosted

deposits, e.g., Eadington (1983), Charoy and Pollard (1989), it is generally difficult to prove this hypothesis since hydrothermal systems also involve widespread alteration (cf. Marignac and Cuney, 1991). By examining a pegmatite system we have shown that cassiterite can crystallize in both magmatic and orthomagmatic systems; a remobilization origin at Nong Sua can be rejected from the lack of pervasive alteration (cf. Linnen and Williams-Jones, submitted; Chapter 2). The magmatic-orthomagmatic model also probably applies to other pegmatite-hosted deposits, and it follows that the orthomagmatic model is a viable explanation of mineralization in granite-hosted deposits.

CONCLUSIONS

The composition of the tourmaline at Nong Sua reflects the intensive and extensive parameters of the melt or fluid from which it crystallized, i.e., it was resistant to exchange with subsequent hydrothermal fluids. The overlap of magmatic and hydrothermal compositions indicates that tourmaline and associated cassiterite crystallized from melt and aqueous fluids at similar physicochemical conditions. Tourmaline also records the involvement of external fluids and Ca-Mg metasomatism.

The composition of muscovite has been modified and indicates water-rock interaction at subsolidus conditions. However, the incipient nature of sericite alteration suggests low water-rock ratios.

Elevated tin abundances in the cores of garnet that are defined by fluid inclusions suggest that the late-stage tin concentration in the melt was high. This supports the hypothesis that some cassiterite crystallized directly from the melt.

The restriction of Ti-Fe-Nb-Ta oxides to aplite-pegmatite suggests that they have a magmatic origin, and their common occurrence in disseminated cassiterite and absence in vein cassiterite support the model that cassiterite crystallized from melt and later precipitated from fluids.

ACKNOWLEDGEMENTS

This research was carried out with the aid of a grant from the International Development Research Centre, Ottawa, Canada. We would like to thank the efforts of S. Bunopas, N. Jungyusuk, S. Khositanont, and P. Putthapiban of the Geological Survey Division, Department of Mineral Resources of Thailand during the course of this project.

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CHAPTER 4

THE EVOLUTION OF PEGMATITE-HOSTED TIN-TUNGSTEN MINERALIZATION AT NONG SUA, THAILAND: EVIDENCE FROM FLUID INCLUSIONS AND STABLE ISOTOPES

by

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ABSTRACT

The Nong Sua aplite-pegmatite complex is host to cassiterite-wolframite mineralization. The orthomagmatic fluid at Nong Sua is preserved as primary fluid inclusions in magmatic garnet. These inclusions have a composition of 3 wt. % NaCl eq. and are linked to tin mineralization by being associated with high tin concentrations in the garnet. The low salinity suggests that, at vapour saturation, tin was partitioned in favour of the melt, which allowed cassiterite to initially crystallized directly from the melt. The orthomagmatic fluid is interpreted to have had a composition of $\delta^{18}\text{O}$ +8.7 to +9.9 per mil and δD of -72 to -78 per mil from $\delta^{18}\text{O}$ analyses of muscovite and quartz and δD of muscovite. The $\delta^{18}\text{O}$ composition of muscovite decreased from 10.1 to 8.0 per mil and δD increased from -106 to -85 per mil from magmatic to hydrothermal stages of pegmatite evolution. These changes are consistent with an influx of metamorphic fluids or evolved meteoric waters. We interpret that the vapour saturation of the melt caused the pressure in the pegmatite to rise. This occurred at approximately 650°C and 3.8 kbar. Fluid overpressure caused the aplite to fracture and veins to form from fluids that migrated into the fracture induced low pressure zones. This event can be modeled by an isothermal decompression to 2.7 kbar, followed by cooling. The orthomagmatic fluids mixed with external high salinity and carbonic fluids; cassiterite deposition is interpreted to have been controlled by increasing $f\text{O}_2$,

whereas wolframite resulted from the mixing of W-rich with Fe-Mn-rich fluids. In both cases cooling may also have contributed to mineralization.

INTRODUCTION

In spite of the fact that aplites and pegmatites have long been recognized to be economically important hosts of cassiterite mineralization (e.g., Derry, 1931), tin-bearing pegmatites have received little scientific scrutiny. Most of the research has concentrated on the internal structure and mineralogy of the pegmatites (e.g. Gouanvic and Gagny, 1987; Foord and Cook, 1989; Kuster, 1990; Robles *et al.*, 1991). Very few fluid inclusion studies have been carried out (Little, 1960; Haapala and Kinnunen, 1982; Mangas and Arribas, 1987; Segalstad and Eggleston, 1990), and we are not aware of any stable isotope studies of tin-bearing pegmatites. The origin of the cassiterite in these deposits has been proposed to be magmatic (e.g., Gouanvic and Gagny, 1983), hydrothermal (e.g., Gocht and Pluhar, 1982; Ollila, 1985) or both (e.g., Little, 1960) and remains poorly understood. Abundant pegmatite-hosted tin deposits that occur in the western Thailand granite belt are well suited for establishing the genesis of this style of mineralization. One of these deposits, at Nong Sua, was therefore selected for an investigation of the magmatic and hydrothermal controls of tin-tungsten mineralization.

The Nong Sua complex, located approximately 180 km south of Bangkok, hosts cassiterite disseminated in pegmatite and to a lesser extent in aplite; vein-hosted wolframite and cassiterite; and cassiterite intergrown with muscovite at the contacts between pegmatite and metasedimentary rocks (contact mineralization).

The complex comprises laterally continuous horizons of garnet aplite, tourmaline aplite, banded pegmatite, and tourmaline pegmatite consisting of quartz, K-feldspar, and plagioclase $\pm$ muscovite $\pm$ tourmaline $\pm$ garnet $\pm$ apatite (Fig. 1). Blocks of psammitic metasediment are contained within the complex and at the contacts with aplite-pegmatite they are altered to tourmaline with minor quartz and muscovite. Textural relationships suggest that the pegmatite crystallized largely at vapour-undersaturated conditions and that K-feldspar is late. Whole-rock compositions are interpreted to reflect an evolution from peraluminous (Na-rich) to more alkalic (K-rich) compositions and a build-up of Sn in the residual melt. This eventually led to cassiterite saturation, at or prior to vapour saturation of the melt (Linnen and Williams-Jones, submitted; Chapter 2). Some tourmaline, particularly that intergrown with disseminated cassiterite, show Ca-Mg enrichment which is interpreted to reflect an influx of external fluids (Linnen and Williams-Jones; Chapter 3). Garnet textures and compositions are consistent with a magmatic origin, and elevated Sn contents in garnet cores defined by primary fluid inclusions suggest high tin abundances in the melt at, or prior, to vapour saturation (op. cit.). The veins are contained within aplite units and terminate abruptly at aplite-pegmatite contacts. This relationship and the compositional similarity of magmatic and hydrothermal tourmaline (op. cit.) suggest that the veins formed from aqueous fluids while the pegmatite was at solidus conditions. The contact mineralization is also interpreted to have precipitated from hydrothermal fluids. Muscovite was more

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susceptible to re-equilibration and the compositions of magmatic and hydrothermal muscovite indicate the involvement of fluids at subsolidus temperatures (op. cit.).

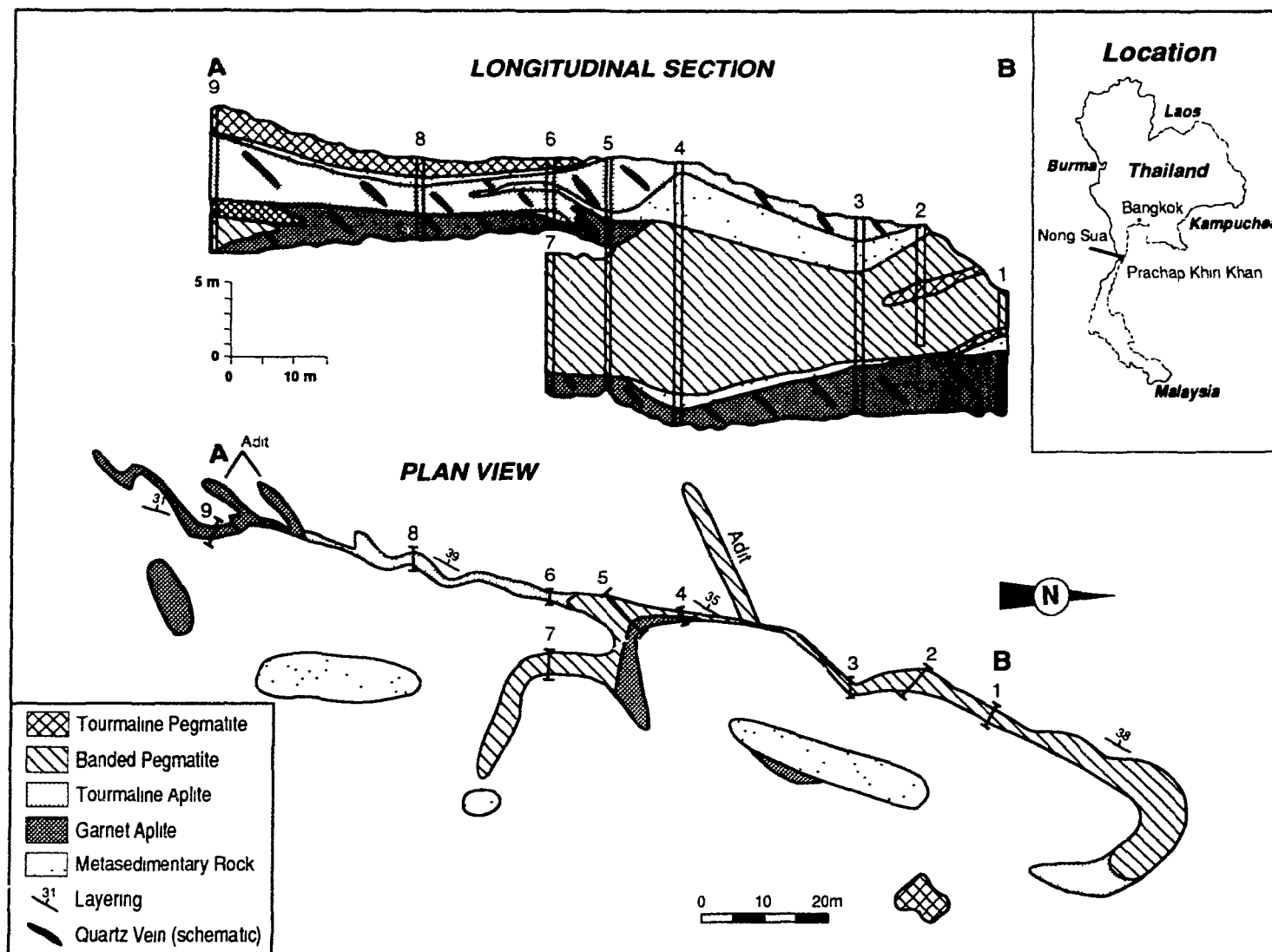
In this study the origin, composition, temperature and pressure of hydrothermal fluids at Nong Sua are interpreted from stable isotope and fluid inclusions analyses. Fluid inclusions in garnet and cassiterite are demonstrably primary and pseudosecondary, respectively, and it is possible to relate these inclusions to the pegmatite evolution and mineralization. The geological and mineralogical context from which fluid inclusions and stable isotopic data can be interpreted is provided in our previous work, summarized above. These data are integrated in a model that appears to explain satisfactorily the fluid evolution at Nong Sua and the controls of Sn-W deposition.

FLUID INCLUSIONS

PETROGRAPHY

Fluid inclusions are observed in cassiterite, quartz, tourmaline, and garnet. Type 1 inclusions are two-phase L-V aqueous inclusions. This type may also contain 1 or 2 birefringent solids. The number and size of the solids are variable, indicating that the solids are probably trapped. Within individual clusters of type 1 fluid inclusions, solid-bearing and solid-free inclusions have similar ice melting and homogenization temperatures and none of the solids homogenized on heating,

Figure 1. A plan view and longitudinal section showing the geology of the Nong Sua aplite-pegmatite, from Linnen and Williams-Jones (submitted; Chapter 2).



further indicating that they are not daughter minerals. Therefore the solid-bearing inclusions are included with L-V inclusions in the type 1 data set. Type 2 inclusions are three-phase L-L-V aqueous-carbonic; type 3 inclusions are carbonic-rich with one, two, or three phases V, V-L, and V-L-L; type 4 inclusions are aqueous halite-bearing L-V-H inclusions, this type may also contain additional birefringent solid(s); and type 5 inclusions are aqueous, solid-rich inclusions. Very rarely up to 8 solids, none of which is halite, are observed in aqueous inclusions hosted by tourmaline. The vapour + liquid in these inclusions occupies <50 % of the volume (in contrast to solid-bearing type 1 inclusions) and potentially these once were melt inclusions.

Many type 1 inclusions contain a carbonic vapour, which is made evident by the formation of clathrate and carbonic ice. These inclusions are distinguished from type 2 inclusions by the absence of visible carbonic liquid. Type 3 inclusions are vapour-rich and form a continuum with type 2 inclusions; if the volume percent of the vapour is near 50 %, inclusions which homogenize to vapour are classified as V-L-L (type 3) and those which homogenize to liquid are classified as L-L-V (type 2).

Type 1 inclusions are the most abundant variety observed in cassiterite, tourmaline and garnet. In garnet they commonly delineate the cores of crystals (Fig. 2a,b) and are therefore interpreted to be primary in origin. Some type 1 fluid inclusions in cassiterite occur along planes which terminate at cassiterite

overgrowths (Fig. 2c) and thus are demonstrably pseudosecondary. The origin of other inclusions in cassiterite is not clear; however, they are commonly isolated or in clusters. These inclusions display microthermometric behaviour similar to that of the primary inclusions in garnet and therefore are probably primary. The origin of type 1 inclusions in tourmaline is equivocal, but many inclusions are aligned parallel to the c-axis and may be primary. Type 1 inclusions in quartz are generally secondary. Type 2 inclusions are most abundant in quartz, where they are secondary, and are less common in cassiterite and tourmaline, where their origin is unknown. Type 3 inclusions are observed in all minerals; in quartz and garnet they are secondary, whereas in cassiterite and tourmaline they are isolated or occur as clusters. Some type 3 inclusions in quartz are present with type 1 inclusions in secondary planes suggesting possible fluid immiscibility. Type 4 inclusions are observed only in secondary planes in quartz. Fluid inclusions are typically 5 to 15 μm in diameter, although rarely inclusions in tourmaline range up to 50 μm across. The primary fluid inclusions and some of the secondary fluid inclusions have rounded or elliptical shapes, whereas pseudosecondary and other secondary fluid inclusions are more irregularly shaped (cf. Fig. 2).

ANALYTICAL METHODS

Fluid inclusion compositions were determined by a combination of microthermometry, laser Raman microscope spectroscopy, and scanning electron

Figure 2. Photomicrographs of fluid inclusions from the Nong Sua deposit.

- a) Primary liquid-vapour (L-V) and solid inclusions in garnet. The fluid inclusions are bounded by crystallographic faces in the core of the garnet. Field of view 0.9 mm across.
- b) Enlargement of Figure 2a, primary L-V and solid inclusions bounded by the crystallographic faces in garnet. Field of view 0.36 mm across.
- c) Pseudosecondary liquid-vapour (L-V) inclusions in cassiterite. The pseudosecondary plane terminates at the dark overgrowth. Field of view 0.53 mm across.

a



b



C



microscopy with an energy dispersive system. The microthermometric analyses were determined on a Fluid Inc. stage calibrated using synthetic fluid inclusion standards. Measurements of phase changes at or below 31°C are accurate to within $\pm 0.2^\circ\text{C}$ and high temperature measurements to within $\pm 2.0^\circ\text{C}$. Raman spectroscopic data were collected on a Ramanor U1000 spectrometer and Nacet microscope (x80 objective) fitted with a Spectra Physics Ar laser (514.5 nm line) calibrated at the Δcm^{-1} 918.55 peak using an Ne lamp. The laser was operated at 200-300 mW, slit widths were 300 μm and scans consisted of 5-10 second counting times using wave number increments of 0.5 cm^{-1} over the appropriate ranges for CO_2 , CH_4 , and N_2 . Peak areas were calculated from 4 to 8 coadded spectra using standard numerical integration methods and CH_4/CO_2 and N_2/CO_2 ratios were calculated using the peak area equations and quantification factors of Wopenka and Pasteris (1987). The quantification factors are machine specific; however, it has been shown that these factors are valid for the McGill spectrometer (cf. Salvi and Williams-Jones, in press). Fluid inclusion decrepitates were analyzed on a Cambridge scanning electron microscope equipped with a Tracor-Northern energy dispersive analyzer after being heated to 400°C on the fluid inclusion stage.

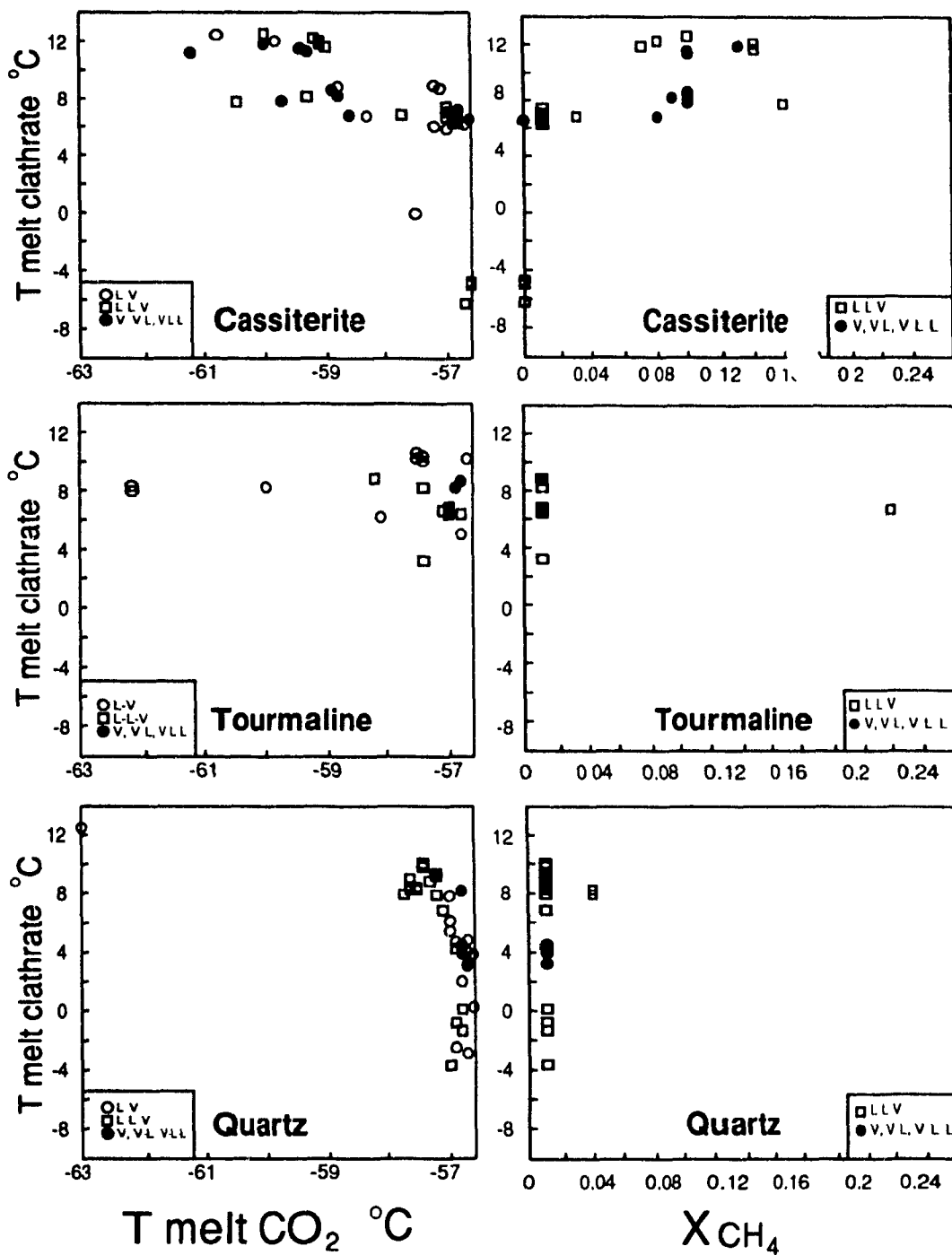
Carbonic Phase Compositions

Carbonic ice melting temperatures were measured for inclusion types 1, 2, and 3 in cassiterite, tourmaline and quartz. There is no correlation between this

melting temperature and inclusion type (Fig. 3) and, within individual clusters, there is no significant difference in the carbonic ice melting temperatures of different inclusion types. In addition, there are no differences in the carbonic ice melting temperatures of vein and disseminated cassiterite. Carbonic ice melting temperatures in cassiterite vary from -56.6° to -60.8°C (Fig. 3). The data for carbonic ice melting temperatures in tourmaline are much fewer. These temperatures range from -56.7° to -62.2°C and, as in the case of cassiterite, the range is independent of the inclusion type. The carbonic ice melting temperatures in quartz are restricted to greater than -58.0°C , with the exception of one determination of -63.0°C . Clathrate decomposition temperatures in inclusions hosted by all minerals range from -10.3° to $+15.2^{\circ}\text{C}$ (Fig. 3) and, as in the case for the carbonic ice melting temperatures, are independent of inclusion type (Fig. 3).

The carbonic ice melting temperature is dependent on the composition of the carbonic phases (CO_2 , CH_4 , N_2 etc.), whereas the clathrate melting temperature is dependent on both the carbonic composition and the salinity of the aqueous phase (cf. Bozzo et al. 1973). If homogenization temperatures of the carbonic phases are greater than clathrate decomposition temperatures, a negative linear correlation between carbonic ice and clathrate melting temperatures indicates constant salinity and variable carbonic composition, whereas if the carbonic ice melting temperature is constant but the clathrate melting temperature varies, salinity is variable and the carbonic composition is constant. It is apparent from Figure 3 that the carbonic

Figure 3. Clathrate melting temperatures versus CO₂ melting temperatures and X_{CH₄} for type 1, 2, and 3 inclusions in cassiterite, tourmaline, and quartz. The methane mole fraction ($X_{CH_4} = CH_4/CH_4+CO_2$) was calculated after the method of Burruss (1981).



compositions of quartz-hosted inclusions are close to that of pure CO_2 but the salinity is variable. By contrast, inclusions hosted by cassiterite and tourmaline vary in both salinity and carbonic compositions, although for inclusions in cassiterite that have clathrate decomposition temperatures $>4^\circ\text{C}$ there is a fairly good negative correlation between carbonic ice and clathrate melting temperatures, indicating that most of the compositional variation is in the carbonic phase.

The mole fraction of methane ($\text{CH}_4/(\text{CO}_2+\text{CH}_4)$) was calculated from carbonic ice melting and homogenization temperatures using the method of Burruss (1981). The carbonic phases in type 2 inclusions in cassiterite and tourmaline generally homogenize to vapour, between $+13.0^\circ$ and $+29.5^\circ\text{C}$, and in quartz typically to liquid between $+25.0^\circ$ and $+30.0^\circ\text{C}$. The homogenization of the carbonic phases in type 3 inclusions is, with a few exceptions, to liquid, regardless of the host mineral. In cassiterite, this temperature ranges from $+9.6^\circ$ to $+29.2^\circ\text{C}$, in tourmaline there were only two determinations, at $+29.5^\circ$ and $+29.7^\circ\text{C}$, and in quartz the range is from $+26.8^\circ$ to $+29.4^\circ\text{C}$. The X_{CH_4} of the carbonic phase, calculated from microthermometry ranges from 0.0 to approximately 0.2 (Fig. 3). Inclusions in cassiterite contain significant CH_4 , whereas those in quartz are CO_2 -rich. Some inclusions in tourmaline may also contain CH_4 , although the data are limited. If the carbonic phase is composed solely of CO_2 and CH_4 , the clathrate melting temperatures and X_{CH_4} should correlate positively at constant salinity. The scatter in the cassiterite data suggests that either salinity is variable or that

additional volatiles are present, e.g., N_2 . It is also important to note that type 2 and 3 inclusions (H_2O - and CO_2 -rich, respectively) both exhibit a similar range of X_{CH_4} .

The presence of CO_2 , CH_4 , and N_2 were confirmed by Raman spectroscopic analysis of type 2 and type 3 fluid inclusions. Representative spectra for CO_2 , CH_4 , and N_2 are shown in Figure 4. Nine inclusions in cassiterite and two inclusions in tourmaline were analyzed. In four of these inclusions only CO_2 was detected, in five inclusions $CO_2 + CH_4$ are present, in one inclusion $CO_2 + CH_4 + N_2$ are present, and in one inclusion $CO_2 + N_2$ are present. The X_{CH_4} ranges up to 0.26 and X_{N_2} (relative to CO_2) ranges up to 0.08. A comparison of the X_{CH_4} calculated from microthermometry and that from Raman spectroscopy is presented in Table 1. Only three inclusions were analyzed using both microthermometry and Raman spectroscopy. For one inclusion the two estimates of X_{CH_4} are close, in the second inclusion the Raman estimate is significantly lower but a possible (non-definitive) N_2 peak may explain this difference. The third inclusion does not contain CH_4 but N_2 is present, hence the X_{CH_4} calculated from microthermometry is an X_{CH_4} equivalent.

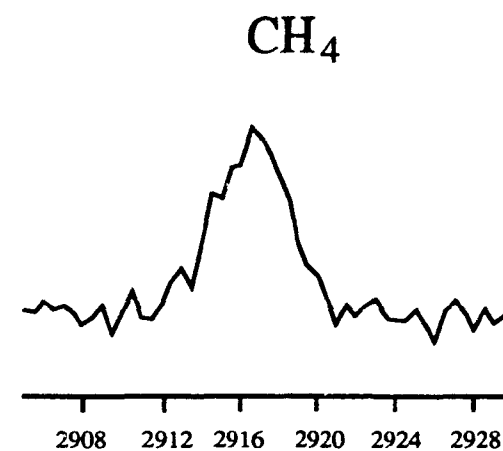
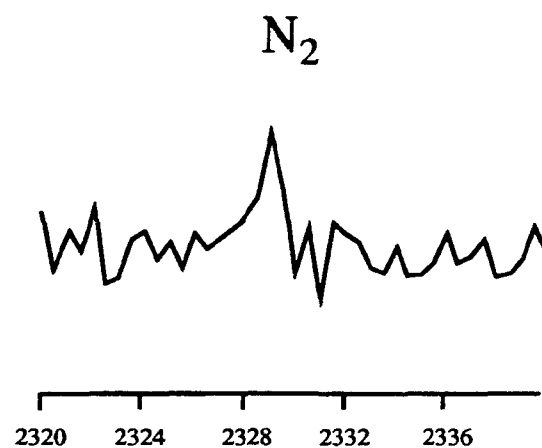
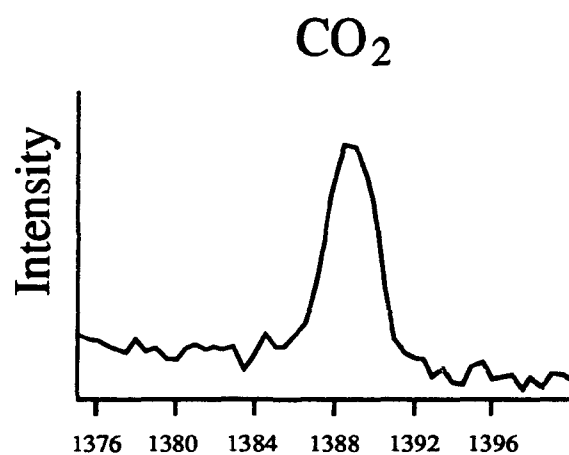
The X_{CO_2} ($CO_{2eq.}/CO_{2eq.}+H_2O$) of type 2 and 3 inclusions was estimated using the method of Bodnar (1983). The cross-sectional areas of the fluid inclusions and the vapour bubbles were accurately measured with the aid of a drawing tube. The depth of fluid inclusions calculated by this method was generally less than the diameters of the vapour bubbles. Elliptical outlines (as

TABLE 1
COMPARISON OF MICROTHERMOMETRIC AND RAMAN SPECTROSCOPIC ESTIMATES
OF CARBONIC PHASE COMPOSITIONS IN TYPE 2 AND 3 INCLUSIONS

Mineral	Type	Microthermometry			Raman Spectroscopy				
		Tm CO ₂	Th CO ₂	X _{CH₄} ¹	Area CO ₂	Area CH ₄	Area N ₂	X _{CH₄} ²	X _{N₂} ²
cass	V	-59.4	15.7(L)	0.10	194.9	81.6	-	0.09	
cass	V	-	-	-	108.8	32.2	?	0.06	
cass	L-L-V	-59.0	18.0(V)	0.14	40.6	2.0	?	0.01	
cass	V	-60.3	-	-	102.6	57.6	?	0.11	
cass	V-L-L	-	-	-	110.3	174.4	?	0.26	
cass	V	-	-	-	107.8	57.6	1.8	0.11	0.02
tour	L-V	-62.2	-	-	45.1	14.9	-	0.07	
tour	L-L-V	-61.8	26.9(L)	0.24	121.6	-	7.4		0.08

Tm CO₂ carbonic melting temperature, Th CO₂ carbonic homogenization temperature (L) to liquid and (V) to vapour, X_{CH₄}¹ mole fraction of CH₄ relative to CO₂ calculated from microthermometric data, Area CO₂, CH₄, and N₂ Raman peaks, ? non-definitive peak, X_{CH₄}² and X_{N₂}² mole fractions of CH₄ and N₂, relative to CO₂ calculated from Raman spectroscopic data.

Figure 4. Typical Raman spectra for a) CO₂, b) CH₄, and c) N₂ in type 2 and 3 inclusions.



Δ Wavenumbers

opposed to circular) of the carbonic phase within aqueous-carbonic inclusions and elliptical outlines of carbonic vapour within carbonic liquid also indicate that the inclusions are relatively flat. The X_{CO_2} can therefore be estimated using the area % of the aqueous and carbonic phases.

The density of the H_2O portion of an inclusion was assumed to be 1.0 g/cm^3 (although saline aqueous fluids have higher densities, this has only a minor affect on the calculated X_{CO_2}) and the carbonic fluid density was calculated using the data of Arai (1971), for the appropriate X_{CH_4} . It was possible to calculate the X_{CO_2} for approximately 20 type 2 and type 3 inclusions. Type 2 inclusions that homogenize to liquid have X_{CO_2} values that range from .03 to .20, whereas for those that homogenize critically, X_{CO_2} ranges from .02 to .08. Type 3 inclusions homogenize to vapour and have X_{CO_2} values that range from .12 to .30. Many type 3 inclusions contain a single visible phase (vapour-only), however, clathrates were observed in these inclusions, indicating the presence of H_2O . The X_{CO_2} of these inclusions is almost certainly $>.30$, but cannot be accurately estimated.

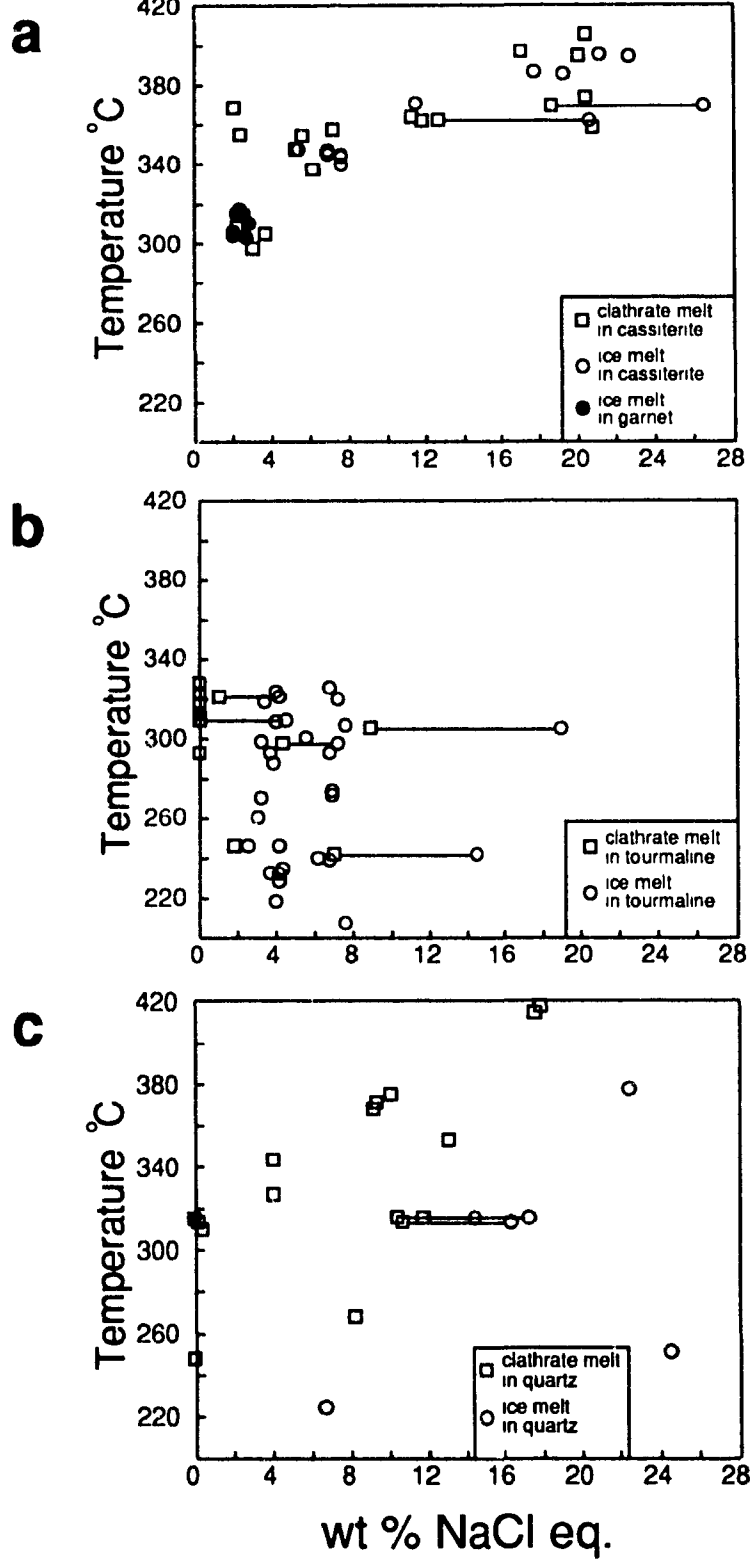
Aqueous Phase Compositions

Salinities of type 1 inclusions were calculated from clathrate decomposition temperatures using the equations of Bozzo et al. (1973) (all clathrate melting temperatures $\geq 10^\circ\text{C}$ were given a salinity of zero), and from ice melting temperatures using the equations of Oakes et al. (1990), assuming that the salinity

can be modelled in the system $\text{H}_2\text{O}-\text{NaCl}$. The ice in some inclusions melts at temperatures lower than -21°C , the eutectic in the $\text{H}_2\text{O}-\text{NaCl}$ system, and the temperature of initial melting in these inclusions is as low as -60°C , indicating the presence of salts other than NaCl . If these inclusions are modelled in the system $\text{H}_2\text{O}-\text{CaCl}_2$ (the justification for this is discussed below) the maximum error in the calculated salinity is approximately 1 wt % (Oakes *et al.*, 1990). Type 1 inclusions in cassiterite, tourmaline, and quartz all display a wide range of salinity, from 0 to 26.5 wt % NaCl eq. (Fig. 5) and the range of salinities is independent of the presence or absence of CO_2 . A similar range of salinity is also found in inclusions from both vein and disseminated cassiterite. Although salinities calculated from clathrate dissolution temperatures are a minimum if CH_4 is present and salinities calculated from ice melting temperatures are a maximum if clathrate is present, there is a good correlation between salinities calculated by both methods, particularly for cassiterite-hosted inclusions (Fig. 5). The salinities of nine secondary halite-bearing (type 4) inclusions in quartz range from 33 to 36 wt. % NaCl eq., calculated from halite dissolution temperatures using the equations of Bischoff (1991). Coincident with changes in salinity are changes in the freezing temperature (T_n) and temperatures at which ice is first observed to melt (T_e). The true eutectic temperature must lie between these two values. For type 1 inclusions the eutectic temperature ranges from approximately -20° to -60°C and for type 4 inclusions is at approximately -60°C (Fig. 6). The lowering of the eutectic

Figure 5. Plots of salinity versus homogenization temperature of type 1 (L-V±S) aqueous inclusions from a) cassiterite and garnet; b) tourmaline; c) quartz. Salinities were calculated from clathrate melting temperatures (Bozzo et al. 1973), and from ice melting temperatures (Oakes et al. 1990).

Type 1 (L-V) inclusions



temperature with increasing salinity is probably caused by an increase in the abundance of a salts such as CaCl_2 or FeCl_2 .

In order to determine the cation compositions, type 1 fluid inclusions were decrepitated and the evaporated residues on the surface of doubly polished wafers of cassiterite were analyzed by an energy dispersive system on a scanning electron microscope (cf. Haynes et al., 1988). Sodium and chlorine were the elements most commonly present as residues; other element were not typically detected. A representative spectrum of NaCl, is shown in Figure 7. The common occurrence of NaCl precipitates without other salts is consistent with the high proportion of inclusions that contain <10 wt. % NaCl eq. and the correspondingly high eutectic temperatures of these inclusions (Fig. 6). Calcium, Fe, Mn, and K peaks are observed in addition to Na and Cl in a subordinate number of inclusions (Fig. 7). The Ca peak overlaps with one of the Sn peaks (from the host cassiterite), and therefore its presence is indicated by the relative heights of the two main Sn peaks (compare the first two spectra on Fig. 7). Many of the decrepitates could contain Ca that was not identified owing to the peak interference. Similarly, because of peak interference, K was identified by an asymmetric broadening of one of the Sn peaks (Fig. 7) thus minor K may also be present in decrepitates where only Na and Cl were identified.

If the high salinity L-V (type 1) inclusions are modeled in the CaCl_2 -NaCl- H_2O system, the minimum $\text{CaCl}_2/\text{NaCl}$ wt. ratio is determined by the intersection

Figure 6. Ice melting temperatures ($T_{m_{ice}}$) versus freezing temperatures (T_n) and the temperature of the initial ice melt (T_e) for type 1 and 4 inclusions. Tie lines join data from individual inclusions.

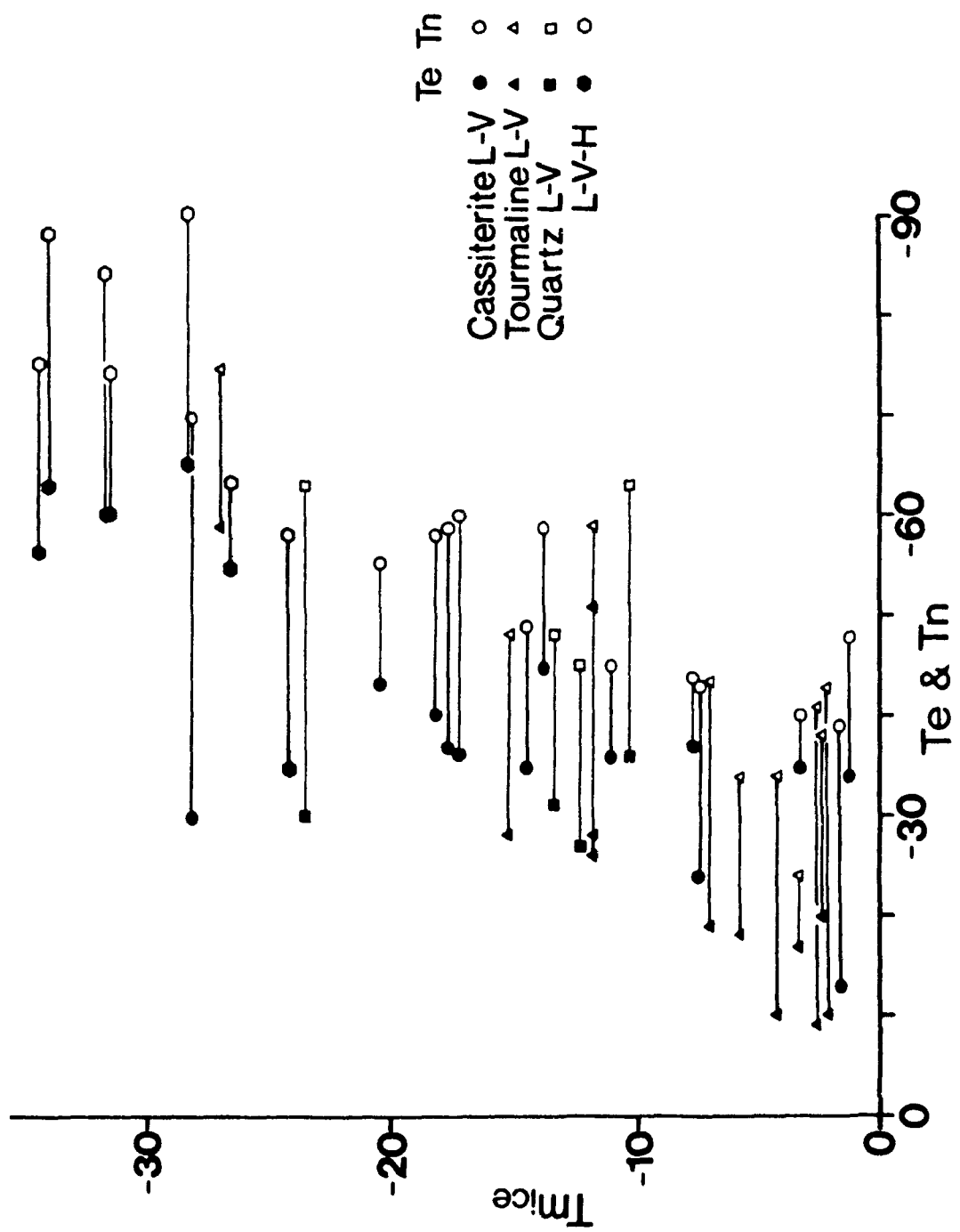
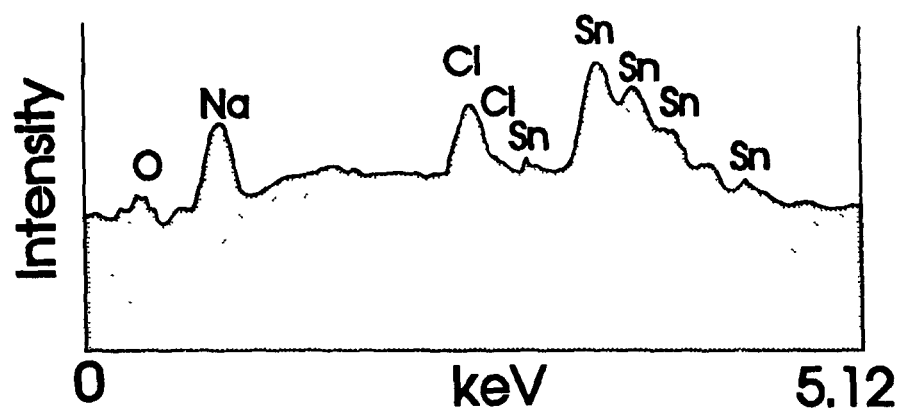
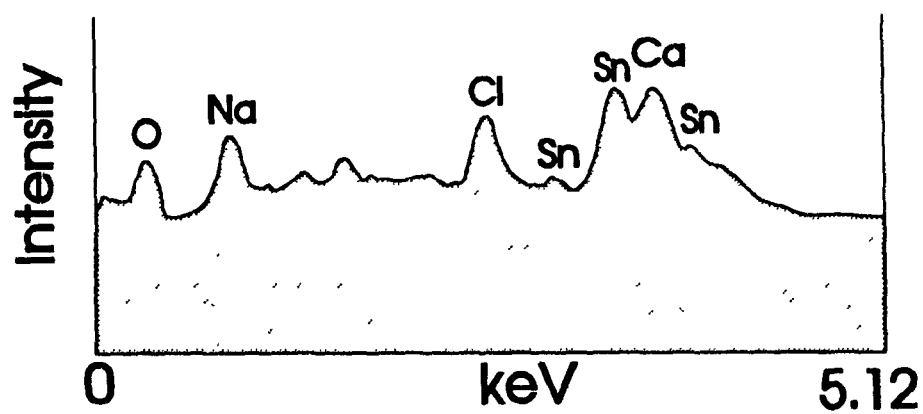


Figure 7. Typical energy dispersive spectra of fluid inclusion decrepitates showing:
a) NaCl, b) NaCl + CaCl₂, and c) NaCl + KCl + Fe & Mn chlorides. The
Sn and O peaks in each case are from the host mineral, cassiterite.

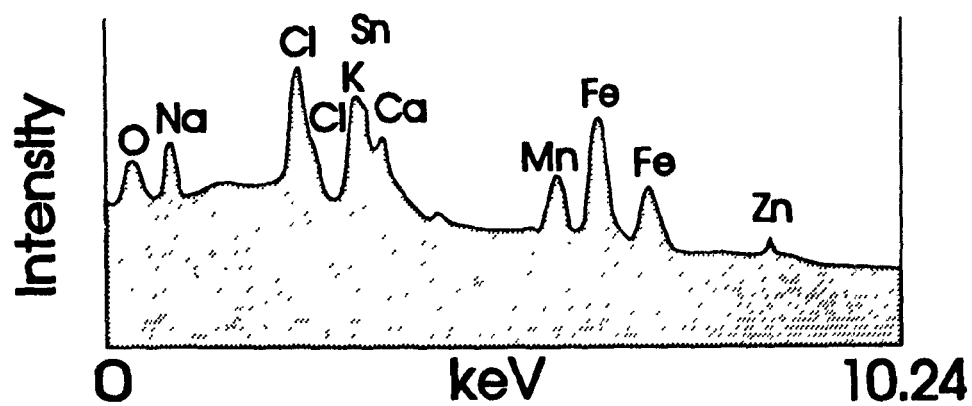
a



b



c

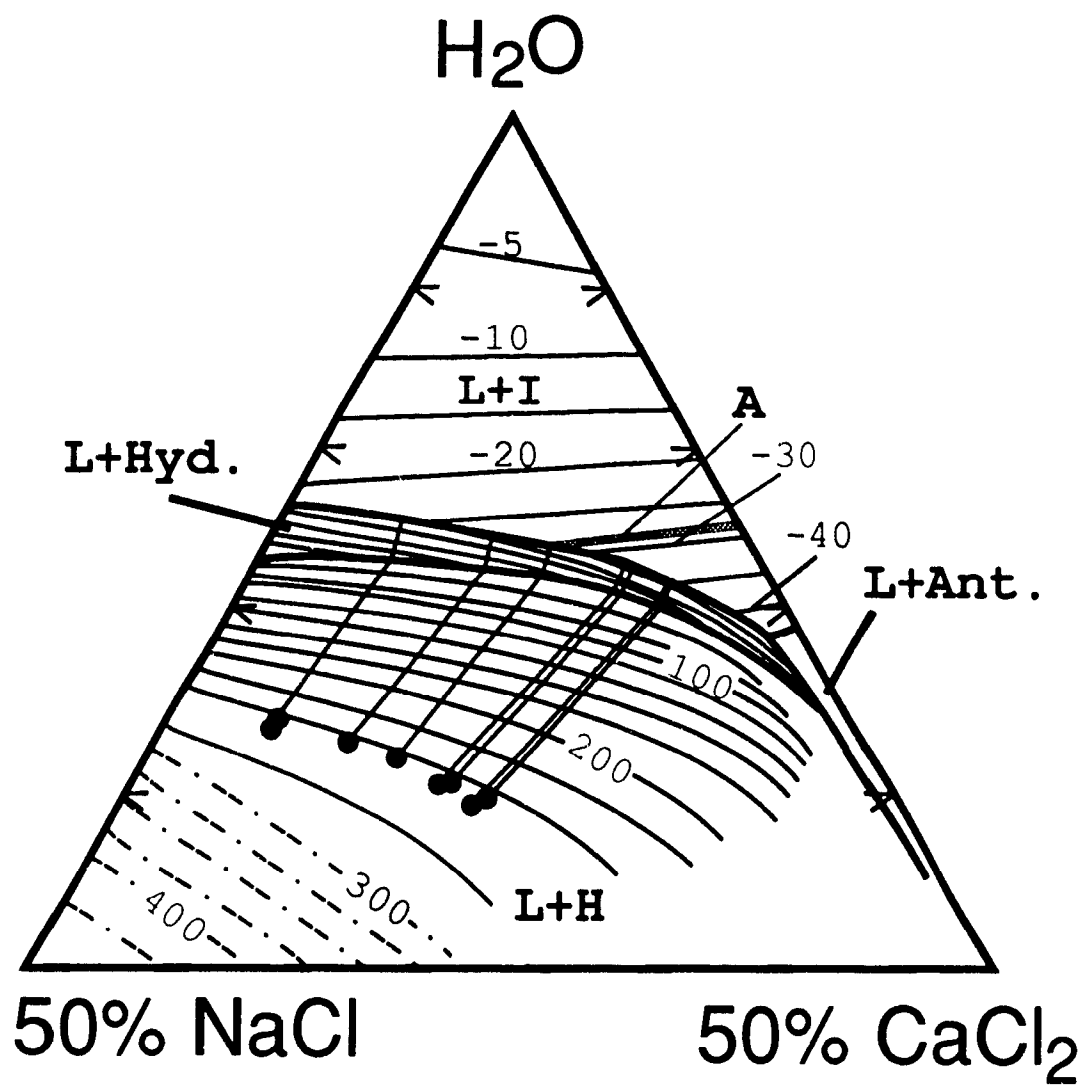


of the ice melting isotherm with the liquid-liquid+hydrate phase boundary (higher $\text{CaCl}_2/\text{NaCl}$ ratios are implied if the ice melts in the presence of liquid-only). The most saline type 1 inclusion (hosted by cassiterite) has a final ice melting temperature of -28.4°C . This corresponds to a minimum $\text{CaCl}_2/\text{NaCl}$ wt. ratio of approximately 1.2 ('A' on Fig. 8) using the isotherms of Oakes *et al.* (1990). The $\text{CaCl}_2/\text{NaCl}$ wt. ratio in secondary type 4 inclusions in quartz ranges from 0.18 to 0.89 (Fig. 8) determined from ice melting and halite dissolution temperatures and modeled in the system $\text{CaCl}_2\text{-NaCl-H}_2\text{O}$ (Williams-Jones and Samson, 1990). The enrichment of CaCl_2 in high salinity type 1 and type 4 inclusions suggests that these two types may be related. This potential relationship is supported by the similarity of eutectic and ice melting temperatures of the two fluid inclusion types (Fig. 6).

AQUEOUS AND CARBONIC FLUID EVOLUTION

Type 1 inclusions in cassiterite, tourmaline, quartz, and garnet homogenize at temperatures between 220° and 420°C (Fig. 5). Primary type 1 inclusions in garnet cores homogenize over a narrow temperature range, from 295° to 310°C , and are of low salinity, approximately 3 wt % NaCl eq. The salinities and homogenization temperatures of some type 1 inclusions in cassiterite are similar, suggesting that these are also primary. The pseudosecondary type 1 inclusions in cassiterite have high salinities and homogenization temperatures, up to

Figure 8. Compositions of Type 1 and Type 4 fluid inclusions modelled in the system $\text{H}_2\text{O}-\text{NaCl}-\text{CaCl}_2$. L+I represents liquid + ice, L+Hyd. liquid + hydrohalite, L+Ant. liquid + antarcticite and L+H liquid + halite. Area 'A' represents the range of compositions possible for a fluid inclusion with a final ice melting temperature of -28°C . The isotherms $<0^\circ\text{C}$ are from Oakes et al. (1990) and $>0^\circ\text{C}$ from Williams-Jones and Samson (1990).



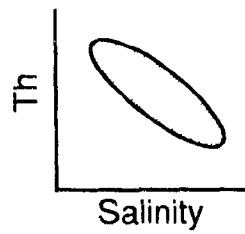
approximately 23 wt. % NaCl eq. and 400°C, respectively. There is a complete range of salinities and homogenization temperatures between the primary and pseudosecondary inclusions in cassiterite, which indicates mixing between two fluids. Type 1 inclusions in tourmaline have similar salinities to those in garnet but tend to homogenize at a lower temperature. However, the homogenization temperatures of type 1 inclusions in tourmaline that are oriented parallel to the c-axis are restricted to the higher end of the range, typically between 270° and 310°C, suggesting that they are probably primary. The data for type 1 inclusions from quartz are not consistent and cannot be meaningfully interpreted.

The low salinity inclusions hosted by garnet are interpreted to represent the orthomagmatic fluid, because of the magmatic and primary origin of the garnet and fluid inclusions, respectively. The high salinity type 1 and late secondary type 4 inclusions are interpreted to have high Ca/Na ratios. This contrasts with the low Ca/Na ratios of the host rock (Linnen and Williams-Jones, submitted; Chapter 2), and suggests that these fluids had an external source. Late Ca-enrichment in tourmaline and late scheelite replacement of wolframite (Linnen and Williams-Jones, submitted; Chapter 2) support this hypothesis. The salinity-homogenization temperature trend displayed by type 1 inclusions in cassiterite (Fig. 5) cannot be explained by the trapping of fluid mixtures at constant temperature and pressure (Fig. 9). A positive correlation between salinity and homogenization temperature can be explained by isothermal trapping of fluids with decreasing pressure and

Figure 9. Salinity versus homogenization temperature relationships for various hypothetical conditions of mixing between halite-undersaturated aqueous fluids.

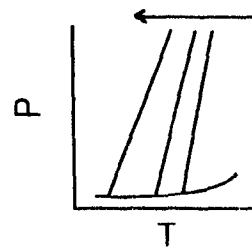
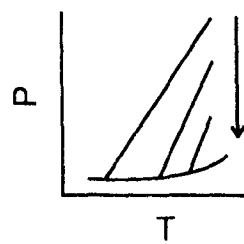
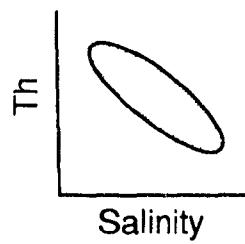
MIXING PATTERNS FOR AQUEOUS L-V INCLUSIONS

constant P & T
variable salinity



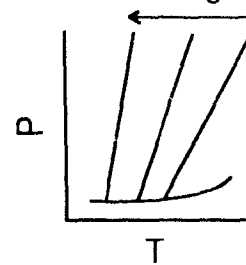
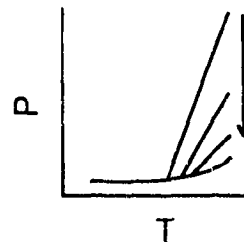
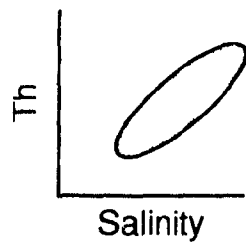
decreasing P
decreasing salinity

decreasing T
increasing salinity



decreasing P
increasing salinity

decreasing T
decreasing salinity

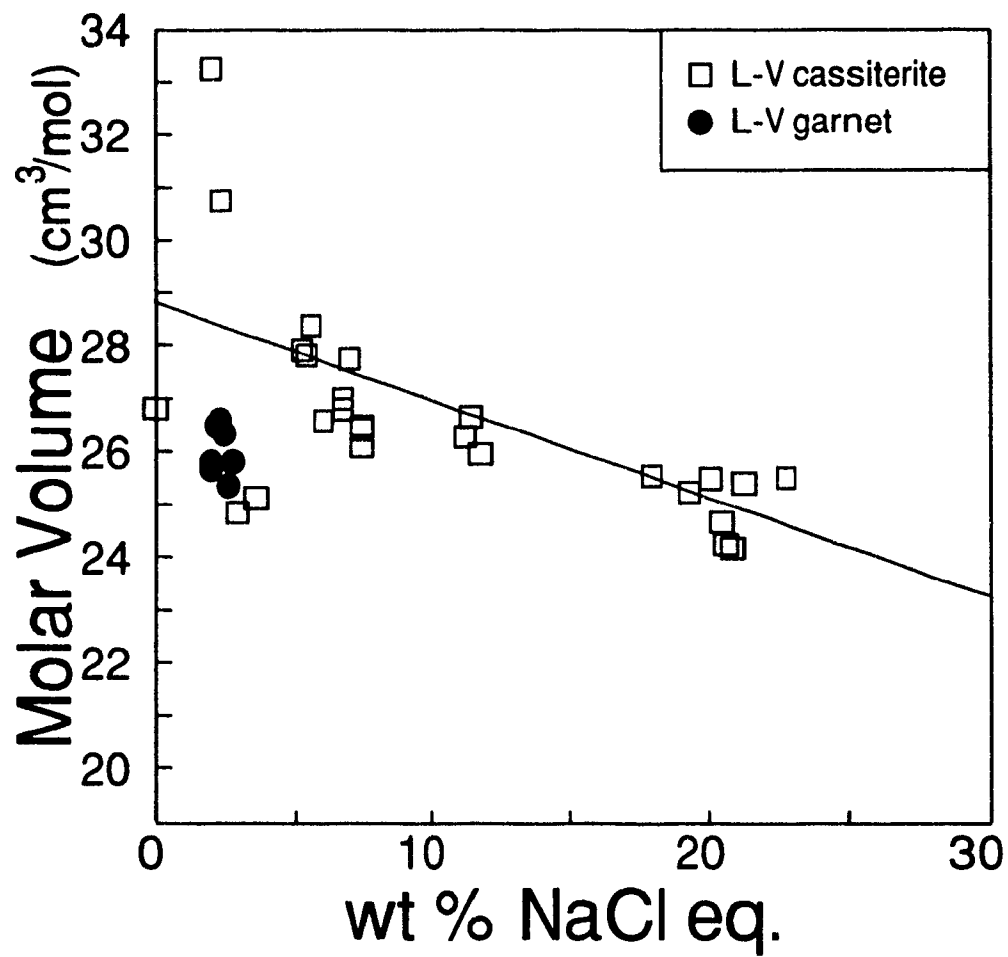


increasing salinity or by isobaric trapping of fluids with decreasing temperature and decreasing salinity (Fig. 9). The trend on Figure 5 is most reasonably interpreted to reflect trapping of fluids that were increasingly more saline at decreasingly lower pressures. This is supported by the observation that the high salinity fluid inclusions are late. Fluid composition can also vary via second boiling. However, if the vapour saturation of a granitic melt occurs at pressures greater than approximately 1 kbar (as discussed below), the initial fluid will have the highest salinity and the salinity of the fluid will decrease as successive aliquots of water are removed from the melt (Candela, 1989; Cline and Bodnar, 1991). This succession is the opposite of that observed at Nong Sua and thus does not explain the salinity variation.

If the mixing occurred such that salinity and pressure varied linearly, this should be evident in the variation of molar volumes with salinity (linear for ideal mixing and parabolic for non-ideal mixing). This appears to be the case for type 1 inclusions with salinities of approximately 6 to 26 wt % NaCl (Fig. 10) and suggests that such mixing did indeed occur. However, the molar volumes of the lowest salinity inclusions (primary garnet-related) lie below the mixing trend. A possible interpretation for this is that they were trapped at pressures higher than that defined by the mixing trend.

The origin of aqueous-carbonic fluid inclusions is less clear, in part because the constraints on timing from fluid inclusion petrography are less definitive. The

Figure 10. Molar volume versus salinity relationships for type 1 inclusions in garnet and cassiterite and type 4 inclusions in quartz. The molar volumes where calculated using the equation of Zhang and Frantz (1987).



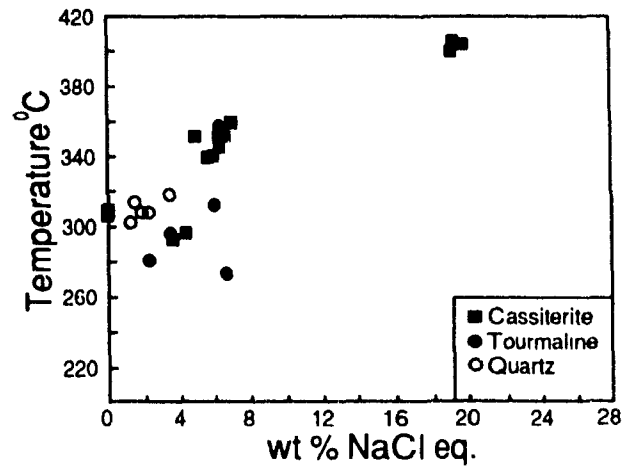
two most important questions concerning the aqueous-carbonic inclusions are: 1) Did mixing or fluid immiscibility have a role in controlling fluid compositions? and 2) What was the origin of the carbonic fluids ?

Plots of homogenization temperature versus salinity for type 2 (homogenization to liquid) and type 3 (homogenization to vapour) aqueous-carbonic inclusions are shown in Figure 11. Type 2, and to a lesser extent type 3, inclusions display salinity-homogenization temperature variations similar to that of type 1 inclusions (Fig. 5). This suggests mixing of CO₂ with aqueous (type 1) fluids (i.e., three component mixing). Where only small quantities of CO₂ mixed with aqueous fluids (high and low salinity), the CO₂ in the resulting fluid inclusions is evident by clathrate formation in type 1 inclusions. Where greater quantities of CO₂ mixed with aqueous fluids the resulting fluid inclusions are three-phase aqueous-carbonic (type 2). The vapour-rich carbonic inclusions, that have only a single visible phase (type 3), potentially represent the end-member carbonic composition for aqueous-carbonic mixing.

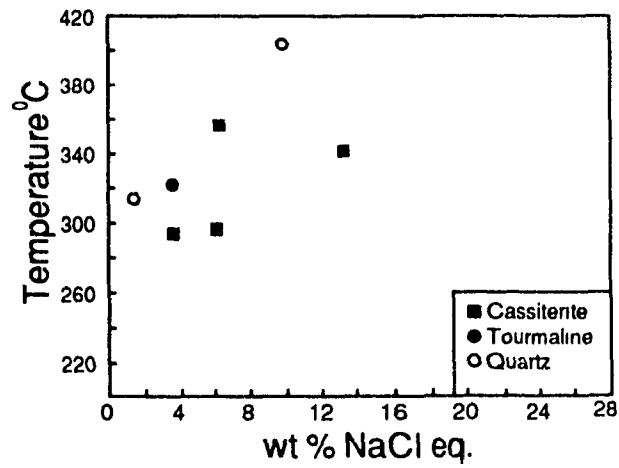
The presence of carbonic-rich and aqueous-rich fluid inclusions is potentially explained by fluid immiscibility. However, if a single fluid becomes immiscible, the carbonic vapour is enriched in CH₄ and the aqueous fluid is enriched in NaCl (Ramboz et al. 1982). The fact that carbonic and aqueous compositions are independent of inclusion type indicates that a single fluid did not undergo phase separation (immiscibility). Although phase separation did not play a role, it is still

Figure 11. Homogenization temperature versus salinity for type 2 (L-L-V) and Type 3 (V-L-L) inclusions. All salinities were determined from clathrate melting temperatures and the equation of Bozzo et al. (1973).

Type 2 (L-L-V) inclusions



Type 3 (V-L-L) inclusions



possible that the carbonic and aqueous fluids were immiscible. If mixing of aqueous and carbonic fluids from different sources occurred at conditions below the two-phase solvus, the H₂O-rich fluid could have dissolved CO₂ up to saturation, or the CO₂-rich fluid could have dissolved H₂O up to saturation. In this manner the H₂O-rich fluid could have a significant carbonic component, and if undersaturated (above the two phase solvus), the X_{CH₄} could be similar to that of the CO₂-rich fluid. Pressures and temperatures from isochore extrapolation, discussed below, are on the order of 2.7 to 3.8 kbar and 650°C. These values are above the two-phase solvus in the system H₂O-CO₂-NaCl (cf. Bowers and Helgeson, 1983). However, the addition of CH₄ (or N₂) (Holloway, 1984a) or CaCl₂ (Zhang and Frantz, 1989; Plyasunova and Shmulovich, 1991) to the fluid expands the two phase solvus (also see Johnson, 1991) and therefore fluid immiscibility (but not boiling) potentially did play a role in controlling fluid composition.

A problem in the interpretation of type 2 inclusions is that at a given salinity the homogenization temperatures of aqueous and aqueous-carbonic inclusions are similar, whereas phase relationships in the H₂O-CO₂-NaCl system indicate that mixing in the one-phase field should result in higher homogenization temperatures of the aqueous-carbonic fluid inclusions (Gehrig, 1979). This is illustrated by one cluster of five aqueous-carbonic inclusions that all homogenize critically at 350° ± 10°C and have compositions of 6.4 ± 0.2 wt % NaCl eq. (estimated from clathrate melting temperatures) and X_{CO₂eq} of 0.06 ± 0.02 (the CO₂/CO₂+CH₄ ratio is 0.01).

These critical homogenization temperatures are close to those in the $\text{H}_2\text{O}-\text{CO}_2$ system but are too low for the system $\text{H}_2\text{O}-\text{CO}_2$ -6 wt % NaCl (Gehrig, 1979). We conclude that an additional component is present in type 2 inclusions which lowers the clathrate melting temperatures, i.e., the estimated salinities are too high.

The most likely sources of CO_2 and CH_4 are the pegmatite melt or the country rocks and the fluids contained within them. Unaltered and tourmalinized metasedimentary rocks from Nong Sua and their regional metamorphic equivalents were analyzed for carbon by the LECO method in order to evaluate whether the metasedimentary rocks could have been a source of carbon. Unaltered pelitic sandstone at Nong Sua contains from 0.06 to 0.21 wt % total C (ave. 0.13, $n=5$) and the regional metamorphic equivalents contain from 0.04 to 6.50 wt % total C (ave. 1.22, $n=6$). By contrast the carbon abundances in the tourmalinized are much lower, 0.02 to 0.03 wt % total C ($n=2$). This suggests that carbon was released during alteration and that the metasedimentary rocks were a plausible source for carbonic fluids. Wilkinson (1991) proposed a similar explanation for the carbonic fluid inclusions observed in the Sn-W deposits of Cornwall. A pegmatitic source of carbon is less likely. Only small quantities of CO_2 are soluble in granitic melts at moderate pressures (Holloway, 1984b), thus the only mechanism by which magmatic CO_2 could have formed a CO_2 -rich fluid (type 3 inclusions) is by boiling. Boiling has been demonstrated above to have been implausible making it unlikely that CO_2 and CH_4 had a magmatic source. This conclusion is supported by a lack

of CO₂ in the primary fluid inclusions hosted by garnet, which are interpreted to be representative of the orthomagmatic fluid. If CO₂ coexisted with the orthomagmatic fluid, either temperatures or pressures well below that of the granite solidus are implied (cf. Bowers and Helgeson, 1983).

STABLE ISOTOPES

Oxygen isotopic compositions of quartz, plagioclase, muscovite, cassiterite and wolframite were determined at the University of Saskatchewan on a Fin Mat 251 mass spectrometer using the BrF₃ extraction technique of Clayton and Mayeda (1963). NBS-28 quartz was used as a standard and the analytical error is ± 0.1 per mil. The δD composition and water contents of muscovite were determined using the method of Bigeleisen *et al.* (1952) as modified by Kyser and O'Neil (1984). Analyses were compared to the NBS-30 biotite standard and the D/H ratio is accurate to within ± 1 per mil.

Primary muscovite from both aplite and pegmatite have $\delta^{18}O$ values of 9.8 to 10.2 per mil (Table 2). Temperatures calculated from quartz-muscovite oxygen isotope thermometry are dependent on which equations are used (Longstaffe, 1982). It has been shown by Linnen and Williams-Jones (Chapter 3) that primary muscovite re-equilibrated with fluids at subsolidus temperatures. We have therefore calculated temperatures from primary quartz-muscovite pairs using the equations of Clayton *et al.* (1972) and O'Neil and Taylor (1969), which are valid up to

TABLE 2
STABLE ISOTOPE DATA

Sample	Description	$\delta^{18}\text{O}_{\text{qtz}}$	$\delta^{18}\text{O}_{\text{musc}}$	$\delta^{18}\text{O}_{\text{cass}}$	$\delta^{18}\text{O}_{\text{plag}}$	$\delta^{18}\text{O}_{\text{wolf}}$	$\delta\text{D}_{\text{musc}}$
PK-17e	qtz-feld-musc-tour pegmatite	12.3	10.1		7.2		-91
PK-22	qtz-tour-wolf-cass vein	12.4				6.0	
PK-25	qtz-feld-musc-tour -gar aplite	12.3	9.8				
PK-33	qtz-feld-musc-tour pegmatite	12.4	9.8		6.1		-106
PK-35b	qtz-cass-wolf vein secondary musc	11.5	8.0	5.6	5.6	5.9	-88
PK-36	disseminated cass secondary musc	12.8	8.5	5.9			
PK-43	qtz-feld-musc-tour pegmatite	12.4	9.8		2.4		
PK-45	disseminated cass secondary musc	12.9	9.4	6.4			
PK-49	contact cass secondary musc	12.6	8.1	5.7			-85
PK-53	qtz-feld-tour-cass vein	12.3		5.6			
PK-58	qtz-feld-musc-tour -gar aplite	12.7	10.2				
PK-62	contact cass secondary musc	11.5	8.5	4.9			-85

Sample	^1T ($^{\circ}\text{C}$)	$^1\delta^{18}\text{O}_{\text{fl}}$	^2T ($^{\circ}\text{C}$)	$^2\delta^{18}\text{O}_{\text{fl}}$	^3T ($^{\circ}\text{C}$)	$^3\delta^{18}\text{O}_{\text{fl}}$	$^4\delta\text{D}_{\text{fl}}$
PK-17e	492	+9.9					(-72)
PK-25	432	+8.9					
PK-33	415	+8.7					(-78)
PK-35b	303	+4.7	460	+ 9.8	450	+8.8	[-65]
PK-36	239	+3.3	395	+10.0	395	+8.6	
PK-43	415	+8.7					
PK-45	303	+6.1	420	+ 9.8	415	+9.2	
PK-49	226	+2.4	395	+ 9.8	395	+7.5	[-55]
PK-53			405	+ 9.7	405	+8.4	
PK-58	432	+9.3					
PK-62	358	+6.4	415	+ 9.1	410	+9.1	[-55]

Abbreviations: qtz - quartz, feld - feldspar, fl fluid, gar garnet, tour - tourmaline, musc - muscovite, cass - cassiterite, wolf - wolframite

¹ Calculated using equations for muscovite-water fractionation from O'Neil and Taylor (1969) and quartz-water from Clayton *et al.* (1972).

² Calculated using equations for quartz-cassiterite-water fractionation from Zheng (1991). ³ Calculated using the quartz-cassiterite equation of Alderton (1989) and quartz-water equation of Clayton *et al.* (1972).

⁴ $\delta\text{D}_{\text{fl}}$ calculated from the equation of Suzuoki and Epstein (1976) at the temperature determined by ($^1\text{qtz-musc}$) and oxygen isotope fractionation and [400°C].

500°C. These temperatures range from 415° to 492°C (Table 2) and are within 75°C of the temperatures estimated by the plagioclase-muscovite thermometer of Green and Usdansky (1986) (Table 3). At these temperatures the $\delta^{18}\text{O}$ values of the coexisting fluid are calculated to have ranged from +8.7 to +9.9 (Table 2). The δD values of two muscovite samples from pegmatite are -91 and -106 per mil (Table 2). The δD values for the coexisting fluid are -72 and -78 per mil (at 492° and 415°C, respectively) using the equation of Suzuoki and Epstein (1976). Magmatic fluids generally are interpreted to have a $\delta^{18}\text{O}$ composition of +5.5 to +10.0 per mil and δD composition of -50 to -85 per mil (Taylor, 1979). The fluid composition at Nong Sua, estimated from primary muscovite thus is consistent with a magmatic origin.

The $\delta^{18}\text{O}$ values of aplite- and pegmatite-hosted plagioclase range from 6.1 to 9.4 per mil. These values are up to 6.3 per mil lower than those of the coexisting quartz, which range from 12.3 to 12.9 per mil (Table 2). At magmatic temperatures, 600° to 700°C, the fractionation between quartz and feldspar is <2 per mil (Clayton et al., 1972, O'Neil and Taylor, 1967). The large fractionation of $\delta^{18}\text{O}$ between quartz and plagioclase therefore indicates isotopic disequilibrium. The compositions of the feldspars were interpreted previously to reflect re-equilibration (Linnen and Williams-Jones submitted; Chapter 2), hence, the isotopic compositions confirm the proposed re-equilibration.

Our isotopic values are similar to those obtained by Kerrich and Beckinsale

TABLE 3
ISOTOPE AND PLAGIOCLASE-MUSCOVITE GEOTHERMOMETRY

Sample	Rock Type	Plag-Musc	$\delta^{18}O_{qtz-musc}$
		$T_{mean} (^{\circ}C)$	$T (^{\circ}C)$
PK-17e	pegmatite	476	492
PK-25	aplite	401	432
PK-33	pegmatite	489	415
PK-43a	pegmatite	472	415
PK-58	aplite	413	432

Plagioclase-muscovite thermometry from Green and Usdansky (1986) (cf. Linnen and Williams-Jones; Chapter 3) and oxygen isotope thermometry from Clayton et al. (1972) and O'Neil and Taylor (1969).

(1988), who analyzed samples from granites in the western Thai granite belt and obtained $\delta^{18}\text{O}$ values that range from 11.6 to 13.6 per mil for quartz and from 2.4 to 11.6 per mil for feldspar. The $\delta^{18}\text{O}$ depletion of feldspar in that study is also reflected in whole-rock compositions that range as low as 5.4 per mil.

Most of the $\delta^{18}\text{O}$ values for cassiterite, regardless of its mode of occurrence, range from 5.6 to 5.9 per mil (Table 2) which suggests that vein, disseminated, and contact styles of mineralization all formed at, or re-equilibrated to, a similar temperature. One sample of contact-style cassiterite has a lower value, 4.9 per mil, which, like the lower value of quartz in that sample (11.5 per mil), may indicate involvement of external isotopically lighter fluids. One sample of disseminated cassiterite has a higher value, 6.4 per mil; however, the quartz in this sample also has a higher value (12.9 per mil). The isotopic fractionation of oxygen between cassiterite and quartz and water has recently been theoretically determined by Zheng (1991). The temperatures calculated from these equations compare very favourably with those estimated from the empirical quartz-cassiterite equation of Alderton (1989). Temperatures of the three styles of mineralization, calculated from quartz-cassiterite fractionation, range from 395° to 460°C and the theoretical and empirical equations agree to within 10°C (Table 2). Using these temperatures and the quartz-water equation of Clayton *et al.* (1972), the $\delta^{18}\text{O}$ composition of the coexisting fluid is estimated to have been +7.5 to +9.2 per mil. If the equation of Zheng (1991) is used, the $\delta^{18}\text{O}$ values for the fluid range up to +10.0 per mil (Table

2). In light of the fluid inclusion evidence presented below, we suggest that cassiterite crystallized at higher temperatures and that the $\delta^{18}\text{O}$ composition of cassiterite represents re-equilibration at lower temperature.

The isotopic composition of quartz from veins and contact-style mineralization is similar to that of magmatic quartz; the $\delta^{18}\text{O}$ values range from 12.3 to 12.6 per mil (Table 2). This suggests that the veins formed from magmatic fluids and supports the interpretation that the veins formed while the pegmatite was at solidus conditions. One quartz separate from a vein and another separate from contact mineralization have lower $\delta^{18}\text{O}$ values, both of 11.5 per mil, which suggests mixing of magmatic fluids with a minor component of an isotopically lighter fluid during mineralization. Albite from a vein gave the lowest $\delta^{18}\text{O}$ value for plagioclase, 5.6 per mil (Table 2), and like plagioclase from the pegmatite, the large fractionation between quartz and plagioclase indicates that they are not in isotopic equilibrium. Muscovite associated with cassiterite mineralization has $\delta^{18}\text{O}$ values that are significantly lower than those from unmineralized aplite and pegmatite, 8.1 to 9.4 per mil (Table 2). Considering that these muscovites are observed to have replaced feldspars, the lower $\delta^{18}\text{O}$ values probably reflect the influx of isotopically lighter fluids. This is supported by the $\delta^{18}\text{O}$ value of secondary muscovite (replacement of albite) from a vein which has the lowest value of the samples analyzed, 8.0 per mil.

Significantly lower temperatures were calculated from quartz-muscovite pairs

from secondary muscovite or that associated with mineralization. These range from 226° to 358°C (Table 2) and reflect the late formation of secondary muscovite, and hence, a lack of equilibration between quartz and muscovite, i.e. the calculated temperatures are not meaningful. Since phengitic substitution in muscovite is indicative of lower temperature alteration (Monier and Robert, 1986), the M^{+2} cation totals were plotted against the $\delta^{18}\text{O}$ compositions of muscovite (Fig. 12). In a closed system $\delta^{18}\text{O}$ and M^{+2} compositions should increase sympathetically as temperature drops. This is not the case at Nong Sua (cf. the reference line on Fig. 12) where the subsolidus magmatic fluids must have mixed with external fluids that had lower $\delta^{18}\text{O}$.

The δD isotopic compositions of secondary muscovites are -85 to -88 per mil (Table 2). The temperature of formation of these muscovites is unknown but given that magmatic muscovite (and cassiterite) re-equilibrated down to approximately 400°C this temperature was used with the fractionation equation of Suzuoki and Epstein (1976) to estimate a δD fluid composition of -55 to -65 per mil. The relatively low $\delta^{18}\text{O}$ and high δD values estimated above are consistent with external fluids of metamorphic or evolved meteoric origins.

An alternate explanation is that the stable isotopic data reflect fluid evolution in a closed system with a very low fluid:rock ratio (cf. Campbell *et al.*, 1984). Figure 13 shows the isotopic compositions of fluids coexisting with a magmatic muscovite ($\delta^{18}\text{O}$ 10.0, δD -100) at 400° to 650°C. This illustrates that in a closed

Figure 12. Octahedral Fe+Mg+Mn versus δO_{18} compositions for magmatic and hydrothermal muscovite. Solid squares represent primary muscovite from pegmatite, solid circles primary muscovite from aplite and open circles secondary muscovite.

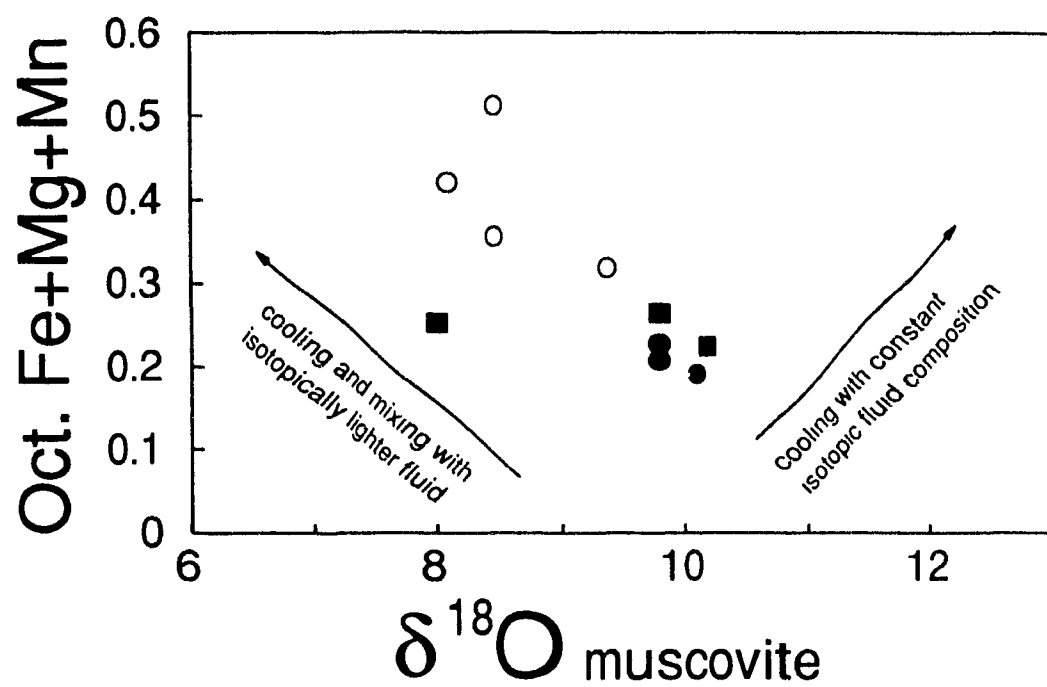
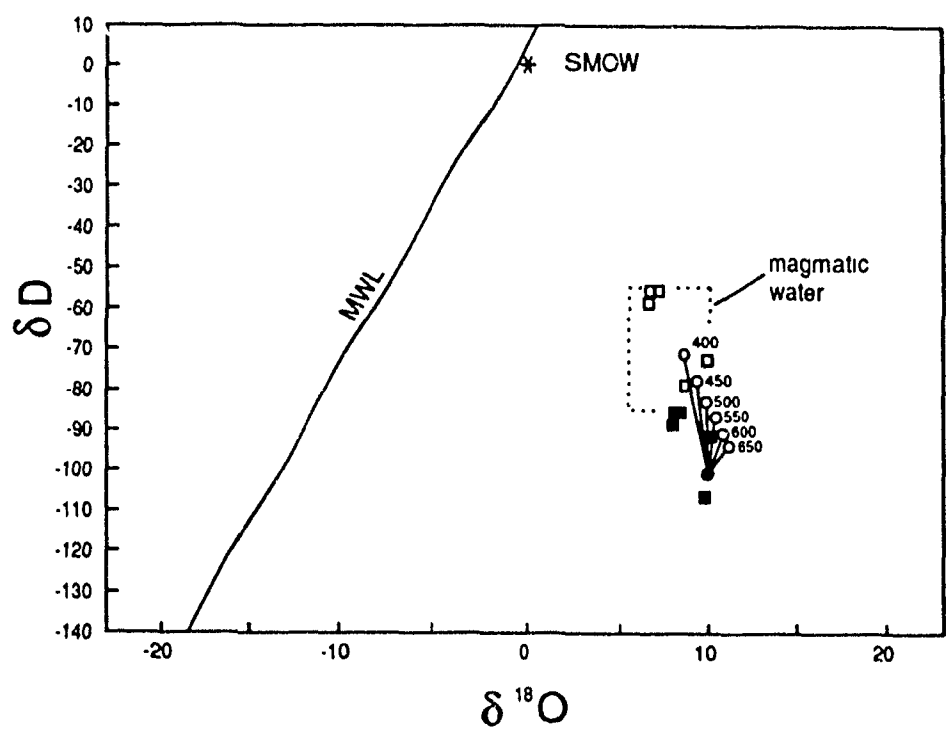


Figure 13. Isotopic compositions of muscovite and calculated fluid compositions.

The solid circle represents a magmatic muscovite. The open circles are the fluid compositions in equilibrium with the magmatic muscovite at 400° to 650°C. The solid squares are the muscovite data from Nong Sua and the open squares are the calculated fluid compositions at 492°, 415° and 400°C (cf. Table 2).



system with a constant whole-rock isotopic composition (very low fluid:rock ratio), cooling results in the coexisting fluid having increasingly lighter $\delta^{18}\text{O}$ and heavier δD compositions. This is very similar to the trend interpreted for the compositions of the fluids at Nong Sua. The lack of widespread alteration suggests that the fluid:rock ratio was low, but Ca-Mg enrichment in tourmaline (Linnen and Williams-Jones, submitted; Chapter 2) and Ca in fluid inclusions indicate an involvement of external fluids. It is most likely that both explanations are partially correct; that the fluid:rock ratio was low, the pegmatite evolved largely as a closed system, but that some external fluids with isotopically lighter oxygen and heavier hydrogen were involved.

PEGMATITE-FLUID EVOLUTION

Linnen and Williams-Jones (submitted; Chapter 2) showed that peraluminous tourmaline-rich pegmatite crystallized early and that the melt evolved toward more alkalic compositions. The concentration of tin in the pegmatite melt at Nong Sua increased with fractionation (increasing alkalinity) and locally, the late-stage melt was saturated in cassiterite (op. cit.). Further support for temporal and/or spatial variation of the tin concentration in the aplite-pegmatite melt is provided by the distribution of tin in garnet (Linnen and Williams-Jones, Chapter 3). Manganese-iron zoning is observed in all garnet types; however, only garnet with Mn-rich cores defined by fluid inclusions contains significant tin (for garnet with a

comparable X_{Spess} composition but lacking fluid inclusions the tin concentration is below the detection limit). The compositional similarity of the different garnet types suggests that they all formed by magmatic processes. The high tin contents in garnet that host fluid inclusions suggest that during the late magmatic vapour-saturated stage, the tin concentration in the melt was high. Garnet that lacks fluid inclusions is tin-poor and probably crystallized earlier, when tin abundances in the melt were much lower.

A magmatic origin of cassiterite is also indicated by cassiterite-tourmaline textural relationships (Linnen and Williams-Jones, Chapter 3). Tourmaline abundances define magmatic layering in the banded pegmatite and the preservation of compositional zoning in all tourmaline types indicates that this mineral did not re-equilibrate with later fluids. The overlap in composition of tourmaline hosted by aplite-pegmatite with that associated with vein or disseminated mineralization suggests that all tourmaline crystallized under similar P-T-X conditions. Since cassiterite-tourmaline intergrowths indicate that these two minerals are coeval, the disseminated cassiterite must also have crystallized at conditions similar to that of magmatic tourmaline.

Whether the transition from magmatic to hydrothermal regimes at Nong Sua was continuous or involved distinct silica-rich and aqueous-rich fluids is difficult to ascertain. There is a continuum of tourmaline and muscovite compositions from demonstrably magmatic to hydrothermal conditions; however, it is not known

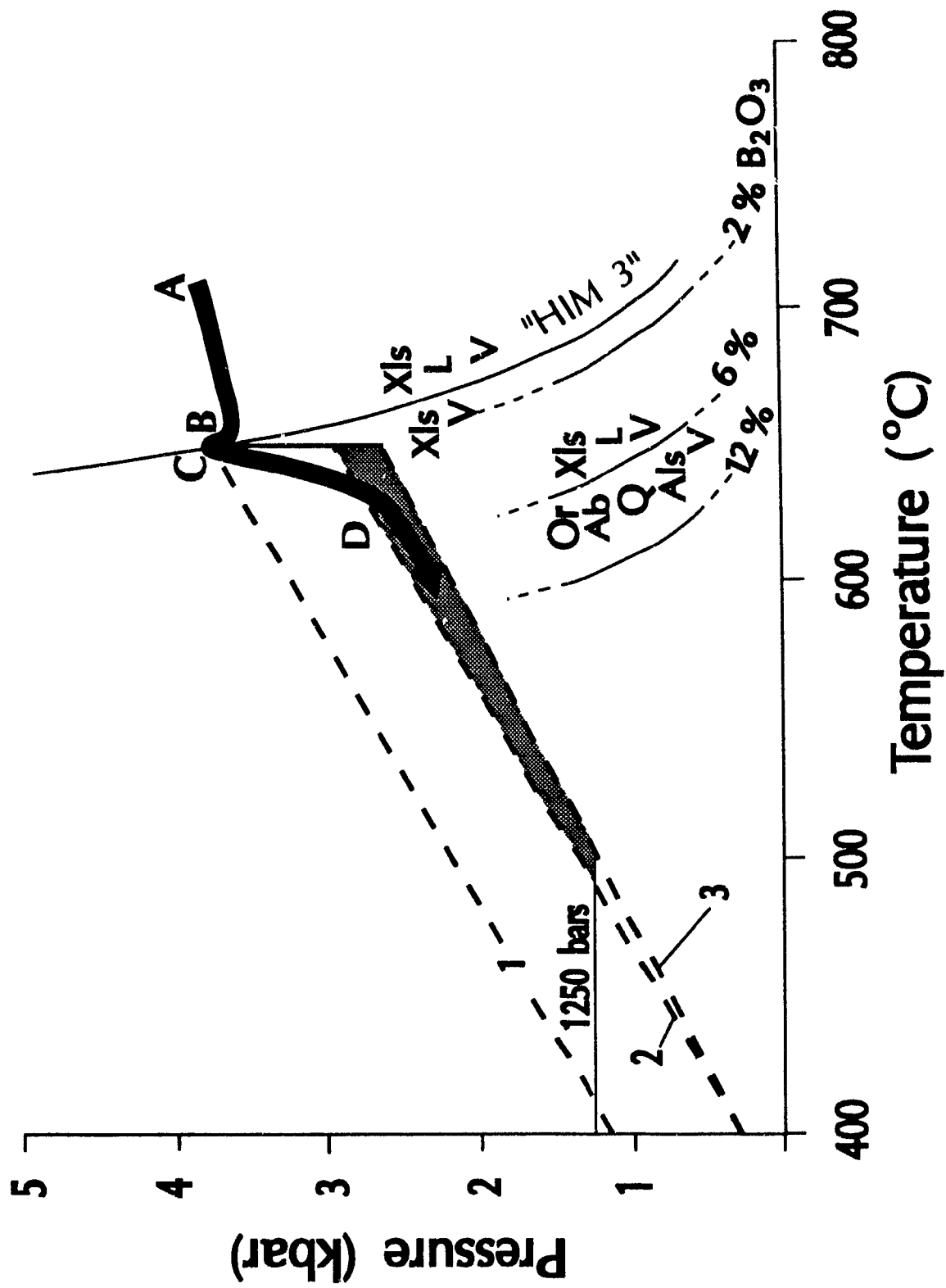
whether mineral-melt and mineral-fluid partition coefficients are significantly different. Hence, the continuum of mineral compositions is not unequivocal evidence for a magmatic-hydrothermal continuum. Evidence for a magmatic-hydrothermal continuum has typically come from "solution-melt" inclusions such as described by London (1990) and the references within. At Nong Sua the solid-rich fluid inclusions in tourmaline have many of the characteristics of solution-melt inclusions and may reflect this stage. Unfortunately these inclusions are very rare at Nong Sua and only one such inclusion was analyzed by microthermometry. This inclusion (in tourmaline from pegmatite with disseminated cassiterite) had a salinity of 5 wt % NaCl eq. and the vapour homogenized at 127°C, but the inclusion decrepitated at 175°C. The low salinity is typical of solution-melt inclusions and is consistent with the salinity of the subsequent orthomagmatic fluids at Nong Sua. We therefore tentatively suggest that the magmatic-hydrothermal transition at Nong Sua could have been a continuous process, although further documentation of the solution-melt inclusions is required.

Irrespective of whether the evolution of the Nong Sua pegmatite involved a solution-melt stage, the earliest hydrothermal event (vapour saturation of the melt) is preserved as primary fluid inclusions in garnet. Since the garnet is magmatic and the fluid inclusions are primary, the pressure and temperature of the crystallization of the pegmatite can be estimated from the intersection of an isochore representative of these inclusions and the granite solidus. We have used the "HIM

HIM 3 solidus of Benard et al. (1985) which is a tourmaline-bearing leucocratic granite with 1.98 and 1.36 wt. % normative anorthite and corundum, respectively, and thus has an analogous composition to the Nong Sua pegmatite (Linnen and Williams-Jones, submitted; Chapter 2). The intersection occurs at approximately 650°C and 3.8 kbars (Fig. 14). The isochore has been extrapolated beyond the 3 kbar upper limit of Zhang and Frantz (1987). However, the position of the isochore can be considered to represent the maximum pressure owing to the tendency for isochores to flatten at higher pressure (Fisher, 1976). Lower pressures and temperatures are obtained if the tourmaline-free B₂O₃ solidi of Pichavant and Manning (1984) are used. According to Benard et al. (1985), boron has only a minor effect on the granite solidus if the boron content of the melt is controlled by tourmaline crystallization. Since tourmaline is present in most of the pegmatite units at Nong Sua, the trapping temperature and pressure are probably closer to the intersection with the HIM 3 solidus rather than the tourmaline-free solidi.

Based on the timing of features such as graphic quartz-feldspar intergrowths, Linnen and Williams-Jones (submitted; Chapter 2) proposed that vapour saturation of the melt occurred very late in the crystallization sequence. Prior to vapour saturation the pegmatite pressure probably decreased with cooling (cf. Knapp and Norton, 1981), depicted as 'A' on Figure 14. At vapour saturation of the pegmatite melt ('B' on Fig. 14), pressure undoubtedly increased (cf. Burnham, 1979; Knapp and Norton, 1981), making it likely that the primary fluid inclusions in garnet and

Figure 14. Fluid evolution of the Nong Sua pegmatite. The solidus for tourmaline leucocratic granite is taken from the "HIM 3" solidus of Benard et al. (1985). Tourmaline-free solidi at 2, 6, and 12 wt % B_2O_3 are from Pichavant and Manning (1984). The abbreviations are Ab albite, Als aluminosilicate, L silicate melt, Or orthoclase, Q quartz, V aqueous fluid and Xls crystals. Isochores were constructed using the equations of Zhang and Frantz (1987). Isochore '1' represents primary type 1 inclusions in garnet, Th=300°C & 3 wt % NaCl. Isochores 2 & 3 represent pseudosecondary type 1 inclusions in cassiterite that homogenize at 400°C. Isochore '2' assumes 22 wt % NaCl, isochore '3' assumes 22 wt % $CaCl_2$ and the shaded area represents the range in which the correct isochore lies. The evolution of the pegmatite and aqueous fluids is shown schematically by the heavy line. 'A' represents cooling and decreasing pressure at vapour-undersaturated conditions in the melt. 'B' represents vapour-saturation of the melt and increasing pressure. 'C' represents the trapping conditions for primary inclusions in garnet and cassiterite. 'D' represents the vein event: fracturing of the aplite followed by isothermal decompression and fluid mixing. The 1250 isobar represent the minimum (hydrostatic) pressure.



cassiterite were trapped at a pressure slightly higher than the emplacement pressure of the pegmatite ('C' on Fig. 14). The occurrence of demonstrable and inferred primary fluid inclusions in garnet and cassiterite, respectively, suggests that orthomagmatic fluid coexisted with melt and crystals (including cassiterite). The temperature and pressure determined by the isochore-solidus intersection (650°C, 3.8 kbar) are therefore interpreted to represent the P-T conditions of disseminated cassiterite mineralization. Although tin forms complexes with chlorine at high temperature (Wilson and Eugster, 1990), tin is not partitioned in favour of the vapour at chlorinities less than approximately 1 M, for the P, T, and fO_2 conditions at Nong Sua (Durasova *et al.* 1986, Taylor, 1988). This implies that the magmatic crystallization of magmatic cassiterite was partly caused by the low salinity of the orthomagmatic fluid; at higher salinities tin would have been partitioned into the vapour and cassiterite would not have crystallized from melt.

We propose that fluid overpressure in the pegmatite caused the aplite to fracture. Although tin was partitioned in favour of the melt, the coexisting orthomagmatic fluid probably contained on the order of several hundred ppm Sn (cf. Linnen and Williams-Jones, submitted; Chapter 2). Vein cassiterite and wolframite were deposited from fluids that migrated to the fracture-induced low pressure zones (represented as 'D' on Fig. 14). The pegmatite was still at solidus conditions, did not fracture, and hence does not host vein mineralization. The maximum temperature of vein formation can be estimated by assuming isothermal

decompression (the fluids did not boil, discussed above and, at these temperatures and pressures the effects of adiabatic cooling are negligible, cf. Helgeson and Kirkham, 1974). The fluid inclusion evidence indicates mixing between high and low salinity aqueous fluids. A pressure estimate, assuming isothermal decompression, can therefore be obtained by extrapolating an isochore representing the high salinity pseudosecondary inclusions in cassiterite to the trapping temperature of the primary type 1 inclusions in garnet. The pseudosecondary inclusions are chemically complex, containing Ca, Fe, Mn, Na and K chlorides. Two isochores, assuming 22 wt % NaCl eq. and 22 wt % CaCl_2 eq., both with a homogenization temperature of 400°C, were therefore constructed (using the equations of Zhang and Frantz, 1987) in order to model the P-T projection; the true isochore very likely lies between these two end-members. The intersection of the 650°C isotherm with the 22 wt % isochore is at approximately 2.7 kbar (Fig. 14).

The minimum pressure that could have prevailed during vein formation is that corresponding to the hydrostatic condition, which would have been approximately 1250 bars. At this pressure, the pseudosecondary inclusions would have been trapped at a temperature of approximately 500°C. Hydrostatic pressure is clearly an unreasonable assumption since the veins are contained within aplite, and the fractures could not therefore have been open to the surface. However, molar volume relationships can be used to evaluate whether there was isothermal decompression or decreasing pressure and temperature. The change in the molar

volume of the fluid preserved as primary inclusions in garnet over a pressure drop from 3.8 to 2.7 kbar is calculated to be from 26 to 28.5 cm³/mol. The latter value corresponds to the value of the 3 wt % fluid extrapolated from the mixing trend between 6 and 22 wt % NaCl eq. (Fig. 10). This suggests that a sharp temperature drop did not accompany the pressure drop during fluid mixing, i.e., supports isothermal decompression. However, some inclusions in cassiterite, with a 3 wt % NaCl composition, have molar volumes that range up to 34 cm³/mol, which suggests that (after the initial pressure drop?) there was a decrease of temperature.

Several lines of evidence point to mixing between low salinity magmatic and high salinity external fluids. These include increasing salinities and contents of bivalent cations with time (primary to pseudosecondary to halite-bearing secondary) and late Ca-Mg metasomatism reflected by tourmaline compositions. This is supported by the stable isotope data: the magmatic fluid is interpreted to have had a composition of $\delta^{18}\text{O}$ +8.9 to +9.9 and δD -72 to -78 per mil, whereas that of the external fluid had lower $\delta^{18}\text{O}$ and higher δD values. This suggests that the magmatic fluids mixed with external fluids of metamorphic or evolved meteoric origin, both of which are consistent with the pressures estimated above.

CONTROLS OF Sn-W DEPOSITION

The solubility of metals is a function of variables such as P, T, $f\text{O}_2$, pH, and the availability of ligands that form soluble metal complexes. The controls of

cassiterite and wolframite deposition at Nong Sua are particularly difficult to assess since none of these variables were constant. Potentially one of the most important controls of mineralization at Nong Sua was increasing fO_2 . The carbonic-bearing fluid inclusions that are hosted by cassiterite are enriched in CH_4 relative to those hosted by quartz, and possibly those hosted by tourmaline. Cassiterite solubility at high temperature is strongly dependent on fO_2 (Taylor, 1988; Wilson and Eugster, 1990). It is therefore possible that the low fO_2 (high X_{CH_4}) inclusions hosted by cassiterite may represent the mineralizing fluid and cassiterite precipitation, may have been controlled by mixing with fluids having a higher fO_2 (low X_{CH_4}). This is supported by the observation that late, secondary aqueous-carbonic inclusions in quartz are CO_2 -rich.

The oxygen fugacities of the aqueous-carbonic fluids can be estimated using the equations of Jacobs and Kerrick (1981) and Ohmoto and Kerrick (1977). The fO_2 of the oxidized fluid is calculated assuming a fluid composition of $X_{H_2O}=0.95$, $X_C=0.05$ and $X_{CH_4}=0.0001$ where X_{H_2O} is the molar ratio ($H_2O/(H_2O+CO_2+CH_4)$), X_C is ($CO_2/(H_2O+CO_2+CH_4)$), and X_{CH_4} is ($CH_4/(CO_2+CH_4)$). A fluid of this composition at 650°C and 2.7 kbar has a log fO_2 of -17.5. The most reduced fluid is probably represented by vapour-rich type 3 inclusions. Assuming a composition of $X_{H_2O}=0.1$, $X_C=0.9$ and $X_{CH_4}=0.1$ the log fO_2 is -19.8. This range of oxygen fugacities slightly exceeds the range for the fayalite-magnetite-quartz and nickel-nickel oxide buffers. In this respect the mineralization at Nong Sua is similar to

many other Sn-W deposits (Dubessy et al., 1987). At constant pH, T (650°C), and chlorinity, an increase of fO_2 by 2 log units results in a decrease of cassiterite solubility of approximately two orders of magnitude (Wilson and Eugster, 1990). Although chlorinity was not constant (salinity increased with mixing) increasing fO_2 probably played an important role to the deposition of cassiterite. Increasing fO_2 may have also contributed to wolframite deposition (Haselton and D'Angelo, 1986; Wood and Vlassopoulos, 1989). However, Fe and Mn chlorides were identified in fluid inclusion precipitates (Fig. 7c); therefore it is more likely that mixing between low salinity (orthomagmatic W-rich) and high salinity (Fe- and Mn-rich) fluids caused wolframite mineralization.

There is some fluid inclusion evidence of cooling at Nong Sua. Stronger evidence of cooling is provided by stable isotopes, which indicate re-equilibration down to approximately 400°C. Cassiterite solubility is temperature dependent (Jackson and Helgeson, 1985; Wilson and Eugster, 1990) and cooling may therefore have contributed to its crystallization. Temperature dependence of tungstate solubility is poorly understood but Wood (submitted) suggested that, in pure water at 300° to 400°C, there is a prograde dependence. In addition, Heinrich (1990) proposed that wolframite deposition can be a consequence of cooling. Cooling may therefore also have contributed to wolframite deposition at Nong Sua. There is very little geological evidence for pH changes at Nong Sua, although the contact-style cassiterite is intergrown with coarse-grained muscovite that has replaced

feldspar. Wilson and Eugster (1990) proposed that increasing pH, resulting from muscovite alteration, is a potential depositional mechanism. However, at Nong Sua muscovite alteration is very limited in extent and is never pervasive. This mechanism is therefore not considered to have been an important control of cassiterite deposition. Wolframite deposition, if linked to $\text{HCl} \rightleftharpoons \text{FeCl}_2$ exchange, is favoured by decreasing pH (Eugster, 1986). As wolframite at Nong Sua is intergrown with feldspars, it follows that pH did not control of tungsten mineralization.

It was shown above that decreasing pressure was important to the hydrothermal evolution at Nong Sua. Unfortunately, the effect of pressure on metal solubility is at present poorly understood. Eugster (1986) suggested that decreasing pressure could lead to the precipitation of cassiterite and wolframite. However, the studies of Arutyunyan *et al.* (1984) and Hemley *et al.* (1986) suggest that base metal solubilities (as chloride complexes) decrease with increasing pressure. A similar result was obtained by Foster (1977) for scheelite solubility. However, at the pressures and temperatures interpreted for Nong Sua the dielectric constant of water would have been significantly lowered by decreasing pressure. This would have resulted in increased association (Seward, 1981) and could have significantly increased tungsten solubility if alkali ion pairs, such as NaHWO_4^0 as proposed by Wood and Vlassopoulos (1989), were important. The dielectric constant is also

lowered if non-polar molecules such as CO_2 , CH_4 , or N_2 are added to water. However, there are conflicting opinions as to whether or not this increases metal solubility. It has been argued by Dubessy et al. (1989) that the addition of such molecules increase cassiterite solubility, whereas Gilbert et al. (1991) propose that mineral solubility in general, and scheelite solubility specifically, are decreased. Whether the solubility of Sn and W increases or decreases with changes in the dielectric constant depends on whether these metals are present as simple or complex ions and on the changes in the other components in solution (e.g., wolframite solubility is dependent on Fe and Mn solubility). Until these competing effects can be accurately modeled it is thus not possible to evaluate the effects of decompression (or CO_2 - CH_4 addition) on Sn-W solubility.

CONCLUSIONS

1) The enrichment of Sn in late-stage magmatic garnet and inclusions of tourmaline with a magmatic composition in cassiterite indicate that cassiterite initially crystallized directly from melt, probably in the presence of aqueous fluids.

2) Vapour saturation of the melt occurred at approximately 650°C and caused the pressure to increase to 3.8 kbar. The orthomagmatic fluid had a composition of 3 wt. % NaCl eq. and as a consequence of this composition tin was partitioned in favour of the melt.

3) The fluid overpressure caused brittle failure in the aplite units. Fluids

migrated into the low pressure fracture zones to form quartz-tourmaline veins and cassiterite-wolframite mineralization. In both the vein and disseminated mineralization three component mixing occurred between low salinity orthomagmatic fluids, external high salinity fluids enriched in Ca, Fe, and Mn, and external carbonic fluids. The fluid mixing can be approximated by isothermal decompression followed by cooling.

4) The carbonic fluids were probably derived from metasedimentary rocks and the high salinity fluid is interpreted to have a metamorphic or evolved meteoric origins. Oxygen and hydrogen isotopes support either origin.

5) The dominant control of mineralization was probably fluid mixing: increasing fO_2 in the case of cassiterite and mixing between W-rich and Fe-Mn-rich fluids in the case of wolframite. Any cooling that took place during this mixing would have aided cassiterite-wolframite crystallization.

ACKNOWLEDGEMENTS

This research was carried out with the aid of a grant from the International Development Research Centre, Ottawa, Canada. We would like to thank S. Bunopas, N. Jungyusuk, S. Khositantont, and P. Putthapiban of the Geological Survey Division, Department of Mineral Resources of Thailand for their support and advice during the course of this project.

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CHAPTER 5

GENESIS OF W-Sn-POLYMETALLIC DEPOSITS IN THE PILOK AREA, THAILAND

by

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ABSTRACT

Wolframite, cassiterite and base metal mineralization in the Pilok area are hosted by veins and stockworks in the apical portion of leucocratic granite stocks, close to the contacts with pelitic metasedimentary rocks. Early, pervasive fluids infiltrated biotite, biotite-muscovite and leucocratic granite which resulted in a "two-phase variant" overprint of primary magmatic textures and possible re-melting of the granitic rocks. This event was followed by tourmalinization. A contact metamorphic-metasomatic aureole was developed in the pelitic metasediments, coeval with the early fluid evolution of the granites. Tourmaline, biotite and muscovite metamorphic-metasomatic zones were imposed on muscovite-chlorite zone regional metamorphism. There is a decrease of metasomatic features and an increase in the carbon contents of the metamorphic rocks with distance from granite. The earliest mineralization event is expressed by rare wolframite $\pm$ cassiterite hosted by pegmatite pods in the leucocratic granite, but pegmatite pods are generally barren. The bulk of the wolframite mineralization is hosted by quartz $\pm$ K-feldspar veins that commonly have K-feldspar alteration haloes. This was followed by cassiterite and sulphide (chalcopyrite-sphalerite) mineralization associated with greisen alteration.

Homogenization temperatures and compositions of fluid inclusions hosted by quartz in wolframite-bearing veins are similar to those associated with greisen

mineralization, but the origin of the former is equivocal. Aqueous-carbonic fluid inclusions are common in cassiterite and quartz from the greisen event and are probably primary. They homogenize to liquid or vapour at 275° to 400°C. The X_{CH_4} of these inclusions is variable, ranging up to 0.3, and the inclusions hosted by cassiterite are methane-rich compared to those hosted by quartz. A lack of partitioning of CH_4 into vapour and NaCl into liquid is interpreted as evidence that immiscibility did not cause mineralization, although some immiscibility may have occurred at other stages. Moderately saline aqueous fluids are interpreted to have mixed with the aqueous-carbonic fluids and also mixed with later low temperature-salinity aqueous fluids of meteoric origin. The $\delta^{18}\text{O}$ compositions of quartz, muscovite and cassiterite range from 8.9 to 11.7, 6.1 to 9.3, and 3.4 to 3.7 per mil, respectively, and the δD composition of muscovite ranges from -100 to -134 per mil. Temperatures calculated from oxygen isotope thermometry suggest that cassiterite was deposited at 300° to 450°C. The aqueous-carbonic fluid inclusions are interpreted to have originated either from metamorphism of carbonaceous pelite or from interaction between magmatic or meteoric fluids with metamorphic rocks. It is proposed that low $f\text{O}_2$ aqueous-carbonic fluids remobilized tin and tungsten from granitic rocks and deposited cassiterite and wolframite in response to increasing $f\text{O}_2$, pH, or decreasing temperature.

INTRODUCTION

The western Thai Sn-W province is known to contain abundant pegmatite-hosted deposits (Garson *et al.* 1975; Manning, 1986). This suggests that much of the mineralization in this belt is related to late magmatic processes. In the investigation of one of these pegmatite deposits, at Nong Sua, it was proposed that the pegmatite melt was locally saturated in cassiterite and that subsequent vein-hosted cassiterite-wolframite mineralization crystallized from early orthomagmatic fluids (Linnen and Williams-Jones, Chapters 2, 3 & 4). However, significant vein and stockwork mineralization are also present in the western Thai province. The granites in the Pilok area are host to vein, stockwork and greisen styles of W-Sn-polymetallic mineralization. Abundant tourmaline alteration associated with these deposits suggests that the source granite is compositionally similar to the boron-rich pegmatite at Nong Sua, but the nature of the mineralization indicates the involvement of post-magmatic fluids. The post-magmatic fluid evolution is fundamental to the formation of these types of deposits (cf. Pollard *et al.* 1987), in contrast to pegmatite-hosted deposits, which form at the magmatic-hydrothermal transition. The Pilok deposits were therefore selected as being a western Thai example of a hydrothermal-dominated system. We have established the fluid evolution at Pilok from the timing and distribution of alteration and mineralization and from fluid inclusion and stable isotope data. On this basis we have proposed

a genetic model for the mineralization. We compare this model to those proposed elsewhere for the evolution of similar stockwork-vein deposits at Cornwall and Panasqueira in order to understand the evolution of Sn-W stockwork-vein systems in general.

GEOLOGICAL SETTING

The Pilok area is located approximately 250 km northwest of Bangkok, along the Thai-Burmese border, and hosts stockwork and vein-type Sn-W mineralization. Brief descriptions of the Pilok deposits are provided in Brown et al. (1951), Tantisukrit (1978), Charoensri (1982) and Mahawat (1988) but they have not been previously studied in detail. The deposits occur in two main areas, around the towns of Ban I Thong and Ban I Pu, and to the southeast in the Pha Pae area (Fig. 1). Most of the mines in the Ban I Thong and Ban I Pu area were closed at the time of this study, but the mines in the Pha Pae area were active, producing wolframite, cassiterite and scheelite. Fieldwork and subsequent investigation therefore concentrated on the Pha Pae area. All the deposits are associated with a biotite to leucocratic granite suite, and typically occur in the leucocratic granite at the contacts with metasedimentary rocks.

Charusiri (1989) obtained a total fusion $^{40}\text{Ar}/^{39}\text{Ar}$ date of 72 Ma on biotite (no error limits were given) from granite 8 km northwest of the Pilok mine. The age of mineralization is constrained from a plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of 76.5 ± 1.1 Ma

for muscovite intergrown with cassiterite in a quartz vein, and a total fusion $^{40}\text{Ar}/^{39}\text{Ar}$ age of 74.4 Ma (no error limits were given) for muscovite in a wolframite-quartz vein (Charusiri, 1989). This suggests that the mineralization is both spatially and temporally associated with the intrusion of the granite.

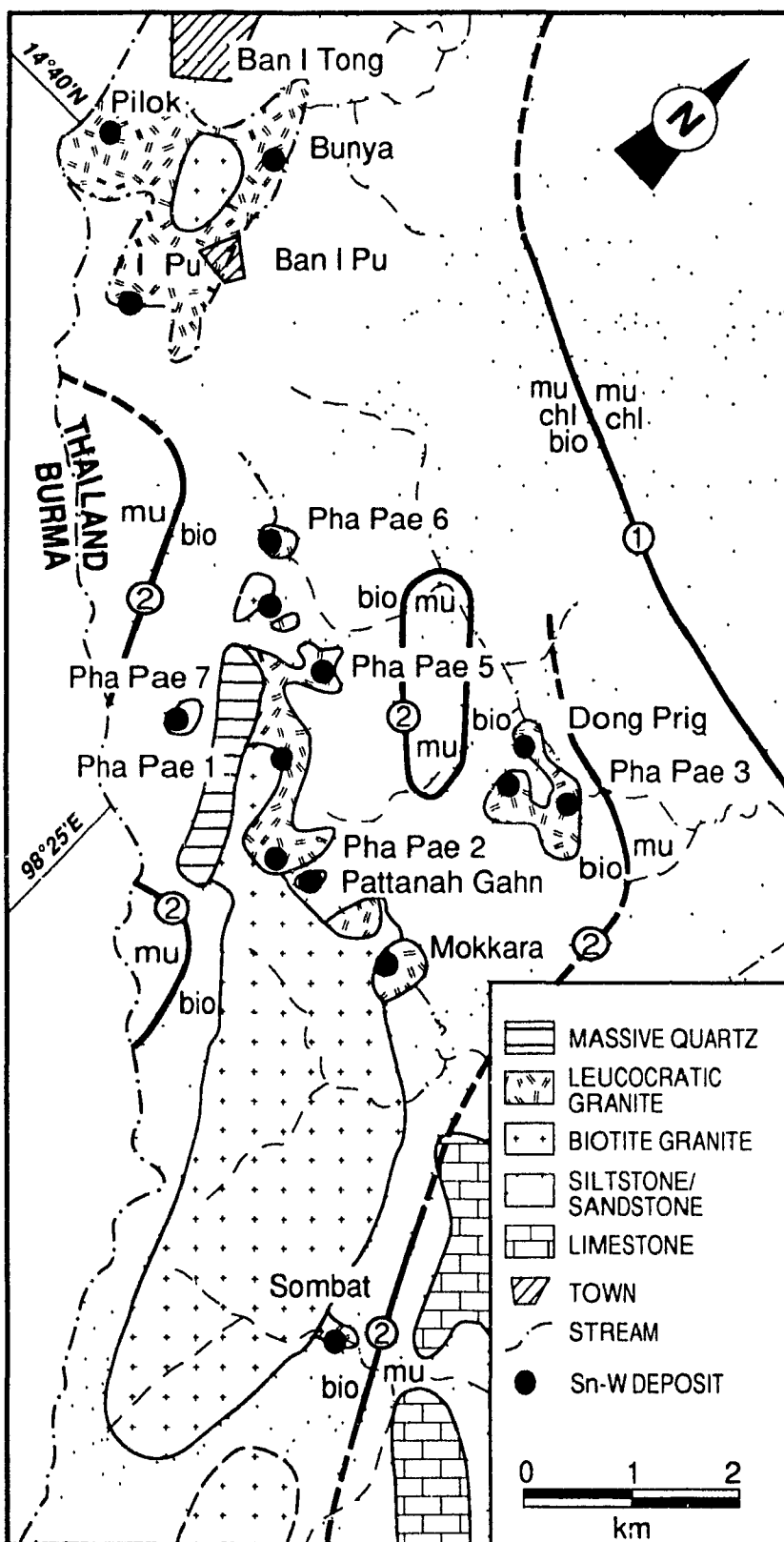
GRANITIC ROCKS

The intrusive rocks in the Pilok area consist of biotite, biotite-muscovite and leucocratic granite and pegmatite. The biotite and leucocratic granites are the most voluminous units whereas biotite-muscovite granite and pegmatite are too small to be represented on Figure 1. The granitic rocks generally outcrop at low elevations so only the upper portion of the system is exposed. The Pilok granites are peraluminous and low in calcium, the $\text{Al}/(\text{Ca}+\text{Na}+\text{K})$ ratio ranges from 1.1 to 1.3, and the normative compositions are close to those of the 1 kbar granite minimum (Lehmann, 1988; Charusiri, 1989; Lehmann and Mahawat, 1989). The granitic rocks contain up to 3000 ppm F (Lehmann, written comm.). However, all of the intrusive phases have undergone textural, and possibly chemical modification as a result of interaction with hydrothermal fluids and/or younger magma.

Biotite Granite

The first intrusive phase is relatively coarse-grained and weakly foliated biotite granite that locally contains K-feldspar megacrysts. Modal analyses of this

Figure 1. A geological map of the Pilok area, showing the setting and location of the principal deposits. Solid line "1" represents the limit of chlorite-muscovite zone regional metamorphism, "2" represents the boundary between the biotite and muscovite contact metamorphic-metasomatic zones.



rock type indicates that it has a monzogranite composition (Charusiri, 1989 and this study). Early coarse-grained K-feldspar and plagioclase show evidence of resorption/dissolution and brecciation. The K-feldspar is generally perthitic, less commonly displays Carlsbad or tartan twinning and contains euhedral plagioclase inclusions. Plagioclase has typically undergone incipient sericite alteration and is locally replaced by fluorite. The primary feldspars may contain subhedral to euhedral inclusions of apatite, or rarely, anhedral inclusions of garnet. K-feldspar, plagioclase and quartz generally show undulose extinction, and where strongly deformed, feldspar twins are displaced. In some places the biotite granite is undeformed, but the primary and secondary features are similar to those in deformed biotite granite.

The coarse-grained primary mineralogy is overprinted by fine-grained plagioclase and quartz with subordinate K-feldspar and biotite. This overprint occurs as veinlets-dykelets that cut individual primary feldspar crystals, or as a matrix around coarser feldspars. Fine-grained plagioclase is rarely twinned, but is commonly myrmekitic and crosscuts and/or replaces primary K-feldspar. Biotite occurs as aggregates of grains along discontinuous veinlets and interstitial clusters; it is rarely in textural equilibrium with coarse-grained feldspar. Biotite has replaced primary plagioclase and is in turn replaced by chlorite. Accessory zircon is also associated with biotite. Minor poikilitic tourmaline is present along fractures or as disseminations.

Biotite-Muscovite Granite

The contact zones between the biotite and leucocratic granites are generally transitional over a few hundred meters, and are characterized by a fine-grained (1 to 2 mm) biotite-muscovite granite. The biotite-muscovite granite is composed of subhedral to anhedral K-feldspar, plagioclase, quartz, biotite, muscovite, and tourmaline, with accessory apatite and zircon. K-feldspar locally has Carlsbad or tartan twins, and perthite and myrmekite is much less common than in the biotite granite. Granophyric quartz-K-feldspar intergrowths occur locally. Biotite and muscovite have some idiomorphic contacts with feldspars, which suggest that they may be primary, but they typically occur as secondary replacement of plagioclase, and are in turn replaced by chlorite. Where intergrown, biotite generally replaces muscovite, but some samples show textural equilibrium.

Leucocratic Granite

Leucocratic granite is fine-grained and occurs at the margins or apices of biotite granite bodies (Fig. 1). Thin leucocratic granite dykes crosscut the biotite granite indicating that the biotite granite is the older intrusive. The leucocratic granite is composed of alkali feldspars and quartz, $\pm$ muscovite $\pm$ tourmaline $\pm$ garnet $\pm$ apatite. This rock type consists of anhedral grains in a matrix of, and/or cut by anhedral finer-grained quartz and alkali feldspars. Locally, an equigranular

hypidiomorphic primary texture is preserved. Quartz and alkali feldspars commonly exhibit undulose extinction and the alkali feldspars may have displaced twins; thus the occurrence of deformation textures and two stages of quartz-feldspar crystallization is similar to that in the biotite granite. K-feldspar is not generally twinned and is rarely perthitic, with very fine albite exsolution lamellae. Albite is typically twinned and, in contrast to biotite granite, incipient sericitization is generally absent. However, argillic alteration (kaolinite group minerals) is common. On the basis of a limited number of electron microprobe analyses the feldspar compositions are Ab_{97} to Ab_{98} and Or_{95} to Or_{97} . Muscovite commonly has a secondary origin, however, in many samples it is in apparent textural equilibrium with primary K-feldspar and albite. Tourmaline is poikilitic and is present along fractures or as disseminations. Garnet is typically anhedral and is intergrown with, or occurs as inclusions in K-feldspar, albite and muscovite. Lehmann and Mahawat (1989) have termed this unit "alkali-feldspar aplogranite" and consider it to be of "transitional magmatic-hydrothermal" origin.

Pegmatite

The leucocratic granite contains abundant pegmatite pods, generally less than 1 m across, and the contacts between pegmatite and leucocratic granite are typically gradational. Pegmatites also occur as dykes (5 to 30 cm thick) with aplitic margins, that crosscut the biotite and leucocratic granites, and rarely the metasedimentary

rocks. Grain size is variable but K-feldspar crystals range up to 20 cm long. Pegmatites are composed of alkali feldspars and quartz $\pm$ muscovite $\pm$ tourmaline $\pm$ beryl $\pm$ apatite $\pm$ garnet. Coarse-grained K-feldspar does not generally exhibit twinning, but may have Carlsbad twins or coarse perthite exsolution. A limited number of electron microprobe analyses indicate that the K-feldspar, compositions range from Or₉₈ to Or₉₀Ab₁₀. Coarse-grained albite has polysynthetic twins, is in apparent equilibrium with K-feldspar and ranges in composition from Ab₉₉ to Ab₇₇Or₂₃. In non-strained pegmatite, fine-grained albite plates (cleavelandite) occur interstitial to, and intergrown with, coarse-grained K-feldspar. Muscovite is typically secondary, although some coarse grains are in apparent equilibrium with alkali feldspars, and may be primary. Less commonly, muscovite is intergrown with quartz in a symplectic-graphic texture. Thin muscovite plates coarsen and radiate outward in optically continuous quartz which may be a primary magmatic feature. Tourmaline generally occurs along fractures but, like muscovite, some grains may be primary. Garnet, beryl and apatite are present as inclusions in, or intergrown with quartz, and the origin of these accessory minerals is equivocal.

Genesis of the Granites

Each of the igneous units show textures indicative of a magmatic and/or hydrothermal overprint. Textures in the Pilok area are similar to those in the Main Range granites of Malaysia (Pitfield et al., 1990) where it was shown that K-

feldspar megacrystic granite may have solidified prior to "disruption" by a fine-grained matrix-replacement. The disruption is believed to be related to "volatile fluxing" and the intrusion of a more differentiated residual granite.

We suggest that the sequence of granite emplacement and modification at Pilok was similar to that proposed by Pitfield et al. (1990): 1) Intrusion of the K-feldspar megacrystic granite. 2) Intrusion of a more differentiated leucocratic granite into a hot but consolidated megacrystic granite. 3) Vapour saturation of the leucocratic granite. The presence of pegmatite pods and graphic and granophyric textures suggest that the leucocratic granite was water-rich. Tourmaline is present dominantly (and possibly exclusively) as a secondary mineral, indicating that boron was probably partitioned into the vapour phase. Since boron increases the solubility of water in granitic melts (Pichavant, 1987), this further suggests a water-rich melt. 4) These fluids (and melt?) pervasively infiltrated the primary granite, in part through hydraulic fracturing as indicated by the microbreccias. Plagioclase, quartz and biotite were the dominant minerals crystallizing during this infiltration (with minor tourmaline?). The common occurrence of myrmekite suggests that the temperature was high and locally the megacrystic granite may have been remelted. This process is similar to the "refusion" proposed by Martin and Bonin (1976). 5) After consolidation of the leucocratic granite a similar sequence was repeated, involving residual melt and orthomagmatic fluids that texturally modified the leucocratic granite. The granophyric textures in these rocks are potentially primary,

but may also reflect local re-melting and re-crystallization of the leucocratic granite. Incipient K-feldspar replacement of albite and albite replacement of K-feldspar are ubiquitous. Thus the fluids that texturally modified the megacrystic and leucocratic granites were very probably metasomatic. However, it is difficult to specifically link feldspar replacement to the magmatic-hydrothermal transition since the leucocratic granite is overprinted by several subsequent stages of alteration.

The presence of two primary alkali feldspars in the leucocratic granite restricts the pressure of emplacement to being greater than approximately 2.5 kbar (Martin and Bonin, 1976). Significantly lower pressures could have been possible if boron and fluorine had depressed the granite solidus, e.g., a melt with 5 wt % B_2O_3 has a minimum pressure of ~800 bars and 4 wt % F ~600 bars (Manning and Pichavant, 1984). The Pilok granites do not show shifts (indicative of high F and/or B) from water-saturated granite minimum compositions toward the albite apex (Lehmann, 1988; Lehmann and Mahawat, 1989). A more realistic minimum pressure for subsolvus granites is ~1500 bars, established for the St. Austell granite by Weidner and Martin (1987).

CONTACT METAMORPHISM-METASOMATISM

The metasedimentary rocks in the Pilok area consist mainly of poorly sorted siltstone-sandstone, and correlate with the Carboniferous to Permian Kaeng Krachan Group (Siribhakdi et al., 1976; Fig. 1). Minor Devonian marble-limestone is also

present but is not known to host any Sn-W mineralization. Regionally metamorphosed chlorite-muscovite facies sandstone-siltstone occur 2 to 3 km from the granite contact. They are composed largely of quartz, muscovite and chlorite and are locally graphitic. Quartz and, less commonly, pelitic (now chlorite) pebbles are supported in a very fine-grained matrix, <0.05 mm. Incipient replacement of chlorite by biotite marks the boundary of the muscovite-chlorite facies of regional metamorphism (Fig. 1). It is not clear whether this replacement is of regional or contact metamorphic origin. Successive muscovite, biotite and tourmaline zones are recognized with decreasing distance from the granite, but as the metasediments have also been affected by hydrothermal alteration the term "zone" is not used in the classic metamorphic sense.

Muscovite zone rocks are composed predominantly of muscovite and quartz with trace to minor amounts of biotite, chlorite and plagioclase. These rocks are slightly more coarse-grained than their regional metamorphic equivalents, <0.05 to 0.1 mm, and the larger muscovite and plagioclase grains are poikilitic. Graphitic horizons occur locally. Chlorite is much less common than in the regional metamorphic rocks and is replaced by muscovite and biotite. Biotite commonly occurs along fractures or with quartz in veinlets, and this is metasomatic. The presence of trace amounts of non-detrital tourmaline is also indicative of metasomatism.

Biotite zone rocks are composed predominantly of biotite, quartz, muscovite

and plagioclase, with variable amounts of chlorite and tourmaline, and very rarely garnet. They are coarser-grained than muscovite zone or regional metamorphic equivalents, generally 0.1 to 0.2 mm and contain sericitized porphyroblasts up to 0.5 mm across. In contrast to the above units, biotite zone metasediments are never graphitic. Biotite occurs in the matrix and along fractures and veins. Muscovite is generally coarse-grained and poikilitic. Andalusite porphyroblasts are reported to occur in the Ban I Thong area (Khositanont, pers. com., 1988). This restricts the pressure of contact metamorphism to <3.8 kbar (Holdaway, 1971), and we suggest that the sericitized porphyroblasts in the Pha Pae area represent replacement of andalusite by muscovite. Disseminated and vein-hosted tourmaline is present locally, as is rare fracture-controlled K-feldspar (otherwise K-feldspar is not present). Chlorite occurs along late fractures and replaces muscovite porphyroblasts. Metasomatic features such as veins, secondary minerals along fractures, the replacement of porphyroblasts and the presence of tourmaline in the biotite zone are much more abundant than in the muscovite zone, suggesting a greater involvement of fluids.

Tourmaline zone rocks are strongly metasomatized. They occur within 10 m of the leucocratic granite and contain up to 80 % tourmaline. These rocks are composed of tourmaline and quartz $\pm$ muscovite $\pm$ rutile $\pm$ plagioclase and the grain size is the coarsest of any of the zones, ranging up to 1 mm diameter.

Granitic systems are generally thought to initiate large hydrothermal systems

(e.g. Cathles, 1977). At Pilok, this would have involved fluid circulation through metamorphic rocks and, as proposed elsewhere (e.g., Williams-Jones, 1982; Linnen and Williams-Jones, 1990; Wilkinson, 1990), fluids probably controlled the contact metamorphism. We suggest that the rarity of metasomatic features in the muscovite zone, their more common occurrence in the biotite zone and the intense metasomatism in tourmaline zone rocks reflect an increasing influence of magmatic and external (meteoric, formational, or metamorphic) fluid-rock interaction.

Analyses of bulk-rock were conducted for CO_2 , SO_3 , Sn, As, Cu, Pb and Zn to determine whether fluids removed or added these components (Table 1). Carbon contents are highest in regional metamorphic and muscovite zones (up to 4.32 % CO_2 in graphitic horizons), and below the detection limit in the biotite zone (<0.05 % CO_2). This is consistent with the observed distribution of graphitic horizons and indicates volatilization or leaching of carbon during contact metamorphism-metasomatism. Tourmaline zone rocks contain minor carbon, which is unexplained. Only the regional metamorphic rocks have sulphur contents above the detection limit (0.01 % SO_3). All muscovite zone and regional metamorphic samples contained <5 ppm Sn. Three of the seven samples from the biotite zone contain 5 to 8 ppm Sn. Although these values are close to the detection limit, they may indicate an interaction between magmatic fluids and metasedimentary rocks. However, tin is significantly enriched in the tourmaline zone, which suggests a magmatic origin for tin and boron. A similar interpretation of the metal distribution

Table 1

**Average Base Metal, Carbon, and Sulphur Abundances
in Metasedimentary Rocks**

	Regional Tourmaline Metamorphic	Muscovite Zone	Biotite Zone	
	n=3	n=8	n=7	n=2
As	27	8	< 5	26
Cu	21	21	23	15
Pb	20	178	30	37
Zn	70	93	116	368
Sn	<5	<5	<5	85
CO ₂	0.73	1.09	<0.05	0.17
SO ₃	0.10	<0.01	<0.01	<0.01

As, Cu, Pb, Zn and Sn were analyzed by atomic absorption. The values are in ppm, with a detection limit of 5 ppm

Total carbon as CO₂ and total sulphur as SO₃ were analyzed by the LECO method. The values are in wt % and the detection limits are 0.05 wt. % for CO₂ and 0.01 wt. % for SO₃

Where samples contained values below the detection limit, a median rather than average value is given for the sample group

in the Cornubian tin province was proposed by Hall (1990). The base metal and arsenic abundances are not significantly different amongst regional metamorphic, muscovite and biotite zones, but we consider the metasediments to have been a potential source of base metals.

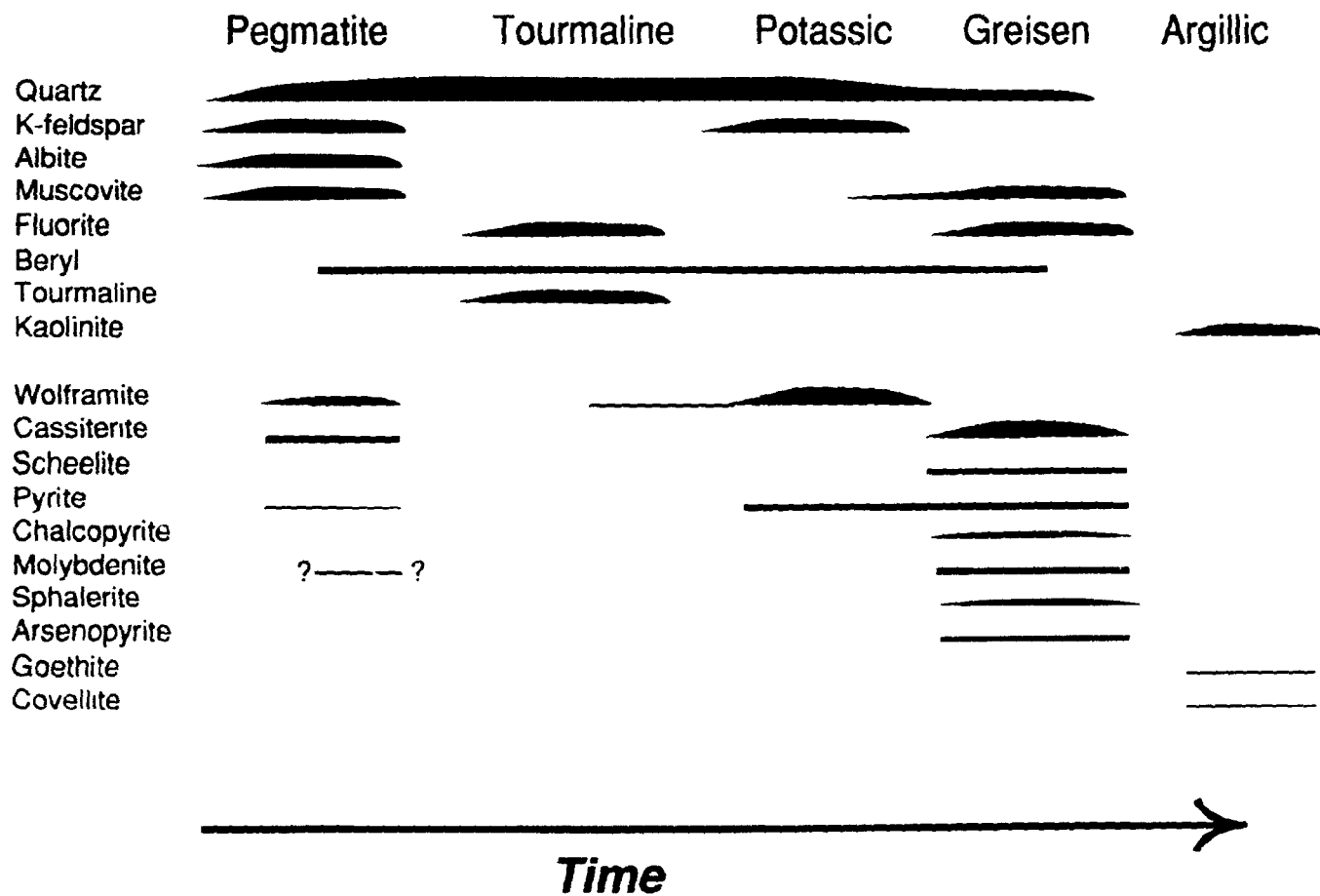
ALTERATION AND MINERALIZATION

Cassiterite is the only tin ore mineral and wolframite is the primary source of tungsten in the Pilok-Pha Pae area; a minor amount of tungsten also occurs in the form of scheelite. Although the abundances of chalcopyrite and sphalerite locally exceed those of cassiterite and wolframite, these minerals are not recovered. The distribution of the ore minerals in heavy metal concentrates presently being recovered is estimated by Charusiri (1989) to be 65% wolframite, 33% cassiterite and 2% scheelite. Almost all the tin and tungsten mineralization is hosted by leucocratic granite and is spatially associated with the contacts between leucocratic granite and metasedimentary rocks (Fig. 1). A minor amount of cassiterite is also hosted by metamorphosed siltstone-sandstone adjacent to leucocratic granite. The paragenesis of oxide, sulphide, and silicate minerals that characterize mineralization and alteration are summarized in Figure 2.

The earliest alteration in the Pilok area is that associated with the magmatic-hydrothermal transition. The local replacement of K-feldspar by albite and replacement of albite by K-feldspar cannot be directly related to veins or fractures

Figure 2. Paragenetic sequence of mineralization and alteration at Pilok.

Paragenesis



and we consider it to represent "magmatic metasomatism". This early metasomatism pre-dates mineralization and fracture-related alteration. The leucocratic granite contains trace amounts of wolframite and cassiterite but these minerals typically occur along fractures and can be placed within the paragenetic sequence established below.

Pegmatites are generally barren but locally contain wolframite $\pm$ cassiterite $\pm$ pyrite, or molybdenite that may reflect the earliest stage of mineralization. Molybdenite invariably occurs along fractures in quartz indicating that it formed at subsolidus conditions, although it may have an early post-magmatic origin. Wolframite, with subordinate cassiterite and pyrite, occurs in pegmatite pods, notably in the Pha Pae 1 and 2 deposits, and K-feldspar or tourmaline haloes are developed around some of these pegmatites. The wolframite is intergrown with quartz, and coarse-grained K-feldspar and muscovite, but where intergrown with K-feldspar, a thin muscovite selvage separates the wolframite from K-feldspar. The relationships between wolframite and muscovite and the presence of pyrite suggest that a magmatic origin (i.e., direct crystallization from melt) is unlikely. Since this mineralization occurs in pegmatites and is not related to veins or fractures, an early post-magmatic origin is probable.

The earliest hydrothermal event that can be related to veins or fractures is tourmaline alteration. Abundant barren quartz-tourmaline veins with trace fluorite, or veinlets that consist solely of tourmaline crosscut the leucocratic granite.

Tourmalinized metasediments occur within 10 m of the leucocratic granite (tourmaline zone, discussed above) and host coeval, tensional quartz veins. Pervasively altered metasediments are massive, containing up to 80 % tourmaline, whereas altered leucocratic granite contains 5 to 10 % tourmaline (along fractures and disseminations). The lower tourmaline abundances in pervasively altered leucocratic granite are undoubtedly related to the lower Fe and Mg contents in this host. Minor wolframite $\pm$ cassiterite mineralization is hosted by quartz-tourmaline veins, indicating that there was a temporal overlap between tourmaline alteration and wolframite-cassiterite deposition. Cassiterite is very rarely associated with tourmaline, except at the Sombat mine (Fig. 1) where it is intergrown with tourmaline and muscovite. However, the tourmaline at this mine has been partially replaced by muscovite, and the cassiterite is potentially related to greisen, rather than tourmaline alteration.

Quartz-tourmaline veins are crosscut by quartz-K-feldspar-wolframite veins and quartz-wolframite veins with K-feldspar alteration haloes. In contrast to tourmaline and greisen alteration, K-feldspar alteration is never pervasive. Quartz-wolframite $\pm$ K-feldspar veins are generally 1 to 10 cm thick, locally up to 1.5 m thick, and host the bulk of the wolframite mineralization. The Sn/W ratio is highly variable amongst different deposits; e.g., the Pattanah Gahn mine is a W-rich deposit and the Dong Prig mine is a Sn-rich deposit (Fig. 1). Wolframite generally occurs as euhedral blades up to 3 cm long, either within the vein or disseminated

in leucocratic granite at the vein selvage. The veins may also contain minor cassiterite, pyrite, beryl, albite and tourmaline. Less commonly, wolframite is intergrown with muscovite, and some quartz-wolframite veins have muscovite alteration haloes. In most of these cases the muscovite occurs along discontinuous fractures at the vein selvage and along fractures that cut vein quartz (i.e., muscovite formation post-dated vein emplacement). However, in other cases the vein and muscovite selvage appear to be coeval, suggesting a temporal overlap between wolframite mineralization and greisen alteration. Virtually all of the scheelite mineralization in the Pilok area occurs as a replacement of wolframite. Scheelite is observed to crosscut wolframite along fractures or in quartz veinlets. In the veinlets, scheelite may be associated with muscovite, pyrite, chalcopyrite, sphalerite or molybdenite.

Greisen alteration overprinted tourmalization and quartz-wolframite-K-feldspar veins. The greisens are composed largely of muscovite and quartz, with minor fluorite and beryl. In partially altered leucocratic granite samples, muscovite has replaced K-feldspar, albite and tourmaline. Greisen alteration hosts the majority of the cassiterite and sulphide mineralization. The sulphides consist of chalcopyrite, sphalerite, pyrite and arsenopyrite, and rarely molybdenite and bismuthinite, and were penecontemporaneous with cassiterite. Sulphides, particularly chalcopyrite, also fill fractures that cut cassiterite. Greisens are generally not related to specific veins; rather they occur pervasively or as haloes around fractures. Quartz coexists

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with muscovite in pervasive alteration zones, indicating that silica saturation occurred during this event. However, cassiterite $\pm$ muscovite $\pm$ sulphides is also observed rarely in etched vugs in quartz-wolframite veins. It seems likely that, during greisen alteration, silica was locally undersaturated, although a supergene origin for the etched vugs cannot be discounted. Cassiterite hosted by quartz veins is commonly secondary, and occurs along fractures with muscovite. In some quartz veins cassiterite is a primary mineral and is coeval with quartz. These veins are commonly shear-related, and locally en-echelon vein sets or slickensides at vein-wallrock contacts are developed. Biotite zone metasediments host a minor amount of cassiterite near the contacts with leucocratic granites. The mineralization consists of thin quartz-cassiterite veins, 1 to 2 cm thick, with muscovite alteration haloes. Chlorite-pyrite along fractures in biotite granite is minor and is not associated with mineralization.

A large, approximately 200 m thick, composite quartz vein crosscuts mineralized leucocratic granite and metasediments in the area near Pha Pae 1 and 7 (Fig. 1). This vein consists of massive quartz, with minor vuggy and brecciated zones and is cut by internal quartz veins <10 cm thick. Smaller quartz breccia-veins are observed elsewhere and a phreatic origin for them was suggested by Charusiri (1989). The large composite vein and breccia-veins are late and clearly not related to mineralization. Their origin is unknown but late massive quartz veins are associated with Sn-W deposits elsewhere in western Thailand, e.g., Takua Pit

Thong and Hub Kapong and, may be related to major strike-slip faults (the Three Pagodas fault in the case of Pilok). Argillic alteration is the youngest hydrothermal stage evident at Pilok. At high elevations (e.g., Pha Pae 2 and 5) clay alteration is very strong, whereas at lower elevations (e.g., Pattanah Gahn and Dong Prig) argillic alteration is absent or weakly developed. It can be argued that the elevation control reflects a supergene origin for kaolinite, or reflects hypogene alteration zoning. Since argillic alteration envelopes have not been observed around fractures or veins we favour a supergene origin. Minor goethite replacement of pyrite and covellite replacement of chalcopyrite is similarly related to weathering or supergene alteration.

FLUID INCLUSIONS

PETROGRAPHY

Six types of fluid inclusions have been recognized in quartz, cassiterite, fluorite and beryl. Type 1 are vapour-rich aqueous-carbonic inclusions (V-L-L and V-L). These inclusions occur in quartz, cassiterite and beryl, typically in isolation or as clusters which indicate that they may be primary. Less commonly they are observed along secondary planes or along short planes within single crystals that are potentially pseudosecondary. Type 2 inclusions are liquid-rich and aqueous-carbonic (L-L-V). They are also commonly present in quartz, cassiterite and beryl,

and have an origin similar to that of type 1 inclusions.

Type 3 are aqueous inclusions (L-V) that are abundant in quartz and fluorite, but are rare in cassiterite and fluorite. Some type 3 inclusions contain carbonic vapour (L-V*), evident from double freezing (cf. Collins, 1979), and carbonic melting and clathrate decomposition temperatures were obtained from these inclusions. Type 3 inclusions are commonly secondary, but are also isolated or occur in clusters and may be primary. The L-V* subset are typically present in the latter mode of occurrence and both L-V and L-V* inclusions are rarely present along short planes within single crystals (pseudosecondary?). Type 4 inclusions are liquid-only aqueous (L). They are always secondary, are hosted solely by quartz, and occur along the same plane as type 3 inclusions with very low vapour/liquid ratios. Type 5 inclusions are aqueous and contain solids other than halite (L-V-S). These inclusions are much less common than type 3 inclusions, are dominantly hosted by quartz and fluorite and are rarely present in cassiterite and beryl. The mode of occurrence of type 5 inclusions is similar to that of type 3 inclusions. Four different solids have been distinguished; S_1 is rounded, generally small and has strong white-yellow birefringence; S_2 is rounded, typically larger than the other solids, and is apparently isotropic; S_3 is rod-shaped, small and has white birefringence; and S_4 is rectangular, small, with strong red birefringence. None of the solids dissolved after heating to 400°C and therefore may be trapped phases. Type 6 inclusions are halite-bearing (L-V-H) and are only present in quartz, where

they are secondary. In addition to halite, they may contain one or two of the above solids.

Inclusions in quartz generally have rounded shapes and range from 5 to 20 μm in diameter. In cassiterite, inclusions generally have rounded shapes, although some type 1 inclusions have negative crystal shapes, and range in size from 5 to 100 μm across. Fluorite-hosted inclusions generally have rounded shapes and range up to 30 μm in diameter. Inclusions in beryl range up to 30 μm in diameter and generally have rounded shapes or rectangular negative crystal shapes.

Microthermometric analyses of fluid inclusions were conducted using a Fluid Inc. stage calibrated with synthetic fluid inclusion standards. Measurements of the phase changes at or below 31°C are accurate to within $\pm 0.2^{\circ}\text{C}$ and the high temperature measurements to within $\pm 2.0^{\circ}\text{C}$. The inclusions in quartz are from quartz-wolframite, quartz-cassiterite and quartz-sulphide veins. There are no apparent differences in the microthermometric results for quartz-hosted inclusions amongst the different vein types. All data from quartz are therefore discussed collectively below. The data for beryl were collected from a quartz-cassiterite-beryl vein and those for fluorite was collected from a vug of uncertain paragenesis.

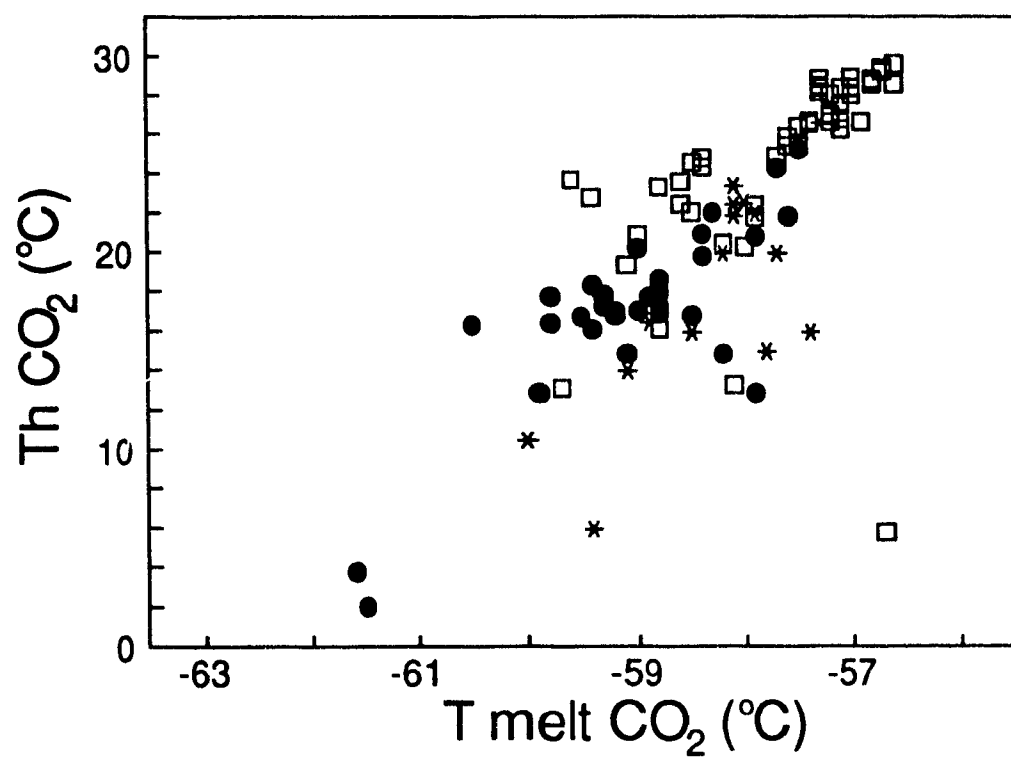
AQUEOUS-CARBONIC INCLUSIONS

The carbonic phase homogenizes to liquid, vapour and critically in quartz, cassiterite and beryl. In quartz the homogenization temperatures are generally

greater than 22°C and the carbonic melting temperature higher than -58°C, but in cassiterite and beryl these temperatures are typically less than 22°C and lower than -58°C (Fig. 3). The carbonic melting temperatures indicate that most inclusions in quartz have compositions close to that of pure CO₂ (T_m_{CO₂} -56.6°C), whereas those in cassiterite and beryl have significant concentrations of additional components such as CH₄ and/or N₂ that lower the melting temperature. Additional components and variable compositions are also indicated in some inclusions by critical carbonic phase homogenization temperatures of 16.3°, 16.6°, 21.0° and 22.4°C, which are lower than that of pure CO₂ (31.1°C). If modeled in the CO₂-CH₄ system, the critical homogenization temperatures indicate X_{CH₄} values of up to ~0.2 (cf. Heyen et al., 1982).

The mole fraction of equivalent methane ($X_{CH_4} = CO_2/(CO_2+CH_4)$) was calculated for other type 1 and 2 inclusions from the melting temperature of carbonic ice and the homogenization temperatures of carbonic liquid or vapour using the method of Burruss (1981) as modified by Heyen et al. (1982), assuming that the carbonic ice melting temperature depression is due solely to the presence of CH₄. For inclusions in which the carbonic phase homogenized at temperatures lower than clathrate decomposition, moderate re-cooling was carried out to establish Th_{CO₂} in the absence of clathrate. This may lead to an underestimation of X_{CH₄} (cf. Seitz et al., 1987), but for the clathrate decomposition temperatures at Pilok, CH₄ is probably not preferentially partitioned to significant degree (cf. Seitz and Pasteris,

Figure 3. Carbonic phase homogenization versus carbonic ice melting temperatures of type 1 and 2 aqueous-carbonic fluid inclusions. Open squares represent inclusions that are hosted by quartz, solid circles by cassiterite and stars by beryl.



1990). Figure 4 shows the frequency distribution of the X_{CH_4} calculated for type 1 and 2 inclusions in quartz, cassiterite and beryl. The inclusions in quartz generally have low X_{CH_4} compared to those in cassiterite and beryl. The range of X_{CH_4} in vapour-rich (type 1) is similar to that of liquid-rich (type 2) inclusions. The frequency distribution of high X_{CH_4} is greater in type 1 than in type 2 inclusions suggesting that CH_4 may have been partitioned into the vapour. However, if the X_{CH_4} values from individual samples are examined (Table 2) there does not seem to be any systematic partitioning of CH_4 into vapour. For example, in sample PP-2-4b (quartz) both type 1 and 2 inclusions have average X_{CH_4} values of 0.03 and have a similar range of values. The same is true in sample DP-1a (cassiterite), but the average X_{CH_4} values of type 1 and type 2 inclusions are both 0.10.

The composition of the aqueous portion of type 1 and 2 inclusions can be estimated from clathrate decomposition temperatures (in the presence of carbonic liquid and vapour, and water) and the equation of Bozzo *et al.* (1973). This equation is only valid in the system H_2O-CO_2-NaCl , but as indicated in Figure 5, there is a linear relationship between clathrate and carbonic melting temperatures. Salinities were therefore calculated by extrapolating the clathrate decomposition temperatures along the linear trend to $-56.6^\circ C$. This yields clathrate decomposition temperatures for inclusions in quartz that range from 9.7° to $7.2^\circ C$, which correspond to salinities of 0.6 to 5.4 wt % NaCl eq. Clathrate decomposition in L-V* inclusions is potentially not invariant (Th_{CO_2} was not observed) but the data for

Table 2

X_{CH_4} of Aqueous-Carbonic Inclusions

Sample	Host	Mineralization	Type	n	X_{CH_4} ave.	X_{CH_4} range
PP-2-4b	qtz	qtz-wolf vein	L-L-V	8	0.03	0.00-0.04
PP-2-4b	qtz	qtz-wolf vein	V-L-L	6	0.03	0.02-0.03
PP-2-5b	qtz	qtz-wolf-py vein	L-L-V	1	0.12	-
PP-2-5b	qtz	qtz-wolf-py vein	V-L-L	4	0.07	0.03-0.15
PP-2-7b	qtz	qtz-cas-ber vein	L-L-V	6	0.04	0.00-0.12
PP-2-7b	qtz	qtz-cas-ber vein	V-L-L	7	0.08	0.02-0.15
PP-2-14	qtz	qtz-asp vein	L-L-V	2	0.12	0.07-0.16
PP-2-14	qtz	qtz-asp vein	V-L-L	9	0.05	0.03-0.07
PP-2-5a	cas	qtz-mu-cas vein	L-L-V	15	0.17	0.10-0.32
DP-1a	cas	qtz-cas vein	L-L-V	6	0.10	0.03-0.13
DP-1a	cas	qtz-cas vein	V-L-L	8	0.10	0.06-0.14
PP-2-7b	ber	qtz-cas-ber vein	L-L-V	4	0.16	0.09-0.26
PP-2-7b	ber	qtz-cas-ber vein	V-L-L	12	0.13	0.06-0.25

Abbreviations are ber=beryl, cas=cassiterite, mu=muscovite, qtz=quartz and wolf=wolframite. V-L-L are type 1 and L-L-V are type 2 inclusions.

Figure 4. Frequency distribution of X_{CH_4} in type 1 (V-L-L) and type 2 (L-L-V) inclusions in quartz, cassiterite (Cas) and beryl.

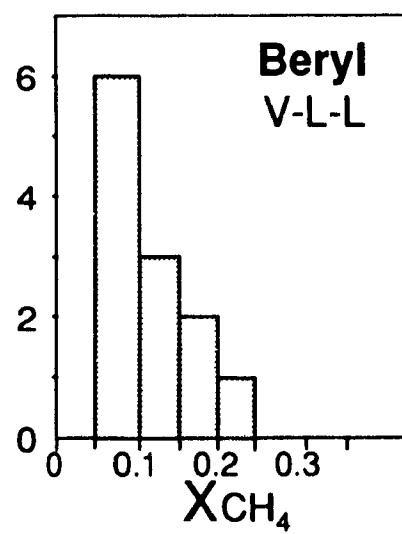
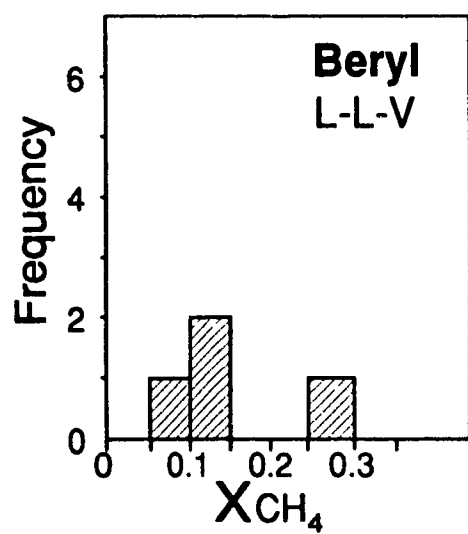
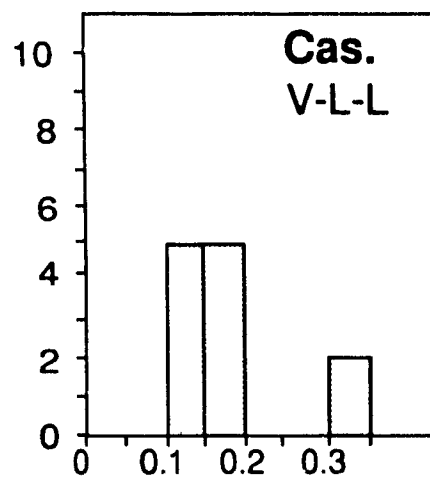
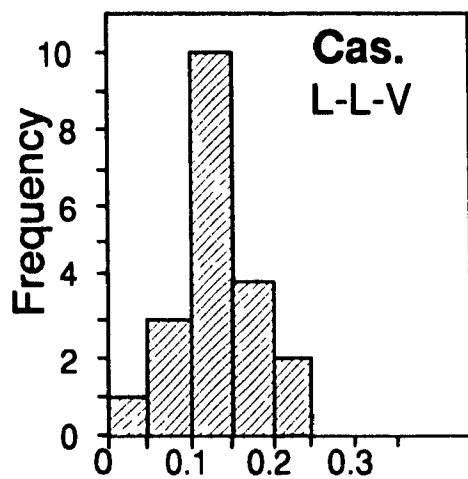
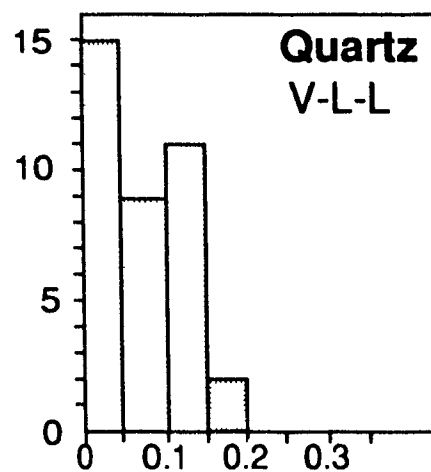
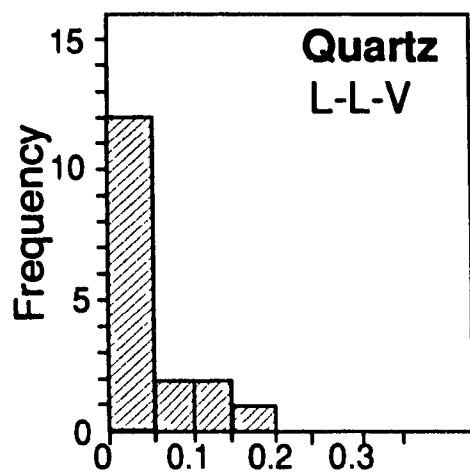
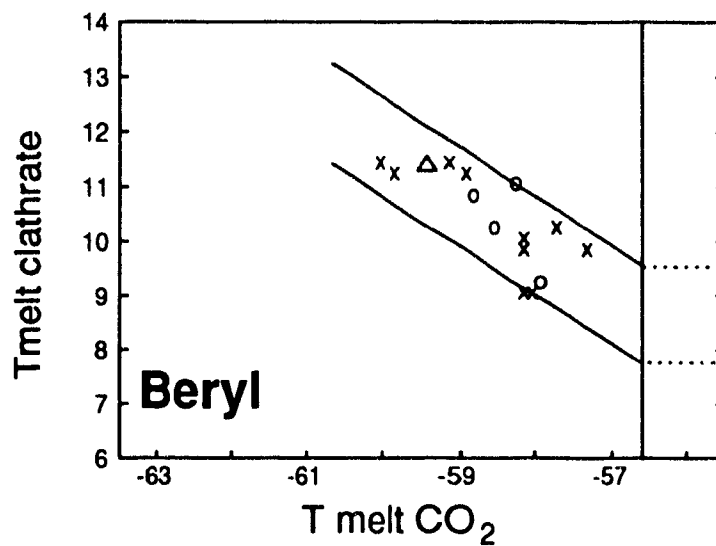
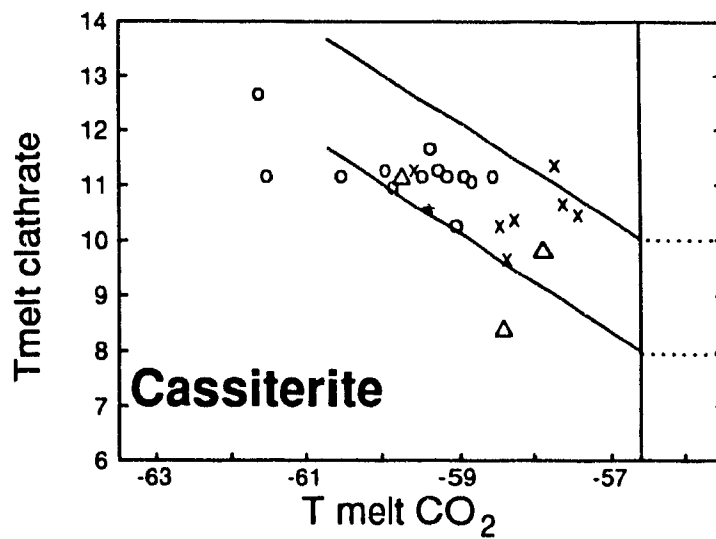
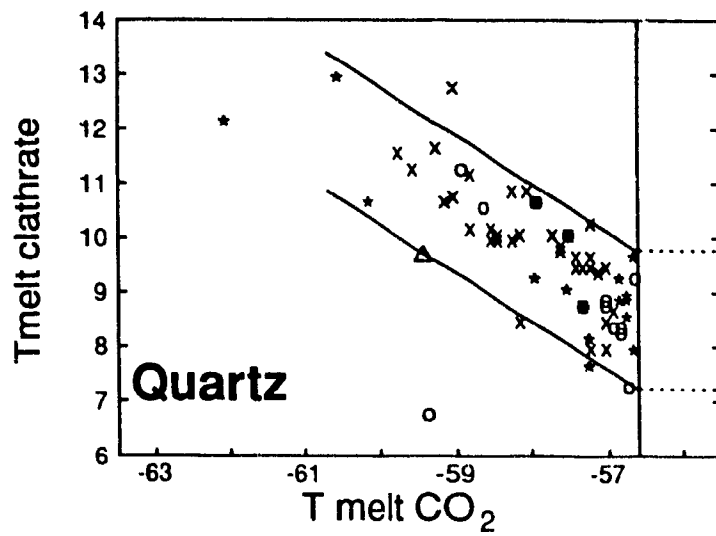


Figure 5. Clathrate decomposition versus carbonic ice melting temperatures for inclusions hosted by quartz, cassiterite and beryl. Crosses represent V-L-L (type 1), triangles critical compositions, circles L-L-V (type 2) and stars L-V* (type 3) inclusions. The solid lines represent the projection of the clathrate melting temperature to that which corresponds with a carbonic melting temperature of -56.6°C .



L-V* inclusions lie within the linear trends and were therefore included in Figure 5. Well defined negative correlations between the clathrate and carbonic melting temperatures are apparent for inclusions in quartz and beryl, suggesting that the extrapolation probably provides a reasonable estimate of salinity. However, the data for cassiterite show a smaller variation of the temperature of clathrate melting with that of carbonic ice melting, and the salinity estimates are less certain (Fig. 5). The Nong Sua deposit is located in the same granite province and metasedimentary sequence. The aqueous-carbonic inclusions in cassiterite at Nong Sua contain N_2 in addition to CO_2 and CH_4 (Linnen and Williams-Jones, Chapter 4) and the presence of N_2 may explain why the distribution of clathrate and carbonic melting temperatures in cassiterite at Pilok differs from that in quartz and beryl.

The X_{CO_2} ($CO_{2eq}/CO_{2eq}+H_2O$) of type 1 and 2 inclusions were estimated using cross sectional areas of fluid inclusions and vapour bubbles, accurately measured with the aid of a drawing tube. Elliptical outlines of the carbonic phase within aqueous-carbonic inclusions and elliptical outlines of carbonic vapour within carbonic liquid indicate that the inclusions are relatively flat. The X_{CO_2} (a maximum value) can therefore be estimated using the relative areas of the aqueous and carbonic phases. The total homogenization temperatures versus X_{CO_2} are plotted in Figures 6 and 7. The data have been subdivided on the basis of whether homogenization was to liquid (H_2O), vapour (CO_2), or occurred critically. The homogenization temperatures in quartz range from 275° to 385°C, and one

inclusion with an X_{CO_2} of 0.18 homogenized critically at 355°C (Fig. 6). The distribution of homogenization temperatures and X_{CO_2} values in Figure 6 lie close to the solvus in the system $\text{H}_2\text{O}-\text{CO}_2$ (350 bars), suggesting that immiscible fluids may have been trapped. The microthermometry indicates at least a four component system, $\text{H}_2\text{O}-\text{CO}_2-\text{CH}_4-\text{NaCl}$, thus it may not be reasonable to model these fluid inclusions in the $\text{H}_2\text{O}-\text{CO}_2$ system. It is important to note that the 6 wt % solvi (Fig. 6) are pseudobinary projections and that true boiling fluids will not lie precisely on the curves as depicted. The $\text{NaCl}:\text{H}_2\text{O}$ for these solvi is constant, which is not the case for coexisting liquid and vapour (cf. Bowers and Helgeson, 1983; Trommsdorff and Skippen, 1986). Unfortunately the partitioning of NaCl (and CH_4) between liquid and vapour in aqueous-carbonic fluids as a function of pressure, temperature and composition is poorly known, but the true solvi (for low salinity) are probably similar to those on Figure 6. The data for cassiterite and beryl (Fig. 7) do not appear to delineate a solvus (there is no decrease in homogenization temperatures at lower X_{CO_2}), and thus do not reflect fluid immiscibility.

AQUEOUS INCLUSIONS

The salinities of type 3, 4, 5 and 6 aqueous inclusions were calculated using ice melting temperatures and the equations of Oakes *et al.* (1990), assuming that the freezing point depression is caused entirely by NaCl. Type 3 inclusions range

Figure 6. Homogenization temperature versus X_{CO_2} for aqueous-carbonic inclusions hosted by quartz. Squares represent V-L-L (type 1), triangles critical compositions and circles L-L-V (type 2) inclusions. The solvi for 0 and 6 wt % NaCl in the system $\text{H}_2\text{O}-\text{CO}_2-\text{NaCl}$ are from Bowers and Helgeson (1983).

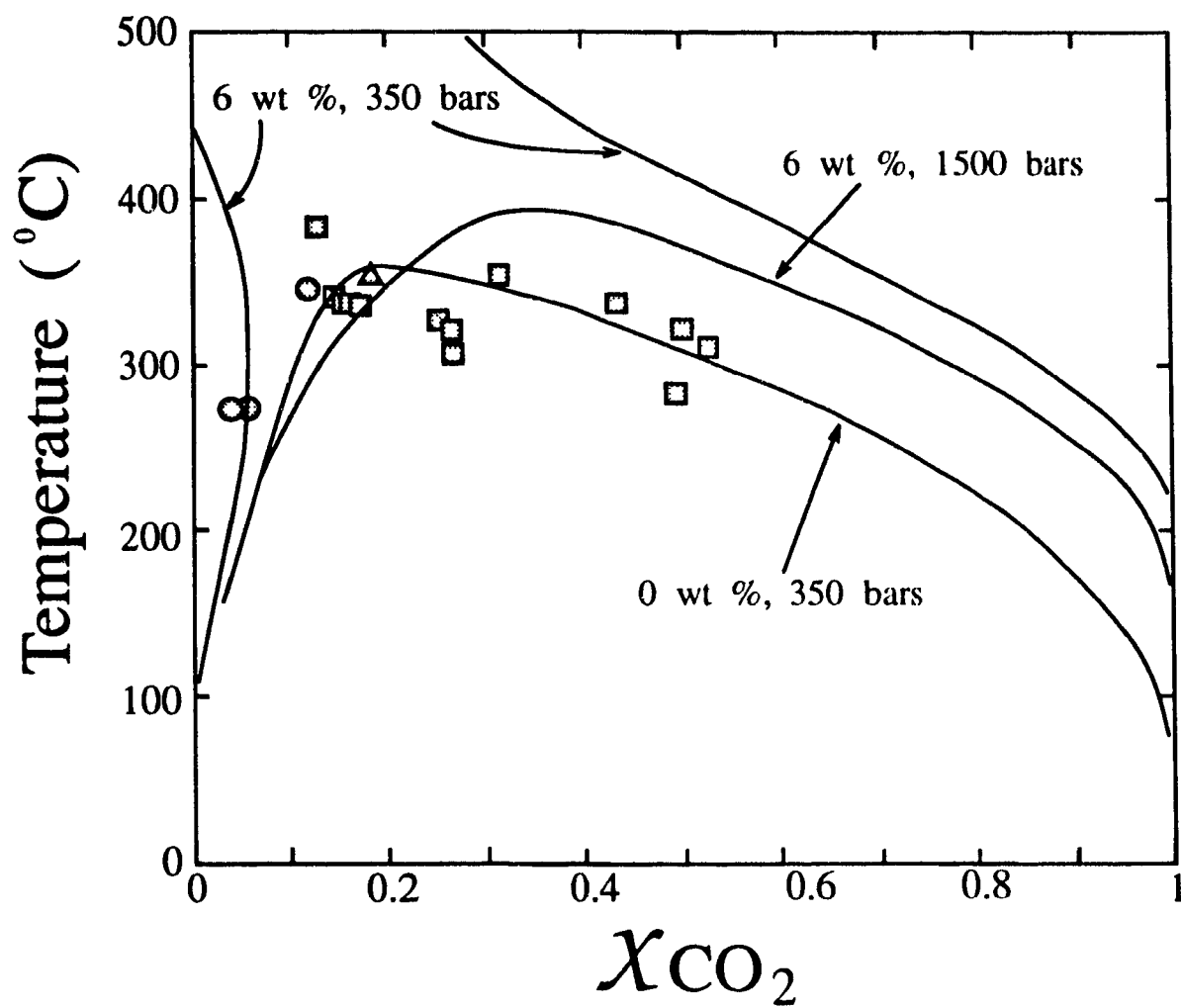
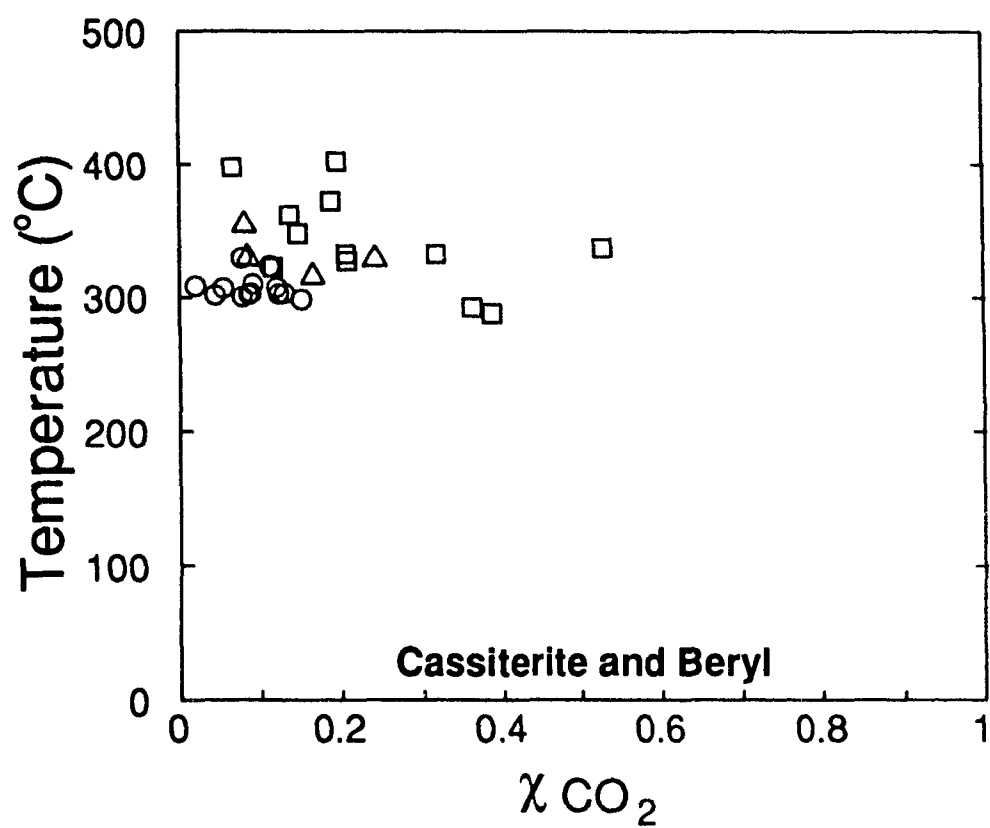


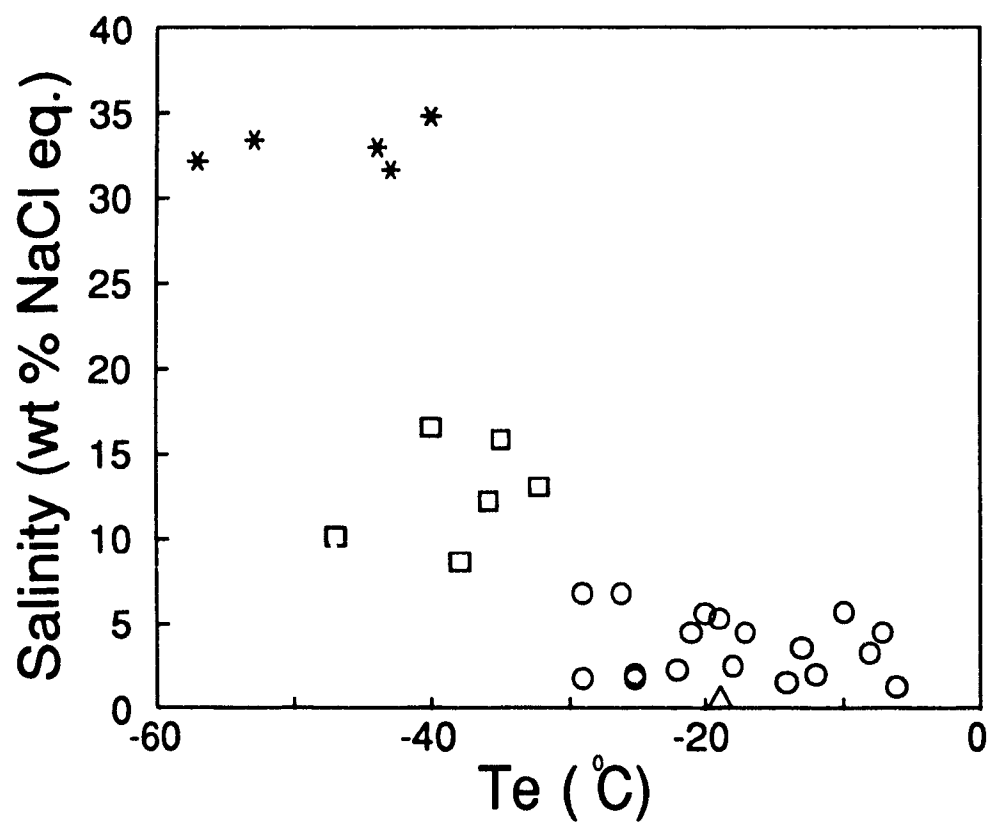
Figure 7. Homogenization temperature versus X_{CO_2} for aqueous-carbonic inclusions hosted by cassiterite and beryl. Squares represent V-L-L (type 1), triangles critical compositions and circles L-L-V (type 2) inclusions.



in composition from approximately 1 to 6 wt % NaCl eq., and type 5 inclusions range from 4 to 18 wt % NaCl eq. (Fig. 8). The salinities of type 4 (L) inclusions are <1.0 wt % NaCl eq., whereas the salinities of halite-bearing (type 6) inclusions, calculated from the equation of Bischoff (1991), range from 32 to 35 wt % NaCl eq. (Fig. 8). Some inclusions, particularly type 6, exhibit eutectic temperatures that are significantly lower than -21°C , indicating the presence of salts such as CaCl_2 , MgCl_2 or FeCl_2 (cf. Crawford, 1981).

Type 6 inclusions typically freeze at temperatures of -70° to -80°C . The first ice melting occurs at temperatures lower than -50°C and the halite cube reacts with the liquid to form hydrohalite. The final ice melting occurs at temperatures as low as -30°C if the halite cube has not reacted completely with the fluid. After final ice melting, the inclusions were re-frozen and melted several times. The halite reacted completely with liquid to form hydrate and the freezing, eutectic, and final ice melting temperatures stabilized at higher values, of approximately -52° , -40°C and -25°C , respectively. We suggest that the NaCl to bivalent cation ratio increased with cycling (reaction of halite) and hence raised the temperatures of phase transitions. This cycling has the opposite effect of that in L-V inclusions (cf. Haynes, 1985). The final low temperature phase transition observed in type 6 inclusions is hydrate decomposition at approximately $+18^{\circ}\text{C}$. Re-nucleation of halite coincided with hydrate decomposition, suggesting that the latter phase is hydrohalite (the inclusions are typically only 10 μm across and the birefringence

Figure 8. Salinity (wt % NaCl eq.) versus temperature of first ice melting (T_e) in type 3 (circles), type 4 (triangles), type 5 (squares) and type 6 (stars) inclusions.

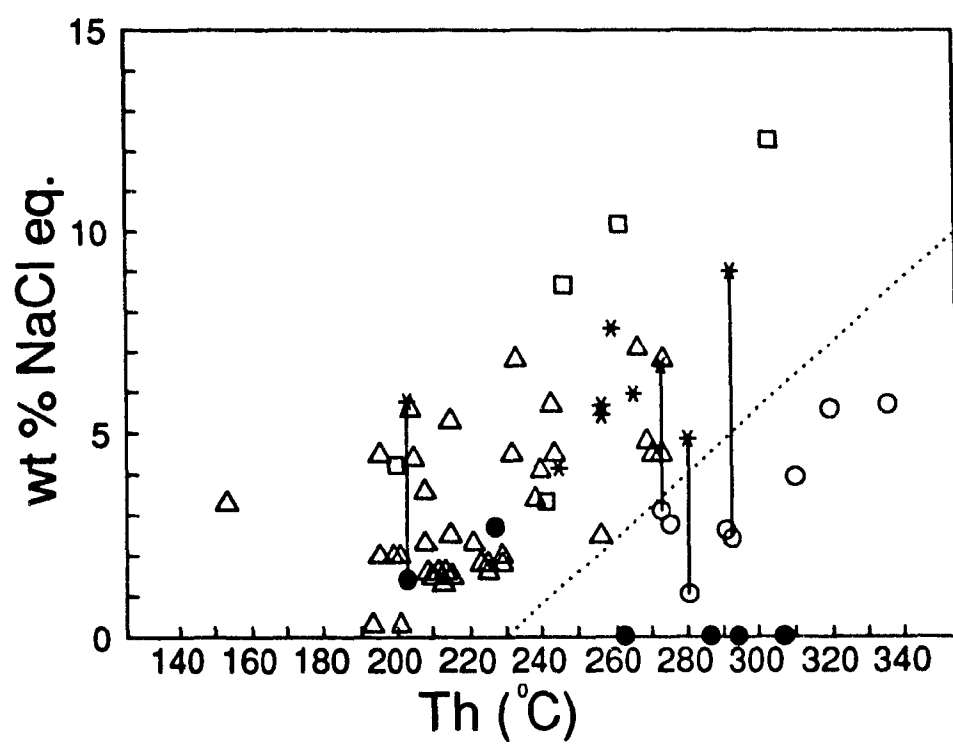


of the hydrate could not be established). The temperature at which hydrohalite disappeared was difficult to reproduce which therefore may reflect metastable decomposition (cf. Roedder, 1984).

Figure 9 shows the relationships between homogenization temperature and salinity. For several of the L-V* inclusions salinities were estimated from ice and clathrate melting temperatures (the tie-lines on Fig. 9). Owing to the opposing affects of NaCl and CH₄, the salinities obtained from ice melting and clathrate melting temperatures are maximum and minimum values, respectively (cf. Collins, 1979). Where carbonic ice and clathrate melting temperatures were measured in the same inclusion, salinities were estimated by the extrapolation method discussed above (the open circles on Fig. 9). Where only clathrate melting temperatures were measured, the salinities were estimated using the equation of Bozzo *et al.* (1973), i.e., assuming pure CO₂ (solid circles on Fig. 9), and inclusions that contain clathrates that melt at temperatures >10°C were assigned a salinity of 0.0 wt. % NaCl eq. Most type 3 and 5 inclusions define a mixing trend from approximately 13 wt % NaCl eq. and Th of 295°C to approximately 2 wt % NaCl eq. and Th of 200°C. Data from single-phase (L) and halite-bearing inclusions are not included on Figure 9. The liquid-only inclusions have salinities of <1.0 wt % NaCl eq. The low salinity of type 5 inclusions and the occurrence of these inclusions along secondary planes with low salinity type 3 inclusions suggests that the mixing trend reflects cooling and dilution. The vapour in type 6 inclusions homogenizes

Figure 9. Salinity versus homogenization temperature for aqueous inclusions.

Triangles represent type 3 (L-V) inclusions, squares type 5 (L-V-S), stars type 3 (L-V*) inclusions (salinity from ice melting temperatures). Open circles represent type 3 (L-V*) inclusions (salinity extrapolated from clathrate melting temperatures) and solid circles type 3 (L-V*) inclusions (salinity from clathrate melting temperatures assuming pure CO₂). The tie lines join data from single inclusions and the dotted line separates data obtained from clathrate and ice melting temperatures.



between 151° and 197°C and the inclusions generally homogenize by halite dissolution at 190° to 256°C.

If the ice melting temperatures of the L-V* inclusions are used to estimate salinity, these inclusions define part of the mixing trend. However, if clathrate melting temperatures are used to estimate salinity, a parallel trend at lower salinity and higher temperature is apparent (Fig. 9). It is not clear whether the ice melting and/or clathrate melting temperatures give unreliable estimates of salinity or whether the higher temperature L-V* inclusions define a separate mixing trend. It is significant that the L-V* inclusions (all but 4 data points are from quartz) typically homogenize at temperatures <300°C. Although the X_{CO_2} cannot be accurately estimated, the X_{CO_2} is low (the lack of visible carbonic homogenization implies low densities), and the corresponding low homogenization temperatures suggest T- X_{CO_2} values that are close to the limbs of the solvi (<300°C) on Figure 6.

STABLE ISOTOPES

The $\delta^{18}\text{O}$ compositions of quartz, muscovite and cassiterite from barren pegmatite, wolframite-bearing pegmatite, veins and greisens are summarized in Table 3. Two barren pegmatite samples contain quartz with $\delta^{18}\text{O}$ values of 10.9 and 10.8 per mil that coexist with muscovite having $\delta^{18}\text{O}$ values of 8.2 and 8.9 per mil. Temperatures calculated from these data using the equations of O'Neil and

Table 3 Stable Isotope Data

Sample	Description	$^{18}\delta\text{O}_{\text{qtz}}$	$^{18}\delta\text{O}_{\text{mu}}$	$^{18}\delta\text{O}_{\text{cas}}$	$\delta\text{D}_{\text{mu}}$
DP-11	pegmatite	10.9	8.2		-100
DP-9b	pegmatite	10.8	8.9		-114
PP-2-4a	pegmatite with wolframite	9.4	9.3		
PP-1-4c	pegmatite with wolframite	9.1	7.1		
PP-5-12b	vein/greisen	10.4	8.3	3.7	-134
PP-3-2a	vein/greisen	11.7	8.5	3.4	-114
PP-2-5a	vein/greisen	10.2	6.1	3.5	
DP-1b	greisen	12.4	9.0		-108
PP-5-31e	greisen	8.9	6.8		

Sample	T ¹	T ²	T ³	$^{18}\delta\text{O}_{\text{fluid}}^1$	$^{18}\delta\text{O}_{\text{fluid}}^2$	$^{18}\delta\text{O}_{\text{fluid}}^3$	$^{18}\delta\text{O}_{\text{fluid}}^4$	$\delta\text{D}_{\text{fluid}}^5$
DP-11	400			+6.8			+6.8	-70
DP-9b	569			+9.4			+6.7	-102
PP-2-4a	-			-			+3.2	
PP-1-4c	541			+7.4			+5.0	
PP-5-12b	515	405	366	+8.4	+6.5	+5.5	+6.3	-118
PP-3-2a	334	336	337	+5.9	+6.0	+6.0	+7.6	-73
PP-2-5a	253	405	547	+1.4	+6.3	+8.6	+6.1	
DP-1b	313			+6.0				-63
PP-5-31e	515			+6.9			+4.8	

Abbreviations are qtz - quartz, mu - muscovite and cas - cassiterite

- ¹ Temperature and fluid $\delta^{18}\text{O}$ composition calculated from quartz-muscovite fractionation and the equations of Clayton et al. (1972) and O'Neil and Taylor (1969).
- ² Temperature calculated from quartz-cassiterite fractionation and the equation of Alderton (1989) and fluid $\delta^{18}\text{O}$ composition from quartz-water fractionation and the equation of Clayton et al. (1972).
- ³ Temperature and fluid $\delta^{18}\text{O}$ composition calculated from muscovite-cassiterite fractionation and the equations of Clayton et al. (1972), O'Neil and Taylor (1969) and Alderton (1989).
- ⁴ Fluid $\delta^{18}\text{O}$ composition calculated from quartz-water fractionation and the equation of Clayton et al. (1972) at 400°C.
- ⁵ Fluid δD composition calculated at the temperature of ¹, using the equation of Suzuoki and Epstein (1976).

The isotopic ratios were provided by K. Kyser at the University of Saskatchewan on a Mat Finnigan 251 gas mass spectrometer, using NBS-28 quartz and NBS-30 biotite standards. The error on the $\delta^{18}\text{O}$ determinations are ± 0.1 per mil, resulting in an error of $\pm 70^\circ\text{C}$ ($\pm 20^\circ\text{C}$ at $T < 400^\circ\text{C}$) and ± 0.8 per mil $^{18}\delta\text{O}_{\text{fluid}}$. The error on δD muscovite determinations are ± 1 per mil, resulting in an error of ± 6 per mil $\delta\text{D}_{\text{fluid}}$ (including temperature error estimated from oxygen isotope fractionation).

Taylor (1969) and Clayton et al. (1972) are 400° and 569°C (Table 3) and, at these temperatures, the coexisting fluids are calculated to have had $\delta^{18}\text{O}$ values of +6.8 and +6.7 per mil. These temperatures and compositions are similar to those obtained for magmatic muscovite at the Nong Sua deposit which was interpreted to have re-equilibrated with fluids at sub-solidus conditions (Linnen and Williams-Jones, Chapter 4). The δD values of pegmatite-hosted muscovite at Pilok are -100 and -114 per mil, and the coexisting fluid is calculated to have had a δD of -70 and -102 per mil (Table 3). These values are lower than those estimated for Nong Sua and are lower than those normally considered to be magmatic (Taylor, 1979). It is therefore probable that the muscovite in the barren pegmatite pods at Pilok formed at magmatic or early sub-solidus conditions but re-equilibrated isotopically with later fluids.

The oxygen isotopic compositions of quartz and muscovite were determined from two pegmatites that contain wolframite $\pm$ cassiterite mineralization (Table 3). In one of these pegmatites, the quartz and muscovite have $\delta^{18}\text{O}$ compositions of 9.1 and 7.1 per mil, respectively. The temperature estimated for this pair is 541°C and the $\delta^{18}\text{O}$ of the coexisting fluid is estimated to have been +7.4 per mil (Table 3). These estimates are consistent with field relationships, that suggest a late magmatic to early hydrothermal origin for the mineralized pegmatite. However, as was the case for the barren pegmatite, the muscovite and quartz in mineralized pegmatite may have undergone re-equilibration. The other quartz-muscovite pair has $\delta^{18}\text{O}$

values of 9.4 and 9.3 per mil, respectively. A fractionation of 0.1 per mil between quartz and muscovite requires an unreasonably high temperature, suggesting that these minerals were not in equilibrium in this sample.

The $\delta^{18}\text{O}$ compositions of quartz and muscovite from three veins with greisen envelopes and two pervasively altered leucocratic granite samples are more variable. The $\delta^{18}\text{O}$ values for quartz range from 8.9 to 11.7 per mil and those for muscovite range from 6.1 to 9.0 per mil (Table 3). The isotopic compositions of three cassiterite samples are more restricted; the $\delta^{18}\text{O}$ values range from 3.4 to 3.7 per mil. The temperatures calculated from oxygen isotopic fractionation between quartz and muscovite range from 253° to 515°C, with an average of 386°C (Table 3). At these temperatures the coexisting fluid is estimated to have had a $\delta^{18}\text{O}$ composition of +1.4 to +8.4 per mil. Similar temperatures, 336° to 405°C, and fluid compositions, +6.0 to +6.5 per mil (Table 3), were obtained from oxygen isotopic fractionation between quartz and cassiterite and the equation of Alderton (1989) (comparable temperatures are obtained using the equation of Zheng, 1991). Isotopic equilibrium can be tested by using three coexisting minerals and by comparing the fractionation of oxygen between the different pairs (O'Neil, 1986). For single samples the temperatures calculated for quartz-muscovite, quartz-cassiterite and muscovite-cassiterite pairs (Table 3) are 366° to 515°C (PP-5-12b), 334° to 337°C (PP-3-2a) and 253° to 547°C (PP-2-5a), respectively. We consider that the first two triplets are close to equilibrium, and that cassiterite mineralization

probably occurred at temperatures of approximately 450° to 300°C. The δD values of three hydrothermal muscovite samples range from -108 to -134 per mil (Table 3). At the temperatures determined by quartz-muscovite oxygen isotope fractionation, the coexisting fluid is estimated to have had δD compositions of -63 to -118 per mil (Table 3). Similar isotopic ratios for hydrothermal muscovite at Pilok, $\delta^{18}O$ of 8.3 ± 0.8 and δD of -119 ± 8 per mil, are also reported by Lehmann *et al.* (1991).

The calculated isotopic compositions of the hydrothermal fluids at Pilok are similar to those of many porphyry systems that have been interpreted to reflect mixing between magmatic and meteoric waters (Taylor, 1979). We consider that the isotopic ratios at Pilok may reflect magmatic-meteoric mixing. Similar conclusions were also reached by Lehmann *et al.* (1991). The paleolatitude of Thailand in the late Cretaceous is interpreted to have been approximately at its current position (Bunopas *et al.*, 1989). Therefore if the low $\delta^{18}O$ and δD values reflect meteoric fluids, a relatively high elevation is implied.

The lack of consistent variation of the isotopic compositions of quartz and muscovite suggest that the fluid evolution may be more complex than simple magmatic-meteoric mixing, i.e., quartz and muscovite from greisen do not have consistently lower $\delta^{18}O$ values. There are several possible reasons for the inconsistencies: a) Isotopic equilibrium may not have been achieved because the minerals did not precipitate from the same fluid at the same temperature or the

different minerals partially re-equilibrated to different degrees; b) If boiling occurred in the system H_2O , the water would have become isotopically heavier with vapour loss or, in the system $\text{H}_2\text{O}-\text{CO}_2$, the water would have become isotopically lighter with vapour loss (Bowers, 1991); c) If more than two fluids mixed, i.e., magmatic, meteoric, metamorphic and formational waters (including fluids with different $\text{CO}_2/\text{H}_2\text{O}$ ratios), the resulting isotopic compositions of minerals precipitating from combinations of the fluids could have varied irregularly, and d) If metamorphic fluids or fluids that had interacted with graphitic or organic-rich pelite were involved, a wide range of δD compositions are possible (Sheppard, 1986). Potentially any or all of these processes may have affected the isotopic compositions of minerals at Pilok, but this cannot be evaluated from the current data.

MUSCOVITE CHEMISTRY

Muscovite is present at several of the paragenetic stages (Fig. 2). A chemical survey, using electron microprobe analyses, was therefore conducted to confirm the paragenetic divisions and to establish whether compositional changes reflect physicochemical differences in the formation of the various muscovite types. The compositions of muscovite, in terms of Ti-Mg-Na and $\text{TiO}_2\text{-FeO-MgO}$, can be used to discriminate between magmatic and hydrothermal origins (Miller *et al.*, 1981; Monier *et al.*, 1984). The muscovite at Pilok is relatively Fe-rich and there

is an overall trend of titanium depletion and iron enrichment from barren pegmatite to mineralized pegmatite to greisen (Fig. 10a). This trend suggests a magmatic origin for the muscovite hosted by barren pegmatite and a hydrothermal origin for the latter two types. A normal magmatic to hydrothermal trend typically involves decreasing Ti and Na with increasing Mg (Miller *et al.* 1981). This is not the case at Pilok (Fig. 10b), but the Mg contents are very low. This diagram may therefore not be useful for discriminating magmatic from non-magmatic muscovite in Fe-rich systems.

Parameters that more clearly distinguish the three muscovite types are the fluorine contents and the paragonite mole fraction (X_{par}). Figure 11 shows that muscovite from barren pegmatites is relatively Na-rich and F-poor, whereas muscovite from mineralized pegmatites is Na-poor and F-rich, and that from greisens is Na-poor and F-poor. The different muscovite types can be also distinguished by their total octahedral M^{+2} contents (Fe+Mg+Mn). Figure 12 shows that muscovite from barren pegmatites has total M^{+2} contents of <0.5 cations, which is consistent with a magmatic origin. The total M^{+2} contents of muscovite from mineralized pegmatite and greisen are both higher, although there is some overlap between greisen and barren pegmatite. The cluster of analyses from mineralized pegmatite that have total M^{+2} contents of >1.0 cations correspond with the highest fluorine contents. The high M^{+2} and fluorine contents distinguish muscovite of mineralized pegmatite from the other types. This strongly supports the paragenesis

Figure 10. Compositions of muscovite in terms of: a) wt % TiO_2 , FeO and MgO , and; b) cation % Ti , Mg and Na . The cations are calculated based on a stoichiometry of 24 (O,OH,F) and total Fe as Fe^{+2} . Solid circles represent muscovite from barren pegmatite, stars from wolframite-bearing pegmatite and open circles from greisen.

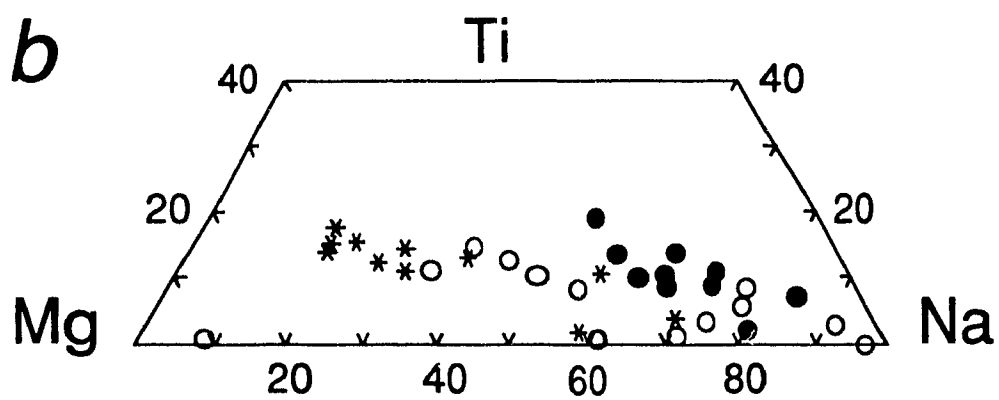
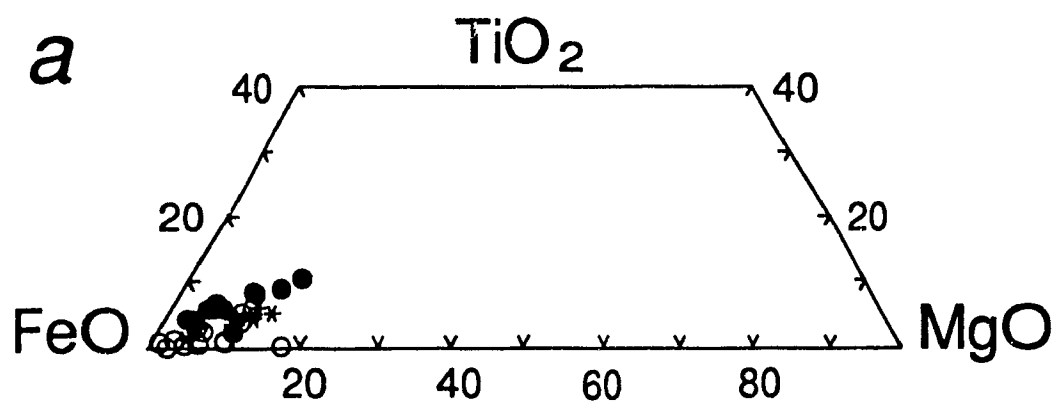
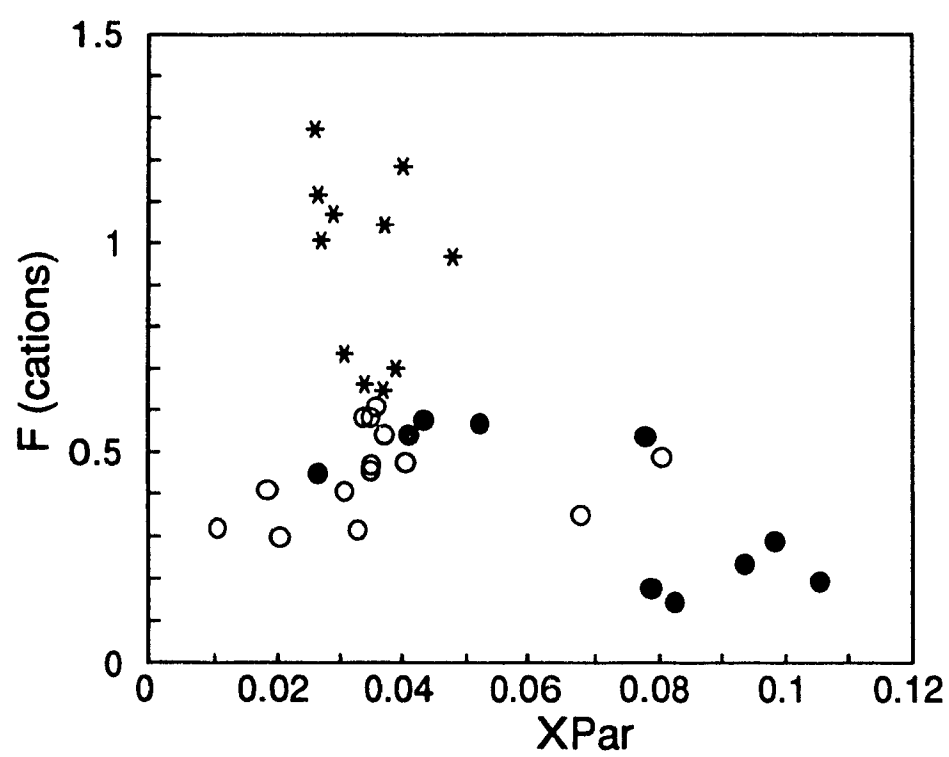


Figure 11. Fluorine versus mole fraction of paragonite in muscovite. The cations are calculated based on a stoichiometry of 24 (O,OH,F) and total Fe as Fe^{+2} . Solid circles represent muscovite from barren pegmatite, stars from wolframite-bearing pegmatite and open circles from greisen.



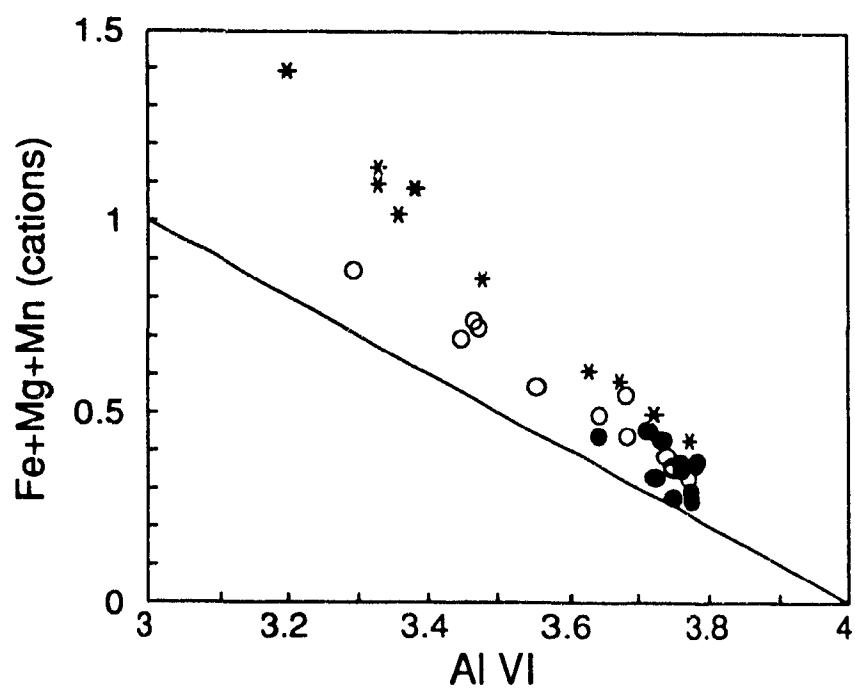
of muscovite established from field relationships, and therefore it follows that the discrimination of a separate pegmatite stage of mineralization is justified.

Higher phengite substitution (greater M^{+2} and Si^{IV}) commonly corresponds to a lower temperature of formation (Monier and Robert, 1986a), and thus it is somewhat surprising that muscovite from mineralized pegmatites contains higher phengite mole fractions than greisen muscovite. A possible explanation is that the muscovites are not Li-free. At constant temperature the M^{+2} total can increase with the addition of Li (zinnwaldite substitution; Monier and Robert, 1986b). The Li contents of whole-rock or mineral samples from Pilok have not been determined and the western Thai granite belt contains both Li-rich and Li-poor intrusions (Charusiri, 1989). The muscovite chemistry cannot therefore be further interpreted without Li analyses. The composition of muscovite from mineralized pegmatite is not unique since Fe-rich muscovite from Portuguese W-Sn deposits (Neiva, 1987) and the East Kemptville, Nova Scotia Sn deposit (Kontak, 1991) have been interpreted to have a magmatic origin.

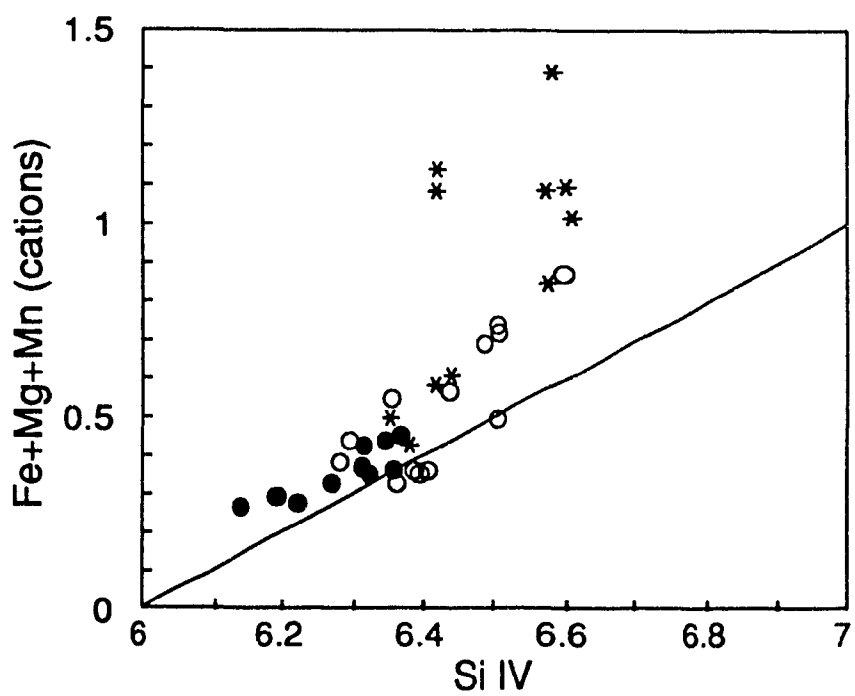
Differences in fluorine contents are also apparent from fluorine intercept values (cf. Munoz, 1984; note that low intercept values reflect high fluorine activities). The intercept values for barren pegmatite muscovite range from 0.77 to 1.35 (ave. 1.00, $n=10$), those from greisen range from 0.79 to 1.11 (ave. 0.89, $n=14$) and those from mineralized pegmatite range from 0.54 to 0.72 (ave. 0.62, $n=11$). The intercept values from the mineralized pegmatite muscovites are

Figure 12. Compositions of muscovite in terms of: a) total octahedral cations versus Al^{VI} , and; b) total octahedral cations versus Si^{IV} . The cations are calculated based on a stoichiometry of 24 (O,OH,F) and total Fe as Fe^{+2} . Solid circles represent muscovite from barren pegmatite, stars from wolframite-bearing pegmatite and open circles from greisen.

a



b



comparable to the muscovites from the Henderson molybdenum deposit that are interpreted to reflect high HF/H₂O fugacity ratios (Gunow et al. 1980).

GENESIS OF MINERALIZATION

The earliest mineralization event in the Pilok area was the development of pegmatite-hosted wolframite ± cassiterite. Intergrowths between wolframite and Fe-rich muscovite suggest that this event was post-magmatic and, in spite of the lack of fluorite and topaz, the fluorine activity was relatively high. The temperature and pressure of this event are unknown, but since the mineralization is contained within pegmatite pods, an early hydrothermal origin is likely.

Most wolframite occurs in quartz ± K-feldspar veins that post-date pegmatite formation and a barren tourmaline alteration event. Restricted potassic (K-feldspar) alteration is associated with wolframite mineralization. The fluid inclusion data from quartz in quartz-wolframite veins are similar to those from quartz in the main stage of cassiterite deposition. This could be the result of both wolframite and cassiterite having precipitated at the same temperature, the inclusions in the wolframite veins being secondary or, quartz having been deposited at a different time than wolframite and cassiterite. Without an independent geothermometer for potassic alteration and wolframite deposition, we cannot establish the pressure, temperature and composition of the W-bearing fluids.

The main stage of cassiterite deposition was associated with greisen

alteration that overprinted potassic alteration and wolframite mineralization. The replacement of K-feldspar by muscovite indicates that the Sn ore-bearing fluids were at a lower pH and/or temperature than the preceding wolframite mineralization (Hemley and Jones, 1964). Sulphide minerals, in particular chalcopyrite and sphalerite, were also deposited during this event.

Aqueous-carbonic fluid inclusions are the dominant type in cassiterite and probably represent the fluid responsible for tin mineralization. There are several possible explanations for the range of compositions in vapour-rich and liquid-rich aqueous-carbonic inclusions (type 1 and 2); the fluids were immiscible (boiling); there was simple mixing between CO₂-rich and H₂O-rich fluids; or there was complex mixing that involved at least one aqueous fluid, a CO₂-rich and a H₂O-rich aqueous-carbonic fluid. If boiling occurred, the vapour would have been enriched in methane and the liquid enriched in salts (Ramboz *et al.*, 1982). There is no apparent partitioning of CH₄ in vapour or NaCl in liquid in cassiterite-hosted inclusions that supports boiling at Pilok. Similarly the T-X_{CO2} relationships for aqueous-carbonic inclusions hosted by cassiterite and beryl (Fig. 7) do not indicate boiling. We conclude therefore that cassiterite was not deposited by boiling (immiscible) fluids. However, the T-X_{CO2} relationships for aqueous-carbonic inclusions hosted by quartz may define a solvus. Thus, potentially there was some boiling, but it was not directly related to mineralization.

Type 1 and 2 inclusions hosted by cassiterite and beryl homogenize

predominantly at temperatures around 300°C, and range up to 400°C. These temperatures are close to those estimated by oxygen isotope thermometry (300° to 450°C). We interpret the 300° to 450°C temperatures to reflect the conditions of cassiterite mineralization. Since the majority of the aqueous-carbonic inclusions hosted by cassiterite homogenize at approximately 300°C, some pressure correction may be required. It is however, very difficult to estimate the pressure of mineralization. The equations of Bowers and Helgeson (1983) and the alternate equations of Brown and Lamb (1989) are valid only above 350°C. Thus it is not currently possible to construct isochores for the majority of the aqueous-carbonic inclusions. A rough estimate of the minimum pressure can be made from inclusions that homogenize critically. Critical homogenizations were observed for inclusions with X_{CO_2} compositions of 0.1 to 0.2 and, because area percentages were used to estimate compositions, these values can be considered to be maxima. Solvi for the systems $\text{H}_2\text{O}-\text{CO}_2$ and $\text{H}_2\text{O}-\text{CO}_2-\text{NaCl}$ are depicted on Figure 6. The data for quartz best fits a solvus at 0 wt % NaCl and 350 bars (critical composition of $X_{\text{CO}_2}=0.2$). At higher pressures the critical composition shifts to higher X_{CO_2} values. Microthermometric data indicate that the salinity of the aqueous-carbonic inclusions ranged up to 5.4 wt % NaCl eq. and the X_{CH_4} up to 0.32. The inclusions cannot thus be modelled in the system $\text{H}_2\text{O}-\text{CO}_2$. In the $\text{H}_2\text{O}-\text{CO}_2-6$ wt % NaCl system the critical composition at 1000 bars is at approximately $X_{\text{CO}_2}=0.25$ whereas at 500 bars, fluid compositions in the X_{CO_2} range of 0.1 to 0.3 are not stable below 400°C

(Bowers and Helgeson, 1983). In summary, we cannot accurately model the aqueous-carbonic inclusions from Pilok but suggest that, owing to the low X_{CO_2} of the critical compositions (maximum estimates), the homogenization pressures are between 350 and 1000 bars. The lithostatic pressure is broadly constrained by the igneous and metamorphic mineralogy to between 1500 and 3800 bars.

Petrographic observations indicate that type 3 (aqueous) inclusions are commonly secondary and post-date type 2 (aqueous-carbonic) inclusions. The fact that some apparently primary aqueous inclusions (L-V*) exhibit double freeze phenomena suggests some mixing between aqueous-carbonic and aqueous fluids. There was also mixing between low salinity-low homogenization temperature and higher salinity-higher homogenization temperature fluids. The occurrence of low salinity single-phase aqueous inclusions (type 4) along planes of type 3 inclusions with low salinities and homogenization temperatures indicates that the late fluid evolution was dominated by cooling and dilution. These low salinities, combined with the stable isotope data, indicate a late influx of meteoric water. Using the lowest homogenization temperature, approximately 160°C, and assuming a geothermal gradient of 40°C/km (the granite was probably still cooling), a depth of approximately 4 km is estimated. This corresponds to lithostatic and hydrostatic pressures of approximately 1000 and 350 bars, respectively. However, type 4 (L) inclusions may be related to the post-mineralization composite quartz vein which may have been emplaced after significant uplift had occurred.

COMPARISON TO OTHER DEPOSITS

It is constructive to compare the Pilok deposits with other W-Sn (or Sn-W) base metal deposits. Two areas which have received extensive study are Panasqueira, Portugal and the deposits associated with the Cornubian batholith, Cornwall, England. The deposits at Panasqueira are associated with a two-mica granite (Kelly and Rye, 1979) and some tourmaline-muscovite granite occur regionally (Oosterom *et al.* 1984). Most of the Sn-W mineralization is associated with greisen alteration and tourmaline is also an important alteration mineral (Kelly and Rye, 1979). The muscovite at Panasqueira contains >1 wt % F (Neiva, 1987), but the greisen at Panasqueira differs from that at Pilok by also containing minor topaz. Boron-rich late-stage granitic rocks occur in Cornwall, as is the case in western Thailand, and some tin mineralization in Cornwall is associated with tourmalinization (Manning, 1986). However, much of the mineralization in Cornwall is associated with later F-Li-rich topaz granites (Manning, 1986).

The hydrothermal evolution at Panasqueira and Cornwall is thought to have involved early "magmatic" (or fluids that equilibrated with granite) and late meteoric waters (Kelly and Rye, 1979; Jackson *et al.* 1989). The early fluids at Panasqueira were aqueous $\pm$ CO₂, moderately saline, and have homogenization temperatures of up to 360°C. The maximum pressure for the formation of the deposit is approximately 1500 bars, but the majority of the mineralization is

interpreted to have occurred at <200 bars (Kelly and Rye, 1979; Bussink *et al.* 1984). Later fluids were at lower temperatures, salinities, and pressures. The early fluids at Cornwall are also moderately saline (Rankin and Alderton, 1985; Bottrell and Yardley, 1988; Jackson *et al.* 1989) although some hypersaline fluids may have also been early (Rankin and Alderton, 1985). The temperature and pressure of Sn-W mineralization is estimated to have been 375° to 450°C and 250 to 400 bars (Jackson *et al.*, 1982; Shepherd *et al.*, 1985). The origin of the moderately saline fluid inclusions at Pilok is unknown, but the pattern of decreasing salinity with decreasing homogenization temperature is similar to that documented at Panasqueira and Cornwall. The low temperature and salinity fluids at Pilok, like those at Panasqueira and Cornwall, suggest a late influx of meteoric water. Boron-rich granites occur in western Thailand and Cornwall (Manning, 1986), and yet tourmaline breccia pipes are developed in Cornwall but are absent at Pilok. We suggest that the fluid evolution at Pilok was similar to that at Cornwall, but the lack of breccias indicates that the depth of formation was somewhat greater.

One of the most important similarities between Pilok and Panasqueira is the association of mineralization with aqueous-carbonic fluid inclusions. At Panasqueira these fluids are interpreted to have boiled, which is supported by CH₄-enrichment in the vapour and NaCl-enrichment in the liquid (Bussink *et al.* 1984). The carbon and nitrogen in these fluid inclusions are interpreted to have had an organic-rich pelitic or graphitic source (Sheppard, 1986). Aqueous-carbonic fluid

inclusions are rare in the granitic rocks in Cornwall (Rankin and Alderton, 1983) but are common in the contact metamorphic aureole (Wilkinson, 1990). Wilkinson (1990) presented evidence of fluid immiscibility and proposed a mixed magmatic-metamorphic evolution for the fluids in the metasediments. Based on the relationship between C, S and Sn in the metamorphic rocks and distance from the granite we propose a similar origin for the aqueous-carbonic fluids at Pilok, although a magmatic origin cannot be entirely discounted. Conceptually this could involve two end-members (cf. Wilkinson, 1990): 1) Contact metasomatism. Expulsion of magmatic fluids from the granite that interact with the metamorphic rocks. These fluids are metasomatic (i.e., have added B in the tourmaline zone and potentially K in the biotite zone) and through wallrock interaction have leached carbon. 2) Contact metamorphism. In this case the heat from the granite was at first transferred conductively and metamorphic reactions produced C-O-H-N fluids that later mixed with magmatic fluids.

Contact metasomatism was undoubtedly responsible for the creation of the tourmaline zone at Pilok, whereas the sericitized porphyroblasts in the biotite zone are indicative of an early contact metamorphic event that was overprinted by a fluid event (magmatic-metamorphic fluid mixing). The primary reason for rejecting fluid immiscibility for the origin of the liquid-rich and vapour-rich aqueous-carbonic fluid inclusions is the lack of CH_4 and NaCl partitioning. However if immiscibility occurred, once the liquid and vapour separated they could react independently with

the surrounding wallrock. Thus it is possible that liquid and vapour could have significant CH_4 if both equilibrated with reduced granite or metasediment.

We suggest that mixing between magmatic and metamorphic fluids produced a range of aqueous-carbonic fluid compositions. Based on the erratic distribution of tin in the leucocratic granite Lehmann and Mahawat (1989) proposed that tin was remobilized at Pilok. We concur, and further propose that early reduced aqueous-carbonic fluids leached Sn and W from the granitic rocks. Cassiterite probably precipitated as a result of increasing $f\text{O}_2$ and/or decreasing temperature and/or increasing pH (cf. Eugster, 1986).

CONCLUSIONS

The W-Sn-polymetallic mineralization at Pilok is strongly associated with the intrusion of a leucocratic granite. Minor wolframite $\pm$ cassiterite mineralization was deposited at an early post-magmatic stage and is hosted by pegmatite pods. The bulk of the wolframite mineralization post-dated tourmalinization and pre-dated greisen alteration, but the temperature and pressure of this event is poorly constrained.

Cassiterite mineralization is associated with greisen alteration of leucocratic granite. The mineralizing fluid is interpreted to have been of mixed magmatic-metamorphic (or evolved meteoric) origin, with carbon and probably nitrogen having been derived from the pelitic metasediments. Some boiling, or fluid

immiscibility, may have occurred during the evolution of the fluid but, if so, was not related to cassiterite deposition.

Although there are differences between the mineralization at Pilok and that at Panasqueira and Cornwall, the fluid evolution in all these areas was broadly similar.

ACKNOWLEDGEMENTS

This research was carried out with the aid of a grant from the International Development Research Centre, Ottawa, Canada. We would like to thank S. Bunopas, N. Jungyusuk, S. Khositantont, and P. Putthapiban of the Geological Survey Division, Department of Mineral Resources of Thailand for their support and advice during the course of this project. We would also like to thank the Somsak Mining Group for their hospitality and help during the fieldwork. The manuscript was improved by comments from J. Clark.

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CHAPTER 6

DEPOSITS IN THE TAKUA PIT THONG AREA

INTRODUCTION

The Takua Pit Thong (also known as Tako Pit Thong) area is located approximately 150 km west of Bangkok (Fig. 1) and hosts several tin and feldspar mines (Fig. 2). The Takua Pit Thong (1 to 4), Huai Nam Sai, and Muang Rae Gow are all sulphide and/or magnetite replacement tin deposits. During the course of the field investigation Takua Pit Thong deposits 1, 2, and 3 were producing approximately 10 tonnes of cassiterite concentrate per month from 1000 tonnes of rock. Tungsten and copper are present in significant concentrations, primarily as scheelite and chalcopyrite, however, neither of these metals were being recovered. Kow Daeng contains "pegmatite", skarn and breccia-hosted mineralization, from which Sn-Ta-Nb are recovered. Feldspar was being mined from the pegmatites in both the Takua Pit Thong and Kow Daeng areas. These pegmatites contain traces of cassiterite although none was observed during this investigation. Feldspar is also being mined from the Nam Pung deposit located south of Takua Pit Thong (Fig. 2). The Huai Saphan area hosts alluvial cassiterite, which, according to the mine owner is "being mined from the soil near the Burmese (Myanmar) border". The Huai Saphan deposit was not visited and will not be discussed further. The

chemistry of the Takua Pit Thong granites were the focus of a study by Suvunsavate (1986) and brief descriptions of this area are also provided by Lehmann (1988) and Charusiri (1989). In this study the geology of the deposits in the Takua Pit Thong area are described and preliminary fluid inclusion data on the sulphide replacement deposits are presented. Tentative interpretations on the evolution of the Takua Pit Thong deposits are proposed based on these data.

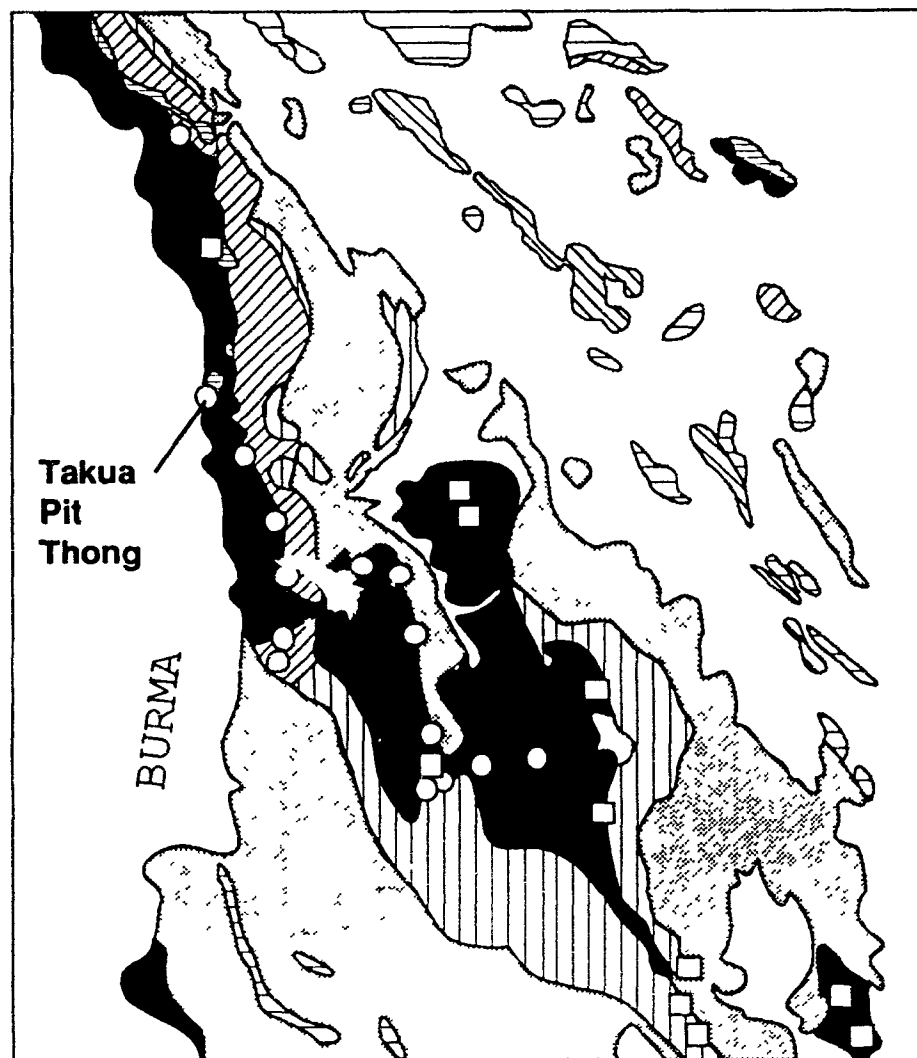
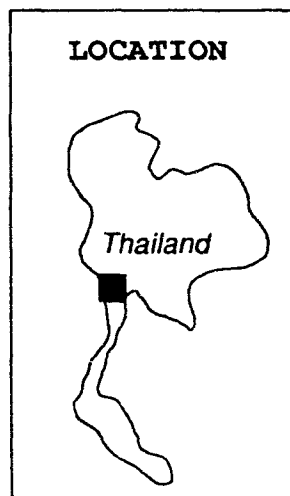
GEOLOGICAL SETTING

Four types of granitic rocks are present in the Takua Pit Thong area. Biotite granite is the most abundant and is either fine-grained and equigranular or is medium-grained with K-feldspar phenocrysts. It is composed primarily of quartz, K-feldspar, plagioclase, and biotite with accessory garnet, tourmaline, and muscovite. The second-most abundant type of granitic rock is leucocratic granite/aplite. This unit is typically fine-grained and is characterized by the occurrence of garnet, and/or tourmaline and/or muscovite. Feldspar compositions were determined by electron microprobe from one leucocratic granite sample which contains albite (Ab_{97} to Ab_{98}) and K-feldspar (Or_{95} to Or_{98}) (Appendix VI). The third unit, pegmatite, is less common. The pegmatites are mineralogically simple, consisting of quartz, K-feldspar, and albite, with minor amounts of garnet, tourmaline, and rare cassiterite and beryl. Non-mineralized pegmatites (those on Fig. 2) are composed of coarse-grained perthitic K-feldspar set in a matrix of quartz

Figure 1. Regional geological map modified after Dheeradilok (1981) showing the location of the tin and fluorite deposits, including Takua Pit Thong.

- Q** alluvial sand and gravel
- K-T** granite
- Mz** sandstone and conglomerate
- P** limestone
- C** pebbly mudstone, sandstone
- c** carbonaceous quartzite
- SD** quartzite
- O** limestone and calc-silicate

- tin deposit
- fluorite deposit



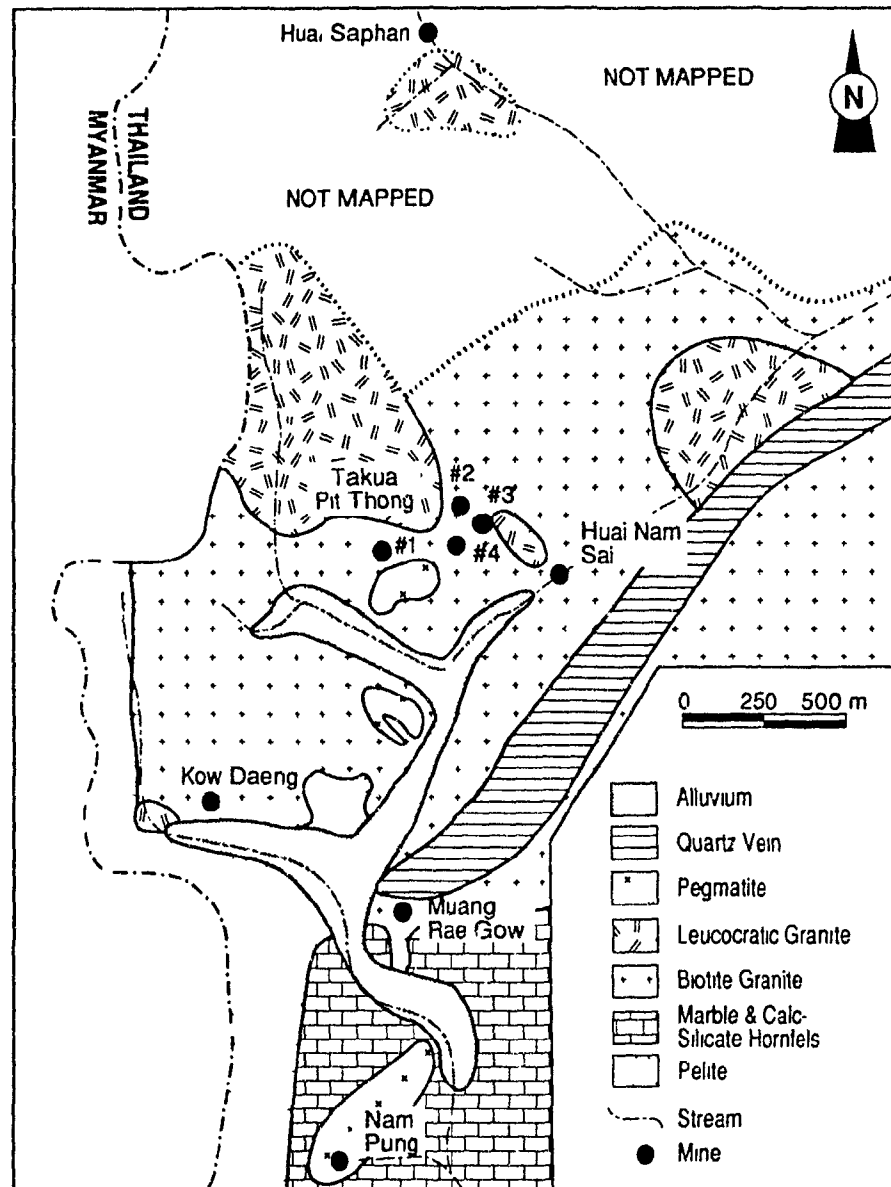
0 10 20 30 km

Figure 2. A geological map of the Takua Pit Thong area, showing the geological setting and location of the principal deposits.

99°10'E

99°11' 6"E

14°41' 8"N



14°39' 5"N

and muscovite $\pm$ albite. The mineralized "pegmatites" at Kow Daeng are albite-rich and are discussed below. The fourth igneous unit occurs only in the vicinity of the Takua Pit Thong ore bodies and has been termed quartz-feldspar rock. It consists of quartz, K-feldspar, and plagioclase, with accessory garnet, tourmaline, and muscovite. This unit has been distinguished from the pegmatite unit because it has both magmatic and metasomatic features. The plagioclase varies in composition from An_0 to An_{13} (Appendix VI), which suggests that the rock was either metasomatized at high temperatures, or that it crystallized directly from a silicate melt or from fluids coexisting with melt.

Two metasedimentary units occur in the area, marble and calc-silicate hornfels, and pelitic metasediment (Fig. 2). The marble and calc-silicate hornfels are composed of calcite, phlogopite, and diopside and are probably of Ordovician age. In the vicinity of the ore deposits calc-silicate hornfels is strongly banded and contains clinopyroxene, garnet, and idocrase-rich layers with minor amounts of quartz, K-feldspar, plagioclase, and titanite. The clinopyroxene is diopside-rich; its composition will be discussed in greater detail in the skarn section. Idocrase and titanite both contain significant fluorine, up to 2.01 and 2.31 wt %, respectively (Appendix VI). Titanite also contains up to 0.1 wt % SnO_2 in unmineralized hornfels adjacent to sulphide replacement deposits. The high concentrations of F and Sn in the hornfels suggest that these rocks have been partly metasomatized.

The pelitic metasediment is carbonaceous, poorly sorted, and fine-grained (<

5 mm), with ellipsoidal quartz grains up to 3 mm in diameter. It is probably either Siluro-Devonian or Carboniferous-Permian in age, and is composed largely of quartz and muscovite, with local biotite replacement of muscovite. In roof pendants, this unit forms a hornfels, composed predominantly of quartz and biotite, with subordinate plagioclase, K-feldspar, and titanite.

The total and reduced carbon contents and the total sulphur and sulphate contents of marble and calc-silicate hornfels adjacent to mineralization, their regional equivalents, and regional pelitic rocks are presented in Table 1. It is apparent that the regional marble and calc-silicate rocks contain higher reduced (insoluble) carbon abundances than they do adjacent to mineralization. The regional metamorphic pelite also contains significant reduced carbon (up to 5.41 wt %; Table 1). The sulphur contents are highly variable, but the calc-silicate hornfels in the zones of mineralization typically contains more sulphur than its regional equivalent suggesting that sulphur may have been added to this rock.

MINERALIZATION

TAKUA PIT THONG

Cassiterite-sulphide and cassiterite-oxide replacement deposits at Takua Pit Thong are hosted by calcareous roof pendants in granite. The age of mineralization determined by plateau and total fusion $^{40}\text{Ar}/^{39}\text{Ar}$ analyses on biotite associated with

Table 1
Carbon and Sulphur Contents of Metasedimentary Rocks

	# of anal.	wt % ΣC	wt % insol. C	wt % ΣS	wt % insol. S
marble regional	4	13.00 (12.52-13.44)	0.14 (0.01-0.35)	0.07 (0.01-0.09)	0.03 (0.01-0.04)
marble deposit	2	13.91 (13.64-14.17)	0.03 (0.02-0.03)	0.02 (0.01-0.03)	0.02 (0.01-0.02)
calc-sil. regional	3	3.93 (1.27-8.07)	0.21 (0.01-0.61)	0.16 (0.01-0.43)	0.03 (0.01-0.05)
calc-sil. deposit	5	0.15 (0.08-0.18)	0.03 (0.01-0.10)	0.81 (0.01-1.63)	0.18 (0.01-0.49)
pelite regional	4	3.04 (0.37-5.41)	2.70 (0.25-5.08)	0.21 (0.01-0.43)	0.02 (0.01-0.05)

Total and insoluble carbon and sulphur analysed by LECO, detection limit 0.01%. The numbers represent mean values, with the ranges in parentheses

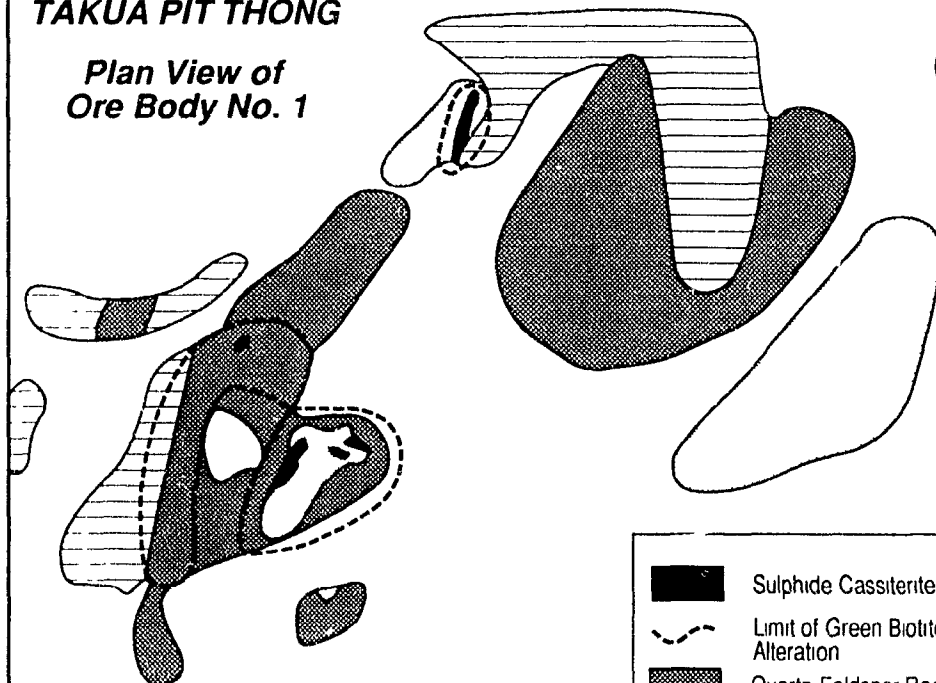
cassiterite-pyrrhotite-pyrite ore is 68.9 Ma and 69.7 Ma, respectively (Charusiri, 1989). Three tin deposits were being exploited during the course of this study, Takua Pit Thong #1, #2, and #3. Cassiterite at ore body #1 is spatially associated with a small calc-silicate roof pendant (Fig. 3). Quartz-feldspar rock surrounds the roof pendant and hosts some tin mineralization. The lack of quartz-feldspar rock distal from the roof pendants suggests that it may be metasomatized granite. There is a strong spatial and temporal relationship between cassiterite and green biotite alteration and pyrite or pyrrhotite replacement of metasedimentary rocks. Cassiterite at the #2 and #3 ore bodies have similar modes of occurrence to the #1 ore body (Fig. 4) except that it also occurs with minor magnetite replacement of hornfels. Not all roof pendants are mineralized; cassiterite-bearing and altered rocks are only localized in roof pendants that are cut by minor faults. This is particularly well developed at ore body #3. The #4 ore body (Fig. 2) was dominantly a magnetite replacement deposit (Suvunsavate, 1986), however, the adit from which it was mined collapsed and thus access to this deposit is not possible. An abandoned magnetite replacement ore body is also located at Muang Rae Gow. The only other replacement body known to occur in this area is Huai Nam Sai (Fig. 2). This is a small showing of granite-hosted cassiterite-sulphide mineralization. There are no metasedimentary outcrops at the showing but it is otherwise similar to the deposits at Takua Pit Thong.

There were three main stages of hydrothermal alteration; skarn, followed by

Figure 3. A detailed geological map of the Takua Pit Thong # 1 deposit.

TAKUA PIT THONG

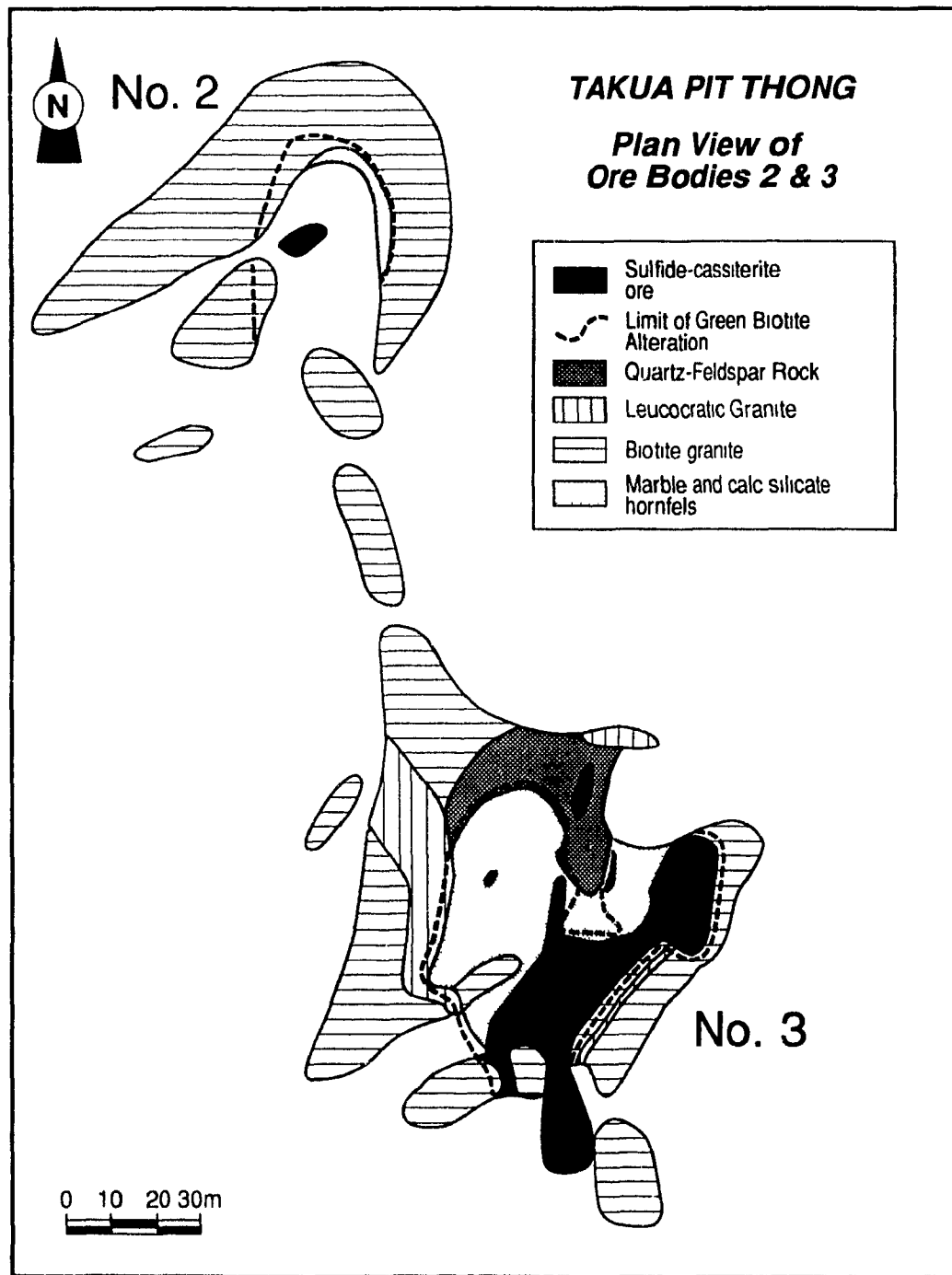
Plan View of
Ore Body No. 1



0 10 20 30m

-  Sulphide Cassiterite Ore
-  Limit of Green Biotite Alteration
-  Quartz-Feldspar Rock
-  Biotite granite
-  Marble and calc silicate hornfels

Figure 4. A detailed geological map of the Takua Pit Thong # 2 and 3 deposits.



retrograde calc-silicate, and finally biotite alteration. Clinopyroxene-garnet skarn is present in minor amounts at Takua Pit Thong. It is more abundant in the Kow Daeng area and hence will be described in the section devoted to that deposit. Retrograde calc-silicate alteration is dark green, medium-grained, and typically occurs as haloes around quartz-sulphide veins or fractures which crosscut the calc-silicate hornfels. It consists of tremolite, clinozoisite, K-feldspar, fluorite and carbonate. Pervasive tremolite-fluorite replacement of hornfels is also locally observed. Tremolite is typically Mg-rich, X_{Mg} (Mg/Mg+Fe molar ratio) ranges from 0.68 to 0.98, is enriched in fluorine, containing up to 1.55 wt % F (Appendix VI). A minor amount of cassiterite is associated with amphibole alteration.

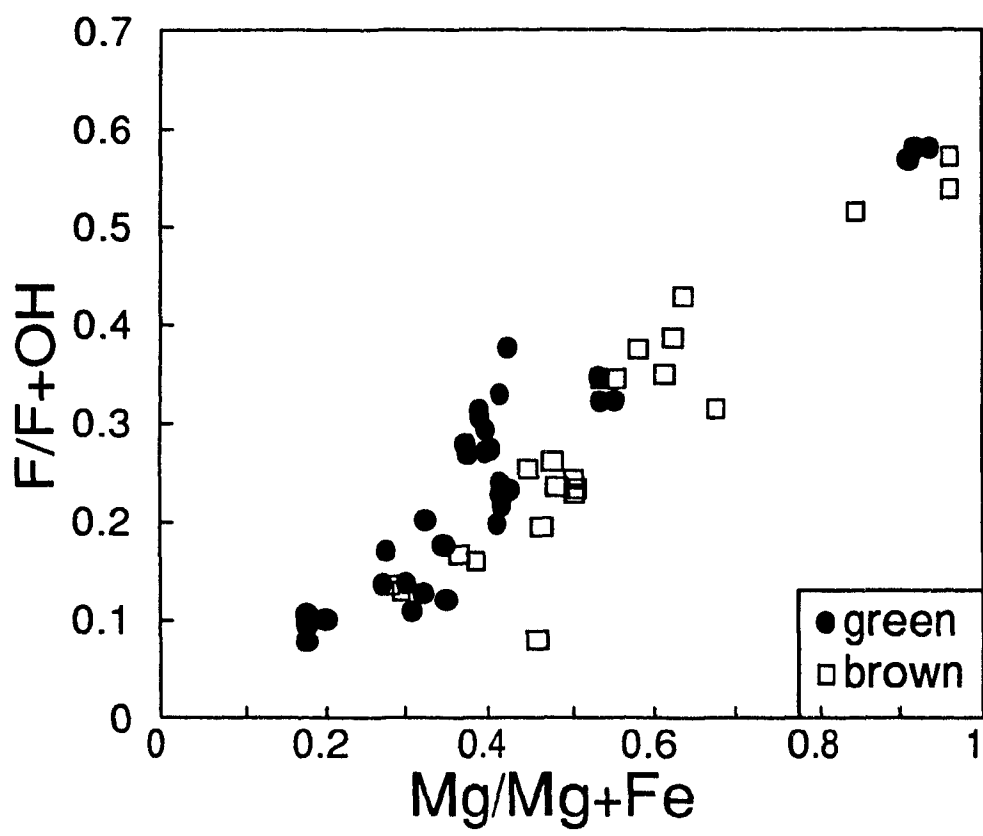
Two types of biotite alteration, early dark brown and late green, have been recognized at Takua Pit Thong. There is a strong spatial and temporal relationship between the green biotite alteration, sulphide and oxide replacement, and cassiterite mineralization. The limits of green biotite alteration in each of the sulphide replacement bodies are shown on Figures 3 and 4. Brown biotite occurs as pervasive replacement of hornfels or as alteration haloes around quartz $\pm$ sulphide veins. In the case of the hornfels, the original layering is preserved as almost monomineralic, alternating, brown biotite and fluorite horizons. Coarse-grained brown biotite also occurs as an early fracture-related alteration of quartz-feldspar rock.

Pervasive green biotite alteration overprints brown biotite in altered hornfels

and granite near roof pendant-granite contacts. This alteration consists of massive green biotite and locally is accompanied by one or more of fluorite, danalite $[(\text{Fe,Mn,Zn})_4(\text{Be}_3\text{Si}_3\text{O}_{12})\text{S}]$, quartz, tourmaline, beryl, and carbonate. Danalite is commonly observed in tin skarns, e.g., at the JC deposit (Layne and Spooner, 1991). At Takua Pit Thong it is dark red, commonly has a triangular crystal form, and is strongly associated with green biotite $\pm$ sphalerite $\pm$ cassiterite.

In spite of the distinct temporal and colour differences, brown and green biotite are chemically similar (Appendix VI). Both types of biotite have a broad range of Fe-Mg compositions, although the green biotite is typically Fe-rich and brown biotite Mg-rich. Individual biotite grains generally have constant compositions, however, in one sample green biotite is strongly zoned with X_{Mg} core compositions of 0.91 and rim compositions of 0.42 (Appendix VI). The fluorine contents of brown and green biotite at a given X_{Mg} composition are similar and there is a linear relationship between F and Mg abundance (Fig. 5). The amount of fluorine in biotite (and other silicates) is strongly dependent on the X_{Mg} composition and to a lesser extent is temperature dependent (Munoz, 1984). The overlap of data for brown and green biotite on Figure 5 suggests that both types of biotite formed from fluids with similar temperatures and $f\text{HF}/f\text{H}_2\text{O}$ ratios. The fluorine intercept values of biotite at Takua Pit Thong range from 1.07 to 1.56, and are comparable to the compositions of biotite at the Moina and Mt. Lindsay F-rich tin skarns in Tasmania (Kwak and Askins, 1981; Kwak, 1983). This further

Figure 5. Cation ratios of $F/F+OH$ versus $Mg/Mg+Fe$ of biotite from Takua Pit
Thong.



demonstrates that Takua Pit Thong is a fluorine-rich system.

Although the bulk of the cassiterite mineralization is contemporaneous with biotite alteration, the highest concentrations of cassiterite are in sulphide or oxide replacement bodies. Since sulphide-quartz veins have biotite alteration haloes, the sulphide replacement bodies are interpreted to have been coeval with biotite alteration. Cassiterite mineralization occurs in three types of sulphide-oxide replacement bodies: pyrrhotite, pyrite, and magnetite. All three types can be either massive or banded. In banded ore, the sulphides/oxides occur in layers, 0.5 to 3.0 cm thick, alternating with biotite-rich layers. Cassiterite is generally intergrown with sulphides or oxides and is less abundant in the biotite-rich layers. Magnetite ore typically contains only trace pyrrhotite, chalcopyrite, pyrite and scheelite. The chalcopyrite, pyrrhotite and scheelite are locally in apparent equilibrium with magnetite, whereas pyrite invariably is present as a replacement product. Pyrrhotite and pyrite ores commonly contain accessory chalcopyrite, arsenopyrite, sphalerite, scheelite, and wolframite. Lehmann (1988) also reported some native bismuth and bismuthinite with the pyrrhotite-rich ore. Pyrite typically replaces pyrrhotite where the two minerals are intergrown, and magnetite and pyrrhotite are observed as inclusions in cassiterite, but pyrite is not. This suggests that pyrite may have formed after cassiterite mineralization.

Figure 6 shows the fO_2 - fS_2 relationships of some of the major alteration minerals at Takua Pit Thong. The diagram was constructed to represent conditions

at 500°C and 500 bars (pressure and temperature are discussed below) using the SUPCRIT data base (Helgeson et al., 1978) and annite to represent biotite. The addition of Mg to annite will expand the biotite stability field, but as green biotite is typically Fe-rich and the use of the annite end member is therefore justified. The assemblages annite-pyrrhotite-chalcopyrite and annite-magnetite-chalcopyrite have very restricted stability fields. The presence of both magnetite and pyrrhotite ore at Takua Pit Thong can therefore be explained by small changes in sulphur or oxygen fugacities.

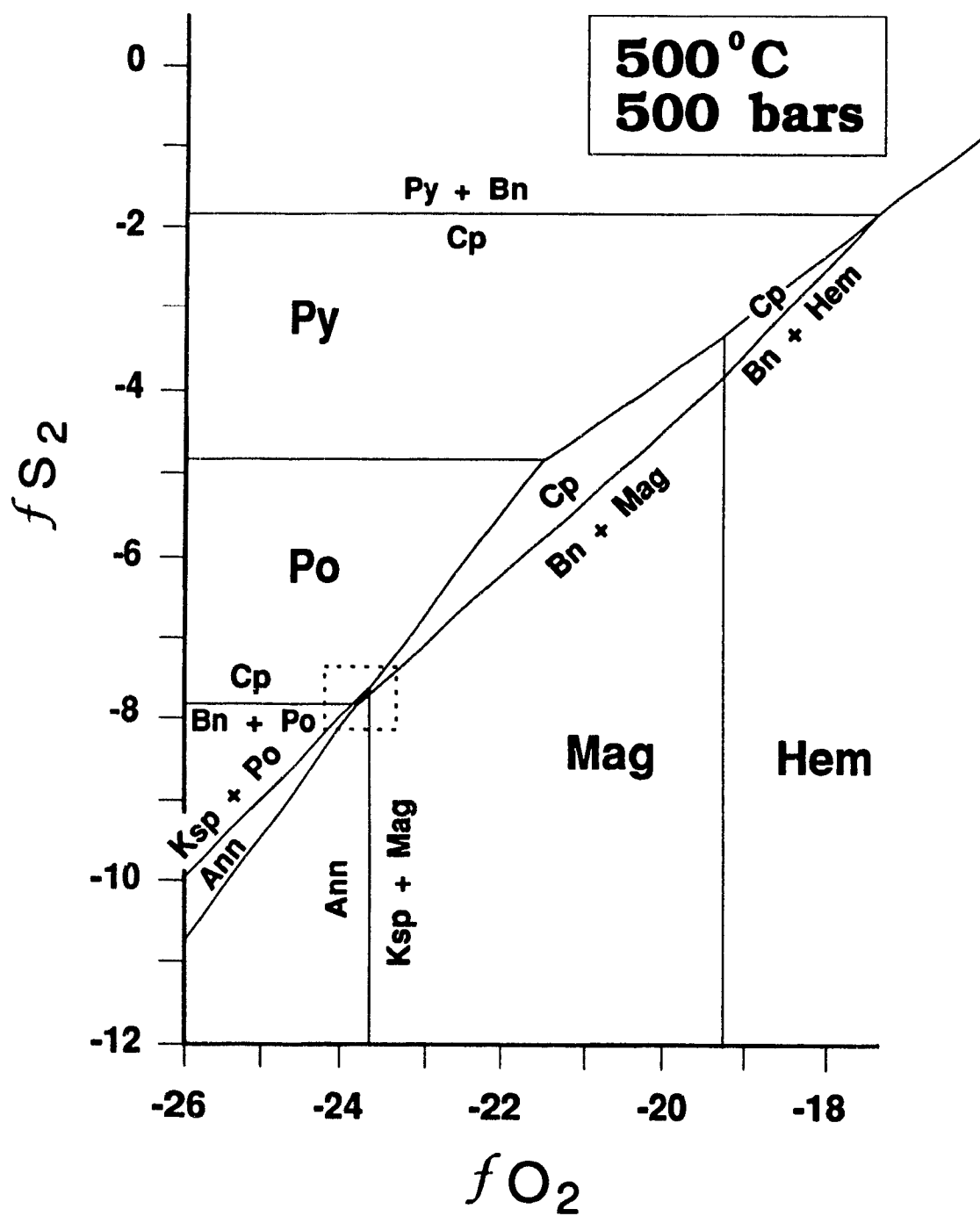
KOW DAENG

Cassiterite mineralization at the Kow Daeng deposit is hosted by "pegmatite", skarn, and breccia (Fig. 7). "Pegmatite" is locally replaced by skarn and both are found as fragments in the breccia. The breccia hosts the bulk of the cassiterite at this deposit, however, it is highly weathered and cassiterite is not visible. Most of the visible cassiterite is pegmatite-hosted with minor amounts present in skarn, although the latter also hosts significant disseminated scheelite.

The "pegmatite" is composed largely of albite with subordinate quartz $\pm$ K-feldspar $\pm$ garnet $\pm$ tourmaline $\pm$ biotite. It is possibly magmatic although a metasomatic origin (albitization) cannot be excluded. This unit is restricted to being adjacent to skarn and, like the quartz-feldspar rock, may indicate a metasomatic rather than magmatic origin. Albite (Ab_{96} to Ab_{98}) and quartz are

Figure 6. $f\text{O}_2$ - $f\text{S}_2$ relationships for pyrrhotite and magnetite replacement ore.

Abbreviations are Ann annite, Bn bornite, Cp chalcopyrite, Hem hematite, Ksp K-feldspar, Mag magnetite, Po pyrrhotite and Py pyrite. The $f\text{O}_2$ - $f\text{S}_2$ conditions for sulphide-oxide replacement at Takua Pit Thong line within the dashed box.

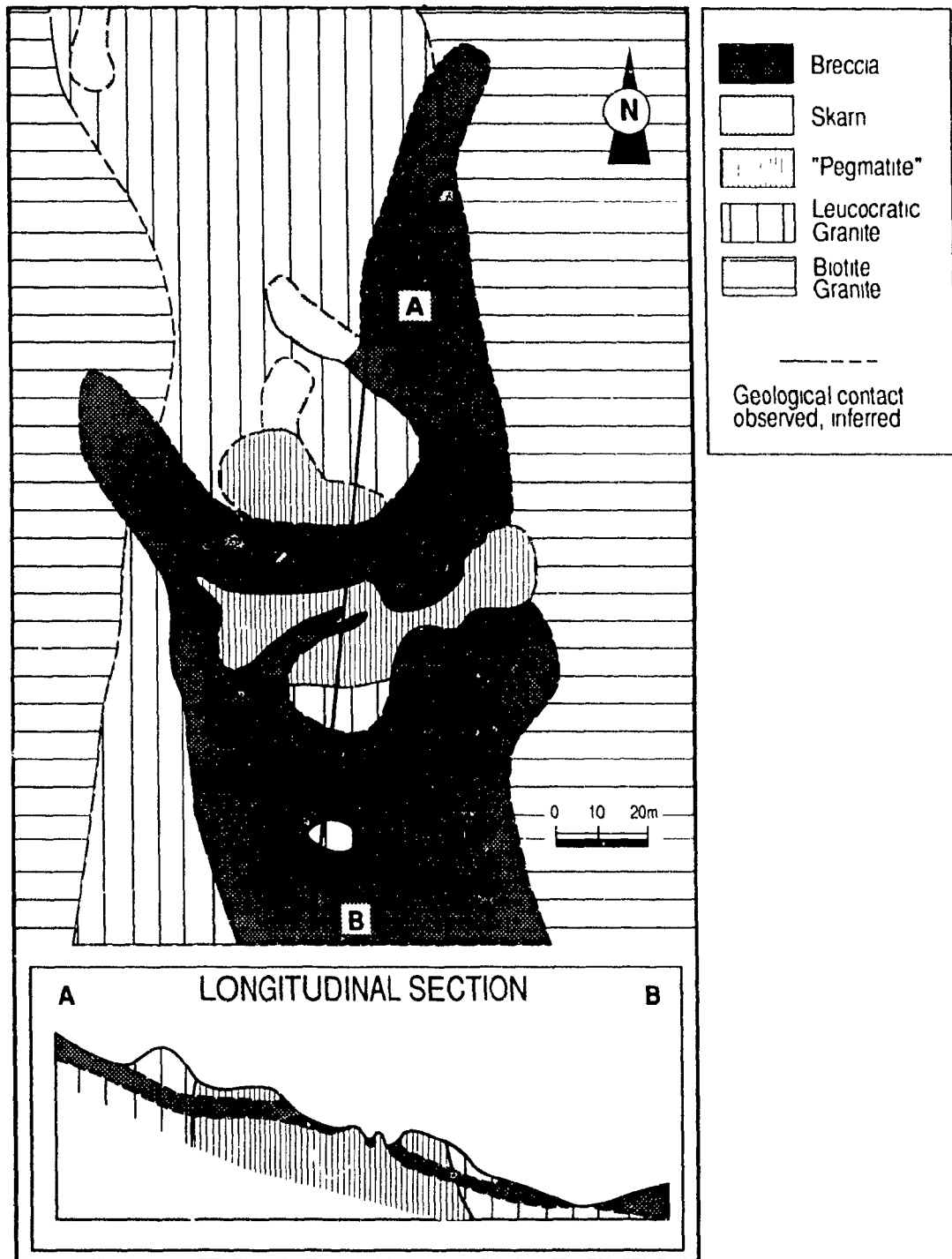


equigranular, and K-feldspar (Or_{95} to Or_{98}) is present either as coarse-grained perthitic phenocrysts or is interstitial and untwinned (for feldspar analyses see Appendix VI). Garnet and tourmaline are disseminated. Cassiterite is also disseminated, as sub- to euhedral crystals up to 3 cm long, and is generally intergrown with albite. It typically contains solid inclusions of Nb-Ta minerals and these inclusions locally define cassiterite growth bands. Two types of biotite are recognized. Dark brown biotite is early, locally interstitial, but has also replaced garnet and therefore is probably an alteration product. Green biotite is later and is clearly metasomatic, occurring with epidote as a replacement of albite along fractures. Minor pyrite and chalcopyrite are associated with green biotite, which indicates that this alteration is analogous to the main stage of mineralization in the Takua Pit Thong deposits (green biotite-sulphide-cassiterite). In contrast to the latter deposits, "pegmatite"-hosted cassiterite pre-dates green biotite alteration which is restricted to fractures with minor associated sulphides. The only pervasive alteration of pegmatite is replacement by skarn.

Skarn at the Kow Daeng deposit consists of prograde and retrograde stages. The prograde skarn is massive, composed dominantly of clinopyroxene and garnet, is typically rich in clinopyroxene, and can contain variable amounts of fluorite, idocrase, titanite, scheelite, and cassiterite. The paragenesis of idocrase, fluorite, and scheelite is poorly constrained. Idocrase generally occurs as large optically continuous grains with inclusions of fine-grained clinopyroxene, and thus is

Figure 7. A detailed geological map and longitudinal section of the Kow Daeng deposit.

PLAN VIEW



synchronous with, or later than the pyroxene. Idocrase is also observed in quartz veins where it is replaced by epidote. Fluorite has a similar mode of occurrence to idocrase, i.e., it is coarse-grained with clinopyroxene inclusions. The retrograde skarn has replaced prograde skarn along fractures or occurs as haloes around quartz veins. It is composed of tremolite, epidote/clinozoisite and calcite, and is therefore analogous to the retrograde calc-silicate alteration at Takua Pit Thong.

Skarn probably formed by replacement of marble roof pendants in granite, however, unaltered (or partially altered) marble is not present at Kow Daeng. Prograde skarn is typically in sharp contact with pegmatite or leucocratic granite, veins are lacking, and albite is replaced by clinopyroxene. Less commonly, quartz veins with prograde skarn alteration haloes cut pegmatite. Minor cassiterite is disseminated in clinopyroxene skarn indicating that it probably precipitated during the prograde skarn event. However, cassiterite is locally replaced by clinopyroxene suggesting that some tin was remobilized. In the prograde skarn cassiterite is also present as inclusions in titanite. Analyses of this titanite show that it contains up to 5.42 wt % SnO_2 (Appendix VI). High tin abundances in titanite are not unusual as this mineral has a tin end-member, malayaite. Scheelite is disseminated in prograde skarn, where, typically, it is spatially related to fluorite.

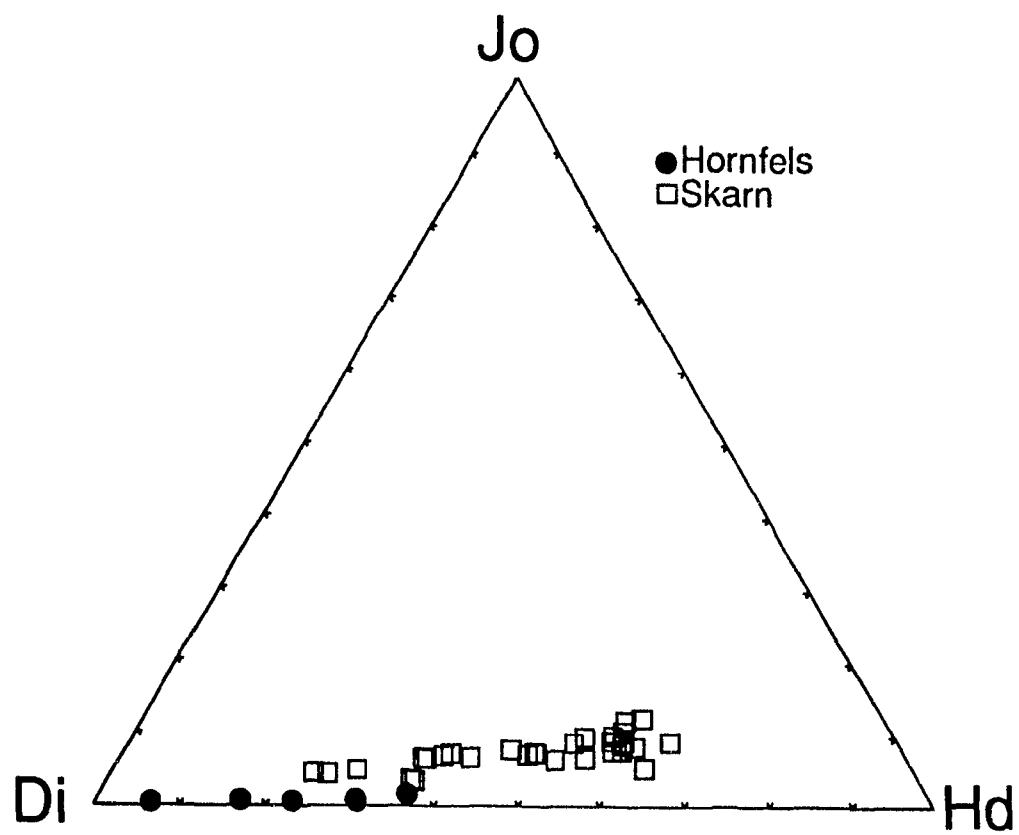
The compositions of clinopyroxene and garnet from skarn have been determined by electron microprobe (Appendix VI). Clinopyroxene from skarn is Fe-rich and there is a clear iron and manganese enrichment trend from hornfels to

skarn (Fig 8). Garnet in skarn has a more consistent composition X_{Gr} from 0.65 to 0.69, X_{Sp} from 0.12 to 0.14, X_{And} from .11 to 0.17, X_{Alm} from 0.04 to 0.09, and $X_{Py} < 0.003$ where X_{Gr} , X_{Sp} , X_{And} , X_{Alm} , and X_{Py} are the mole fractions of grossular, spessartine, andradite, almandine, and pyrope, respectively. These pyroxene-garnet compositions are similar to those obtained from other tin skarns (Einaudi *et al.*, 1981).

Most of the cassiterite at Kow Daeng is breccia-hosted. The breccia consists of angular to subangular fragments of pegmatite, skarn, and vein quartz up to 10 cm across, supported by a fine-grained dark brown matrix. Unfortunately the breccia is pervasively weathered and no cassiterite is visible. The breccia was therefore not studied in detail. From the longitudinal section based on the surface exposure, the breccia is interpreted to be a thin, shallowly dipping, sheet (Fig. 7). This suggests that it may have a tectonic origin which is supported by the presence of slickensides along some shear planes. However, the matrix is typically massive, lacking a structural fabric, which suggests that there was also a hydrothermal component to its formation.

The striking contrast of F-Be-Cu-Zn-As-S enrichment at Takua Pit Thong and Nb-Ta enrichment at Kow Daeng suggests that the mineralizing fluids at the two deposits were different. The two deposits can however, be related using the timing of the different alteration-mineralization events. Pegmatites at Kow Daeng host the earliest mineralization and are overprinted by successive skarn and green

Figure 8. Compositions of clinopyroxene from Takua Pit Thong (hornfels) and Kow Daeng (skarn) in terms of mole fractions of johannsenite (Jo), diopside (Di), and hedenbergite (Hd)



biotite alteration. Sulphides are associated with the green biotite alteration at Kow Daeng which strongly suggests that this event was coeval with green biotite-sulphide-cassiterite mineralization at Takua Pit Thong. The breccia at Kow Daeng contains fragments of pegmatite and skarn and thus is the youngest style of mineralization.

Mineralization, as constrained by the timing of alteration, therefore evolved from early Sn-Nb-Ta to late Sn-Cu-Zn-As-S-Be-F stages. There are two explanations for this evolution. Either there were two different source granites or the compositions of magmatic fluids from a single granitic source evolved during late-stage crystallization. The conclusion that fluorine-rich fluids were late suggests a single granite source where fluorine was partitioned into the melt at the onset of vapour saturation and was released into fluids only at the latest stages of granite crystallization or at early sub-solidus conditions (cf. Manning and Pichavant, 1984). The close proximity of Takua Pit Thong to Kow Daeng suggests that two different granite sources are unlikely. Takua Pit Thong and Kow Daeng are also linked by the compositions of the silicate minerals which show iron enrichment trends. In the case of Takua Pit Thong iron enrichment in amphibole and biotite is spatially and temporally related to cassiterite crystallization, and at Kow Daeng, iron enrichment of clinopyroxene is similarly related.

FLUID INCLUSIONS

PETROGRAPHY

Five types of fluid inclusions have been distinguished at the Takua Pit Thong deposits in cassiterite, scheelite, quartz, and fluorite. Type 1 inclusions consist of aqueous liquid and vapour (L-V). Some of these inclusions contain a carbonic vapour, that is evident from double freeze phenomena (Collins, 1979), and may also contain one or two solids. These solids have either a yellow-white or red-yellow birefringence. Type 2 inclusions are vapour-rich and aqueous-carbonic consisting of vapour (V) or vapour and liquid (V-L) at 25°C. Type 3 inclusions are three-phase, containing aqueous and carbonic liquid and carbonic vapour (L-L-V). Type 4 inclusions are halite bearing (L-V-H). In addition to halite they also commonly have an additional birefringent solid similar to those observed in type 1 inclusions. Type 5 inclusions are liquid only (L). Fluid inclusions of each of the types identified above generally have rounded or elliptical shapes and are typically less than 30 μm across. Fluid inclusion types 1, 2, 3 and 5 are observed in quartz, cassiterite, clinopyroxene, fluorite, scheelite, and idocrase at the Kow Daeng deposit.

Takua Pit Thong

Type 1 inclusions are the dominant type in cassiterite, type 2 inclusions are

less common, and type 4 inclusions are rare. Cassiterite crystals are typically euhedral and strongly zoned. Inclusion types 1, 2, and 4 occur in the centres of grains and are surrounded by clear overgrowths of cassiterite (Fig. 9). The boundaries between the overgrowths and the crystal centres show resorption of the central (inclusion-bearing) portions of the crystals. This indicates that these inclusions are primary. Solid inclusions of green biotite and quartz needles are also common in cassiterite. Less commonly, the cassiterite contains inclusions of pyrrhotite, magnetite and wolframite. Green biotite inclusions typically occur near the crystal centre/overgrowth boundary and indicate that cassiterite precipitated during green biotite alteration.

Quartz contains inclusion types 1, 2, 3, and 4, most of which are secondary. Type 1 inclusions are the most abundant and type 2 are the second most abundant. Some type 1 and 2 inclusions have been observed together along secondary planes. Type 4 inclusions are less common and also delineate secondary planes.

All the inclusion types are represented in fluorite. Inclusion types 1, 2, 4 and 5 are all common, whereas type 3 inclusions are rare. Almost all the inclusions lie along secondary planes. Type 1 inclusions coexist with type 2 inclusions along the same secondary plane, as in quartz. Type 5 inclusions (one phase L) generally occur in the same secondary plane as type 1 inclusions with very low V/L ratio.

Figure 9. Representative photomicrographs of primary type 1 (L-V) fluid inclusions in cassiterite from the Takua Pit Thong deposits.

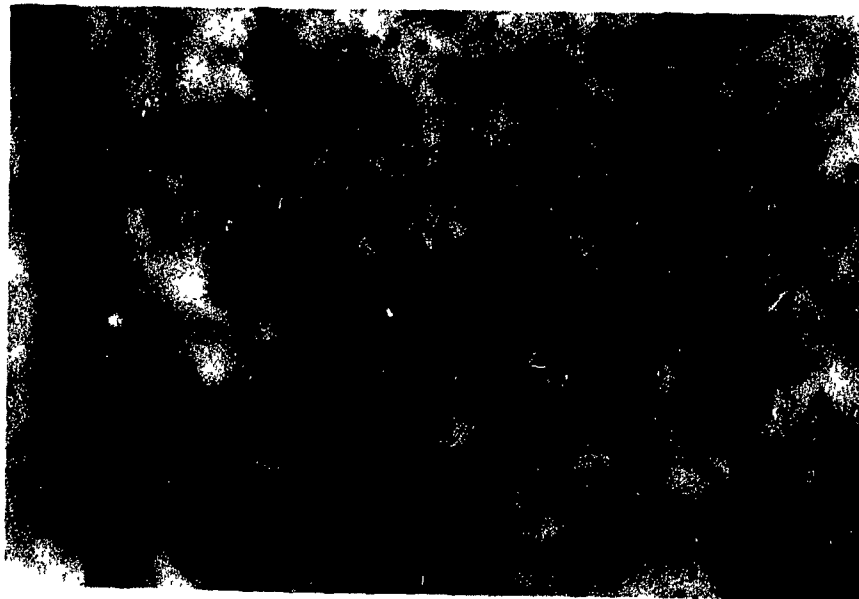
a) A fluid inclusion-rich central portion of the cassiterite grain is embayed and overgrown by clear cassiterite. Field of view 1.4 mm across. The box shows the area of Fig. 4b.

b) Primary type 1 (L-V) inclusions in cassiterite at the clear cassiterite overgrowth boundary. Field of view 0.45 mm across.

a



b



Kow Daeng

Four inclusion types are observed in quartz at the Kow Daeng deposit. In quartz-epidote veins the dominant inclusion type is type 1, type 2 inclusions are also common, and type 5 are less abundant. A single solid with second order birefringence is commonly present in the type 1 inclusions. All four inclusion types are observed along secondary planes and type 1 and 2 inclusions, or type 1 and 5 inclusions coexist along the same secondary plane. Only fluid inclusion types 1 and 2 are present in pegmatite quartz. Both types are commonly observed along secondary planes and locally they coexist along the same plane. Quartz from skarn and pegmatite at the Kow Daeng deposit hosts abundant fluid inclusions, which range up to 20 μm , across.

"Pegmatite"-hosted cassiterite contains similar proportions of types 1 and 2 fluid inclusions. Clear cassiterite, between growth bands rich in inclusions of Ta-Nb minerals (columbite?; described above), contain primary type 1 and 2 inclusions that lie parallel to the growth bands. Type 1 and 2 inclusions also commonly occur along short planes within crystals, approximately 100 μm long, and are thus probably pseudosecondary in origin. Inclusions in cassiterite are typically <5 μm across and have spherical or irregular shapes.

Clinopyroxene from skarn hosts two fluid inclusion types. Type 1 ($\pm$ solid) inclusions are dominant and type 2 inclusions are less common. The solid in type

1 inclusions generally has white birefringence. Type 1 ($\pm$ solid) inclusions are present along planes that do not cross grain boundaries and therefore are probably pseudosecondary. Fluid inclusions commonly have negative crystal shapes, controlled by the two pyroxene cleavages, and range up to 10 μm across.

In other skarn minerals, i.e., idocrase, fluorite, and scheelite, only secondary type 1 and 2 fluid inclusion are observed; both fluid inclusions types are common. The fluid inclusions in fluorite and scheelite have spherical shapes whereas those in idocrase they have negative crystal shapes, elongated parallel to the c-axis.

MICROTHERMOMETRIC ANALYSIS

Fluid inclusion compositions were determined by microthermometry and scanning electron microscopy with an energy dispersive system. The microthermometric analyses were determined on a Fluid Inc. stage calibrated using synthetic fluid inclusion standards. Measurements of phase changes at or below 31°C are accurate to within $\pm 0.2^\circ\text{C}$ and high temperature measurements to within $\pm 2.0^\circ\text{C}$. Fluid inclusion decrepitates were analyzed on a Cambridge scanning electron microscope equipped with a Tracor-Northern energy dispersive analyzer after being heated to 400° and $>600^\circ\text{C}$ on the fluid inclusion stage. Fluid inclusions in cassiterite, quartz, and fluorite from the Takua Pit Thong deposit have been analyzed and the microthermometric results are presented in Appendix VII. There are no apparent differences in the data collected for solid-bearing or solid-

absent type 1 inclusions and none of the solids homogenized or changed shape at temperatures up to 500°C. Therefore the solid-bearing inclusions have not been classified as a different type. No microthermometric data from the Kow Daeng or other deposits in the area have been collected.

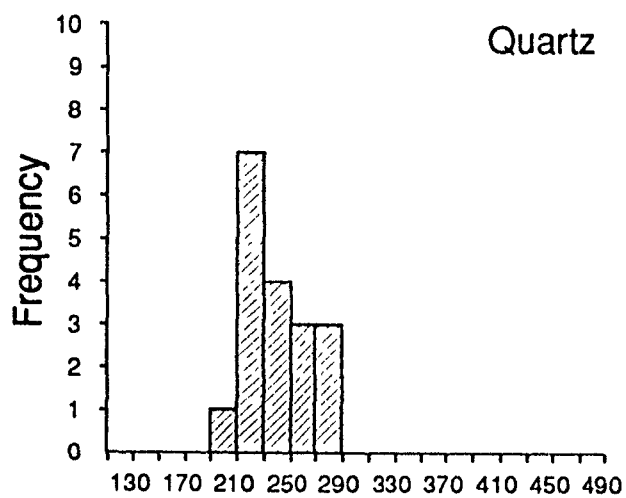
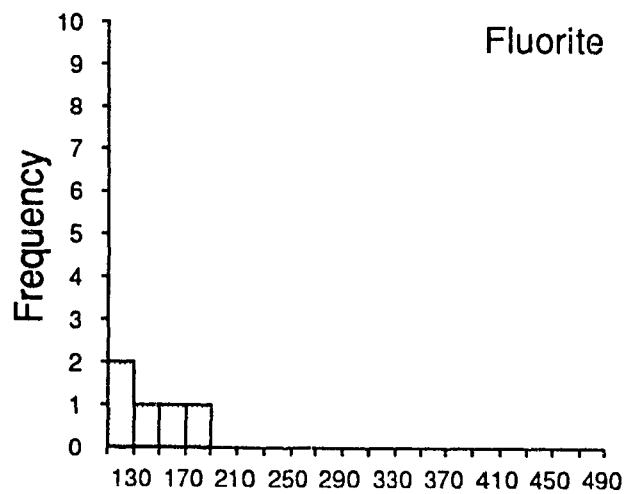
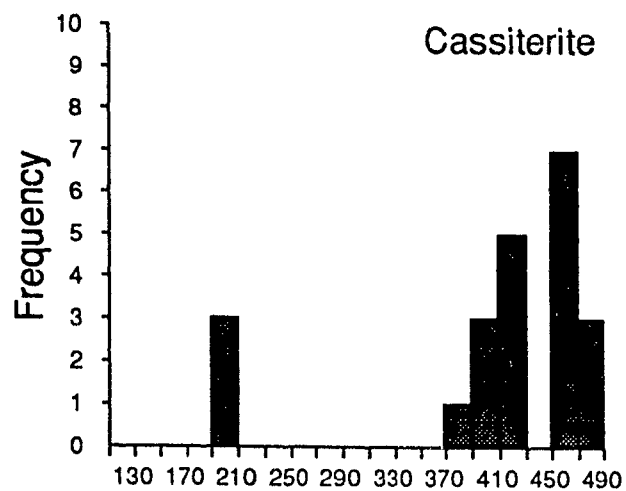
Aqueous Inclusions

Two types of aqueous inclusions, type 1 (L-V $\pm$ S) and type 4 (L-V-H), have been analyzed. The inclusions analyzed in cassiterite are primary, whereas those in quartz and fluorite are secondary or of unknown origin. Low eutectic and ice melting temperatures in type 1 inclusions, discussed below, indicate the presence of a salt(s) other than NaCl and/or KCl. In order to calculate salinities we have used the equation of Oakes *et al.* (1990) and assumed that for $T_{\text{melt ice}} > -20^\circ\text{C}$ the X_{NaCl} is 1.0 and for $T_{\text{melt ice}} < -20^\circ\text{C}$ the X_{NaCl} is 0.1, where X_{NaCl} is the weight ratio $\text{NaCl}/(\text{NaCl} + \text{CaCl}_2)$. The true X_{NaCl} is unknown but any error in the assumed value yields a maximum salinity error of approximately 1 wt % salt. All type 1 inclusions homogenize by vapour disappearance and for cassiterite these temperatures range from 195° to 475°C (Fig. 10 and Appendix VII).

The cassiterite data can be subdivided into high temperature ($>400^\circ\text{C}$) and "low" temperature ($<400^\circ\text{C}$) groups. The freezing and eutectic temperatures in the high temperature group were typically not visible (in one inclusion a T_e of -64°C was observed). Ice melting temperatures, bracketed by freezing, range from -5.8°

Figure 10. Histograms showing the distribution of homogenization temperatures of type 1 (L-V) inclusions from the Takua Pit Thong deposits.

L-V inclusions



Th (°C)

to -25°C (Appendix VII) and no hydrates are present at the last ice melt. These ice melting temperatures correspond to salinities of 7.0 wt % NaCl eq. and 23.1 wt % NaCl+CaCl₂ eq., respectively. Some of these inclusions contain carbonic vapour and their clathrates melt over a wide temperature range, -18.4° to $+7.9^{\circ}\text{C}$ (Appendix VII). The "low" temperature group is characterized by lower eutectic temperatures, approximately -74°C , and lower ice melting temperatures, -27° to -28°C , than the high temperature type 1 inclusions. They also differ by containing a hydrate which melts at $+14^{\circ}$ to $+16^{\circ}\text{C}$ (Appendix VII). The hydrate melting temperature was difficult to reproduce and may represent metastability (cf. Roedder, 1984). The salinities calculated from final ice melting temperatures in these inclusions is approximately 24 wt % NaCl+CaCl₂ eq. (Fig. 11).

Type 1 (secondary) inclusions in quartz typically have lower homogenization temperatures (Fig. 10) and lower salinities (Fig. 11) than those in cassiterite. Some inclusions melted by hydrate disappearance at approximately $+14^{\circ}\text{C}$, i.e., at a temperature similar to that of the "low" temperature group in cassiterite. Other type 1 inclusions contain carbonic vapour, and clathrates in these inclusions melted at temperatures between 8.9° and 10.6°C (Appendix VII). Type 1 inclusions in fluorite can be subdivided into two groups based on salinity (Fig. 11). The low salinity group have high eutectic temperatures, approximately -15°C , and melt by ice disappearance at -0.5° to -1.7°C , which corresponds to 0.9 to 2.2 wt % NaCl eq. These inclusions homogenize to liquid between 125° and 180°C (Fig. 10). The

second group of type 1 inclusions are moderately to highly saline. The ice melting temperature is approximately -32°C , which corresponds to a salinity of approximately 25 wt % $\text{NaCl}+\text{CaCl}_2$ eq. (Fig. 10).

Figure 12 is a plot of homogenization temperature versus salinity and shows that there are complex populations for type 1 inclusions at Takua Pit Thong; high and low homogenization temperatures with both high and low salinities. Primary fluid inclusions in cassiterite are high temperature and high salinity and may reflect a magmatic or modified (magmatic, metamorphic or meteoric) fluid. Secondary inclusions in quartz and are low temperature and low salinity and probably represent an influx of late meteoric waters. High temperature-low salinity and low temperature-high salinity fluid inclusions have unknown origin.

The data collected for type 4 (L-V-H) inclusions is dominantly from secondary inclusions in quartz and fluorite. One primary type 4 inclusion in cassiterite has also been analyzed. The hydrate and halite dissolution, and vapour disappearance temperatures for this inclusion are -19° , 173° (30.6 wt % NaCl eq.), and 469°C , respectively, (Appendix VII), suggesting that it is related to the high temperature type 1 inclusions in cassiterite.

Halite-bearing inclusions in quartz have eutectic temperatures that range from -60° to -70°C , ice melting temperatures of -30.2° to -36.5°C , hydrate melting temperatures of $+12.1^{\circ}$ to $+23.1^{\circ}\text{C}$, and halite disappearance temperatures ranging from 193° to 285°C (Appendix VII). The hydrate melting temperature was difficult

Figure 11. Histogram showing the distribution of salinities of type 1 (L-V) inclusions from the Takua Pit Thong deposits, calculated using the equations of Oakes et al. (1990).

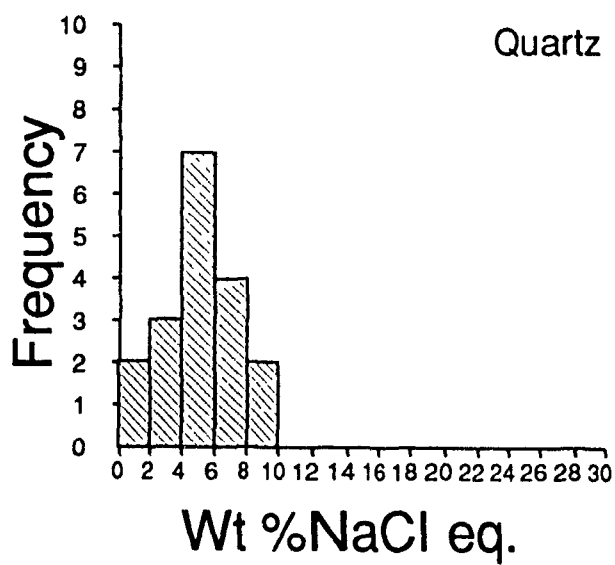
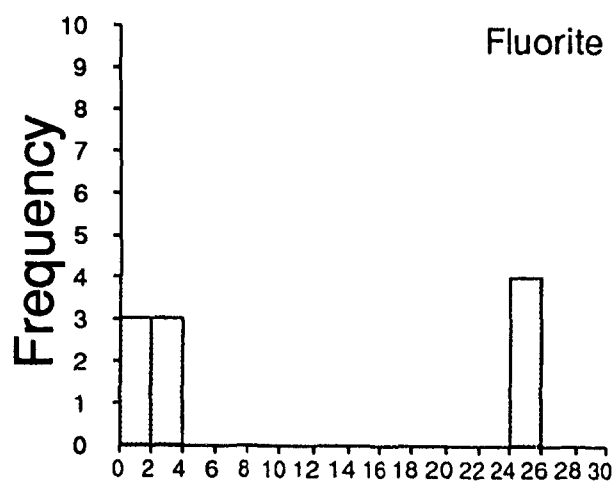
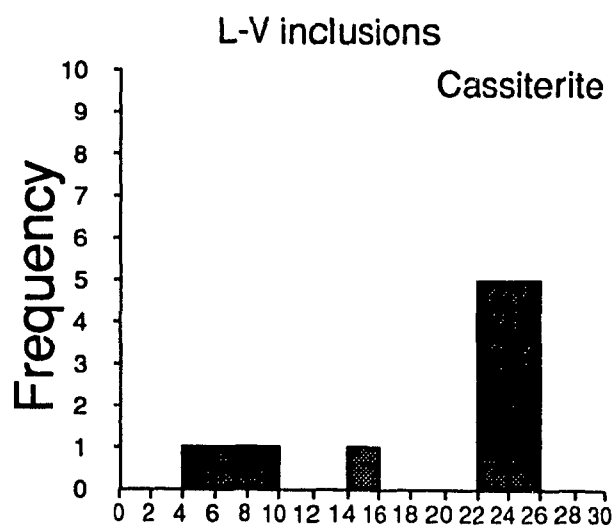
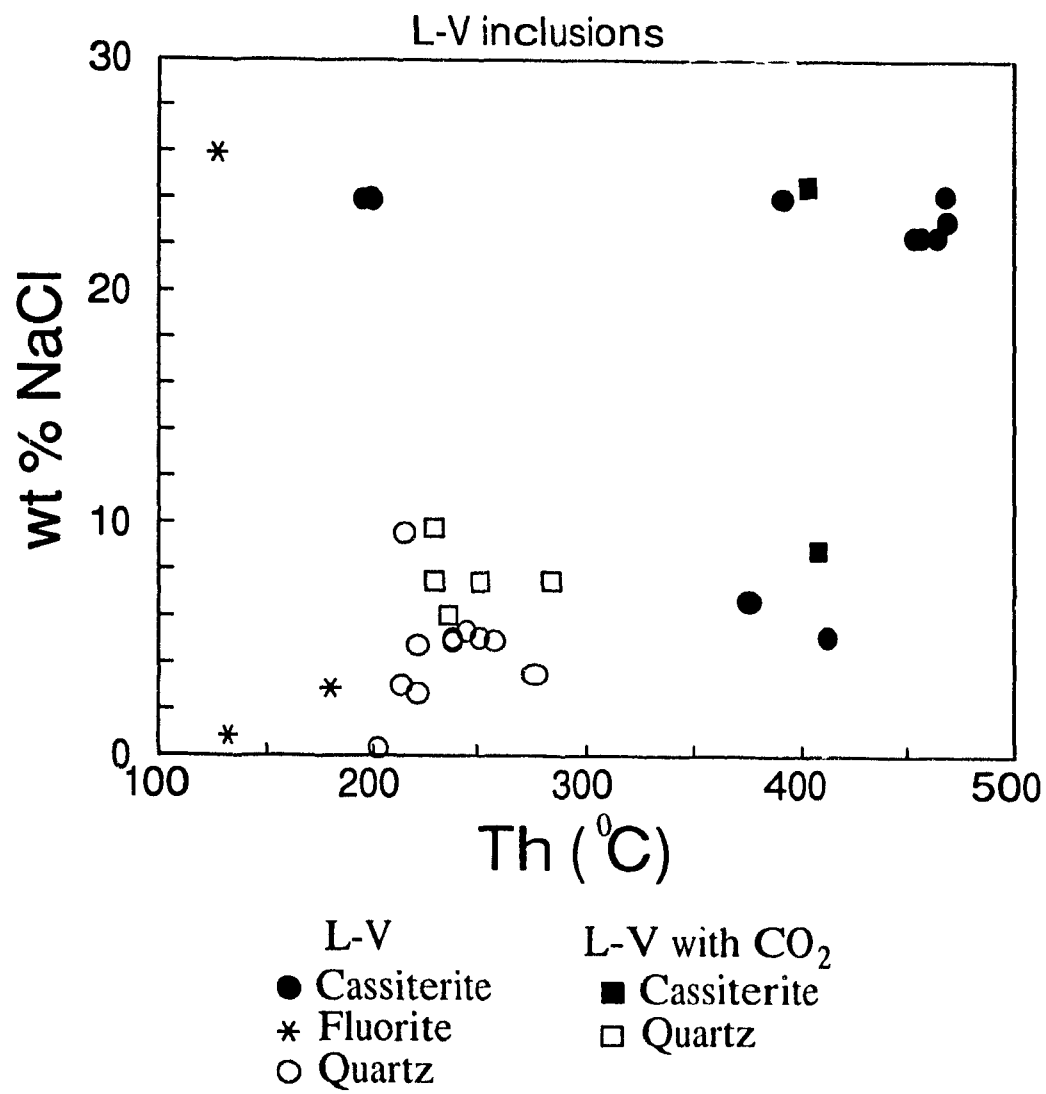


Figure 12. Salinity versus homogenization temperatures for type 1 (L-V) inclusions from the Takua Pit Thong deposits.

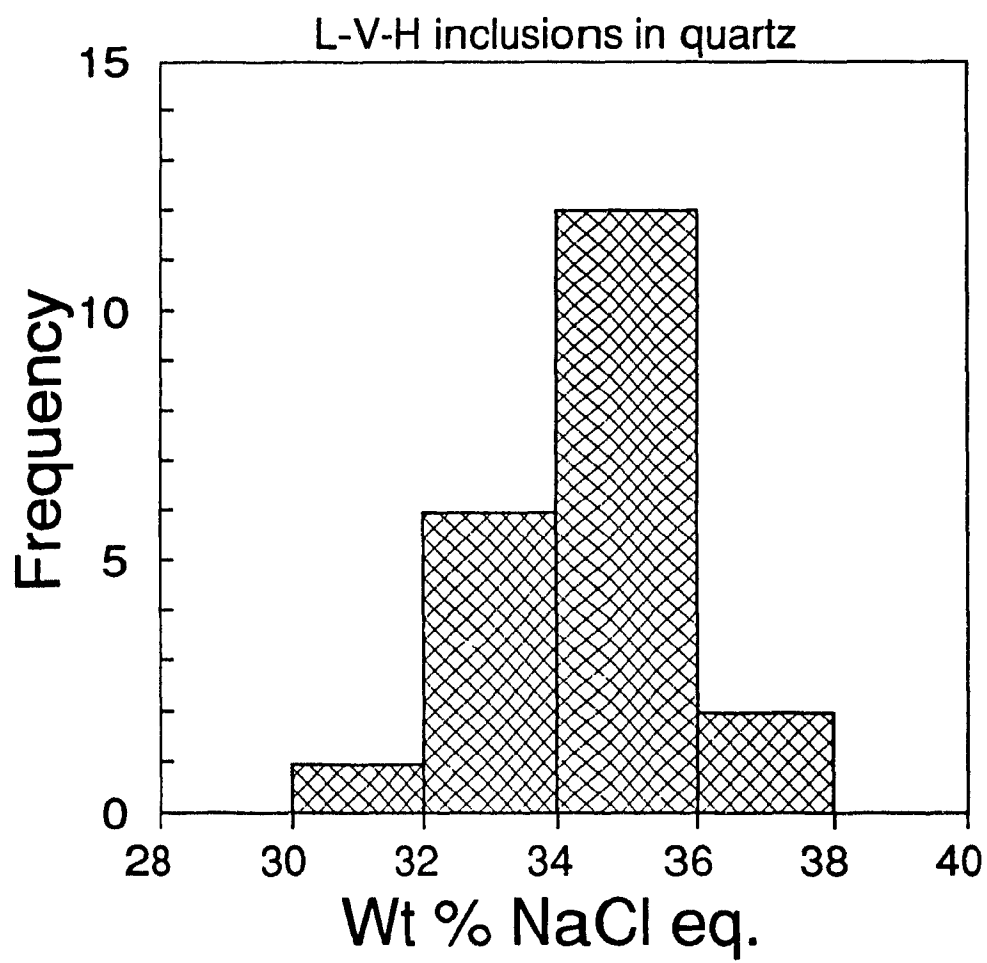


to reproduce and may represent metastability, similar to that discussed above. Vapour homogenization generally preceded halite dissolution and salinities were calculated as wt % NaCl eq. using the equation of Bischoff (1991). Figure 13 shows that the salinity of type 4 inclusions is relatively constant at 32 to 38 wt % NaCl eq. Some of these inclusions also contain a carbonic vapour, evident from double freezing and clathrate melting (-9.5° to -17.7°C); in one inclusion the melting of carbonic ice and carbonic homogenization were observed, at -65° and -41°C , respectively (Appendix VII).

Type 4 fluid inclusions in fluorite have a eutectic temperature of approximately -80°C and undergo melting in two different ways. Some inclusions melt in a manner similar to that of type 4 inclusions in quartz and cassiterite; ice melting at -20° to -32°C , followed by hydrate dissolution at $+13^{\circ}$ to $+14^{\circ}\text{C}$ (Appendix VII). Other type 4 inclusions melt by hydrate disappearance at -21° to -25°C . The behaviour of the latter inclusions is unusual; the eutectic is marked by the appearance of a birefringent hydrate rather than water. All type 4 inclusions in fluorite leaked at 90° to 120°C , prior to halite or vapour homogenization.

The very low eutectic and ice melting temperatures and the high temperatures of hydrate dissolution indicate high concentrations of salts other than NaCl or KCl. Considering that cassiterite mineralization is associated with fluorite and biotite alteration, and is related to the replacement of calcareous roof pendants, the fluid inclusion data is probably best explained by the presence of CaCl_2 or

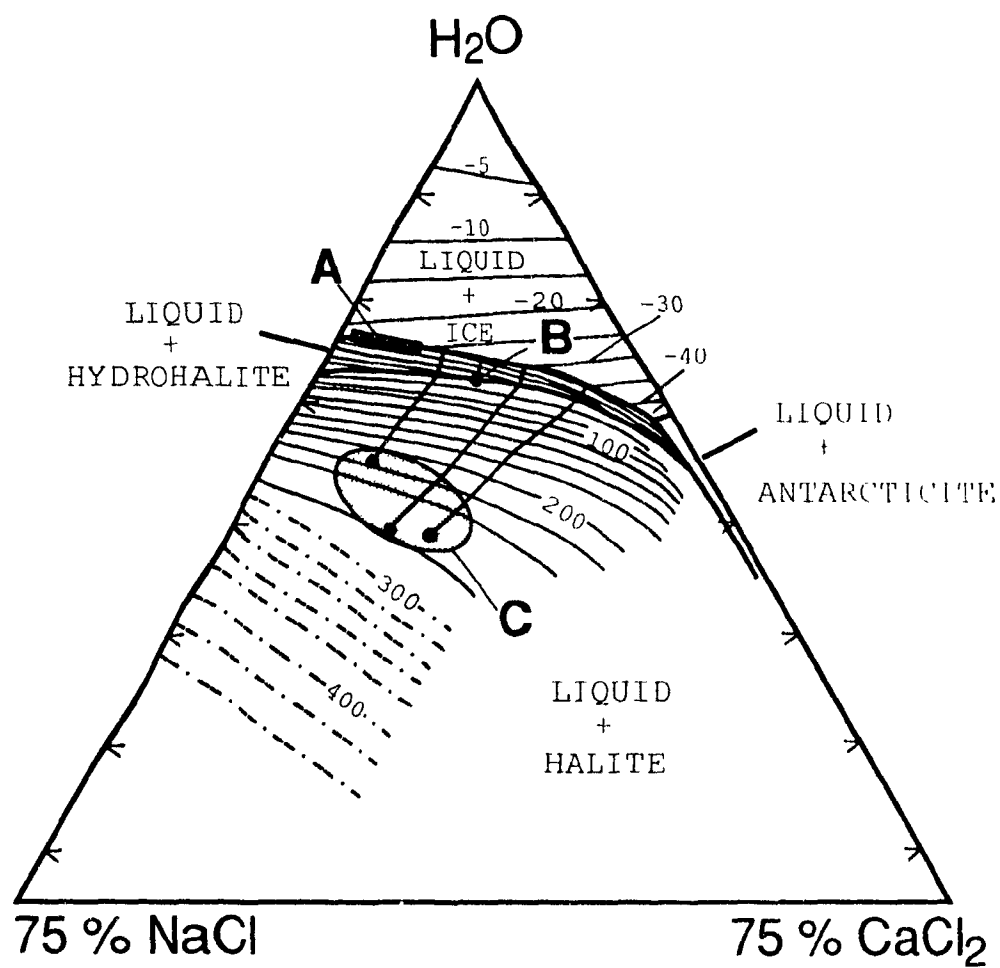
Figure 13. A histogram showing the distribution of salinities for type 4 halite-bearing inclusions in quartz from the Takua Pit Thong deposits.



FeCl_2 . These salts have been shown to be present in type 1 inclusions in cassiterite from scanning electron microscopy, discussed below. Phase equilibria in the H_2O - NaCl - FeCl_2 - CaCl_2 system are poorly understood, therefore the above microthermometric data is modelled assuming that the compositions are in the system H_2O - NaCl - CaCl_2 .

Figure 14 shows the compositions of various fluid inclusions interpreted in terms of weight percent H_2O , NaCl , and CaCl_2 . Area "A" represents the compositions of the high temperature type 1 inclusions in cassiterite which typically have final ice melting temperatures of -23° to -25°C . The minimum $\text{CaCl}_2/\text{NaCl}$ ratio is determined from the intersection of the ice melting temperature isotherms with the liquid + hydrohalite field and ranges from 0.05 to 0.4. Point "B" represents the compositions of the low temperature type 1 inclusions in cassiterite which have ice and hydrate melting temperatures of approximately -27°C and $+15^\circ\text{C}$, respectively. The $\text{CaCl}_2/\text{NaCl}$ ratio in these inclusions is roughly 0.95, the intersection of the $+15^\circ\text{C}$ isotherm in the hydrohalite + liquid stability field with the cooling path projected from the ice melting temperature. The $\text{CaCl}_2/\text{NaCl}$ ratio can also be estimated for type 4, halite-bearing, inclusions in quartz. These inclusions typically have final ice melting temperatures of -25.9° to -39°C and halite homogenization temperatures of 215° to 275°C . The projected fluid composition path of these inclusions as they are heated is perpendicular to the isotherms in the hydrohalite + liquid field after the ice has finally melted. The path

Figure 14. A plot of fluid inclusion compositions from the Takua Pit Thong deposits represented by the system $\text{NaCl-CaCl}_2\text{-H}_2\text{O}$ using the phase relationships of Oakes et al. (1990) and Williams-Jones and Samson (1990). Area 'A' represents L-V inclusions in cassiterite that melt by ice disappearance. Point 'B' represent L-V inclusions in cassiterite that melt by hydrate disappearance. Area 'C' represents halite-bearing inclusions in quartz.



then evolves along a tie-line to the NaCl apex as halite dissolves until the halite homogenization temperature. The latter point defines the fluid composition in the H_2O -NaCl- CaCl_2 system (Williams-Jones and Samson, 1990). The composition of type 4 inclusions in quartz is shown by area "C" on Figure 14; the path of three fluid inclusions with the widest range of compositions are shown, and have $\text{CaCl}_2/\text{NaCl}$ ratios ranging from 0.25 to 0.65.

Scanning electron microscopy

In an effort to establish what salts are present in type 1 inclusions in cassiterite, doubly polished wafers were heated to 400° and 650°C in two different experiments (owing to the high homogenization temperatures only a few inclusions decrepitated at 400°C). Some of the inclusions in these samples decrepitated and as a result a residue of non-volatile salts was precipitated on the wafer surface. The inclusion precipitates were then analyzed by an energy dispersive system on a scanning electron microscope.

For the inclusions that decrepitated at 650°C, Ca, Mn, and, in particular, Fe are the most common elements present in the surface residue. A representative spectrum is shown by Figure 15. Owing to the high decrepitation temperature, Cl, Na, and K were volatilized along with H_2O (cf. Haynes *et al.*, 1988). Native bismuth, scheelite, and wolframite were also interpreted to be in the precipitates. Figure 16 shows octahedral magnetite crystals, along with disk-shaped native Bi in

an inclusion precipitate. Spectra of the native Bi and of a wolframite-scheelite grain are also shown in Figure 16. It is not clear whether the W and Bi represent residues from volatilized fluid, are daughter/trapped minerals, or are solid inclusions. Opaques within fluid inclusions are rare but solid inclusions in cassiterite are common. However, the fact that several decrepitates contain W and/or Bi suggests that high concentrations of these elements are present as aqueous complexes in fluid inclusions.

Precipitates from inclusions in cassiterite that decrepitated at 400°C contain Na, K, and Cl in addition to Fe, Mn, and Ca. A representative spectrum of an inclusion precipitate is shown in Figure 17. Note the relative peak heights on this figure, which indicate that Na is not the predominant cation, rather that these are complex fluids composed of Fe-Ca-Mn-Na-K chlorides (although some Na may have been volatilized). Tungsten was detected in some precipitates, as was the case in the high temperature decrepitates, discussed above.

Type 1 inclusions hosted by cassiterite commonly contain a birefringent solid. In an attempt to identify the solid, cassiterite grains were immersed in liquid nitrogen and then broken. The solids in opened fluid inclusions were then analyzed using the energy dispersive system as described above. The most common mineral observed remains unidentified. This mineral has Si, Fe, and Cl peaks and possibly is either a beryllium or boron mineral (Fig. 18). Figure 18 also shows a euhedral quartz crystal in one inclusion and magnetite and wolframite in another inclusion.

Figure 15. Representative spectrum of cassiterite-hosted fluid inclusion precipitates from Takua Pit Thong, decrepitated at $>600^{\circ}\text{C}$.

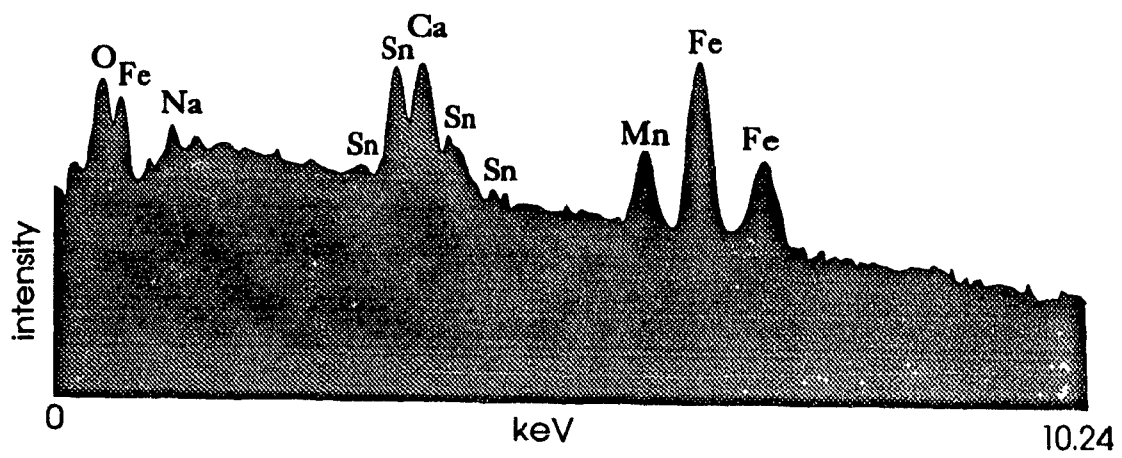
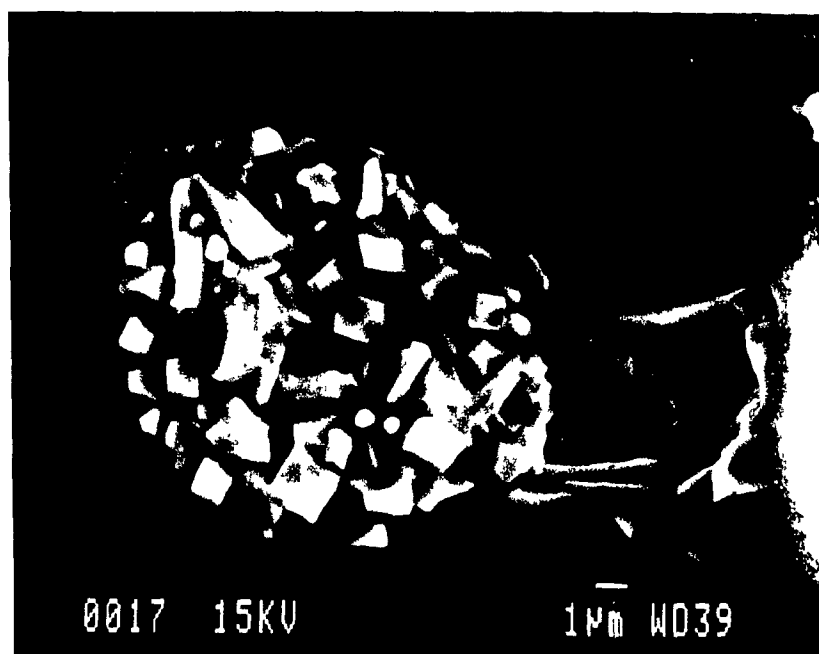
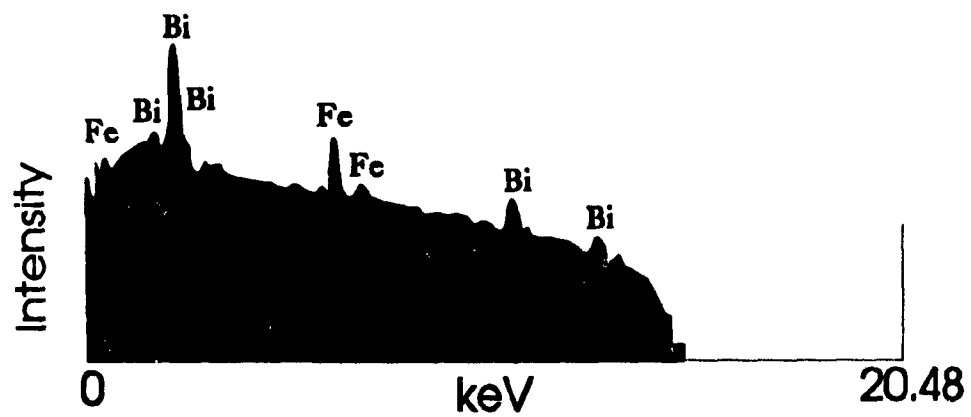


Figure 16. a) Scanning electron microscope image of a cassiterite-hosted fluid inclusion precipitate decrepitated at $>600^{\circ}\text{C}$, with octahedral magnetite and disk-shaped Bi; b) spectrum of the Bi precipitate; c) spectrum of a wolframite-scheelite precipitate; all from Takua Pit Thong.

a



b



c

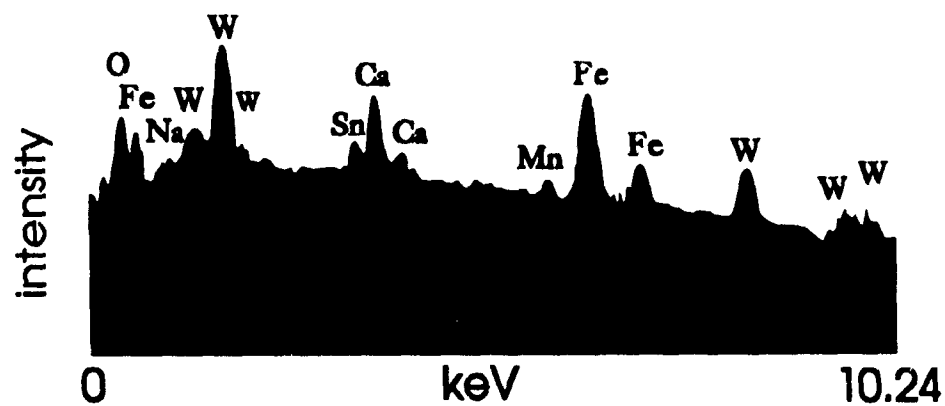


Figure 17. Representative spectrum of an inclusion precipitate decrepitated at 400°C.

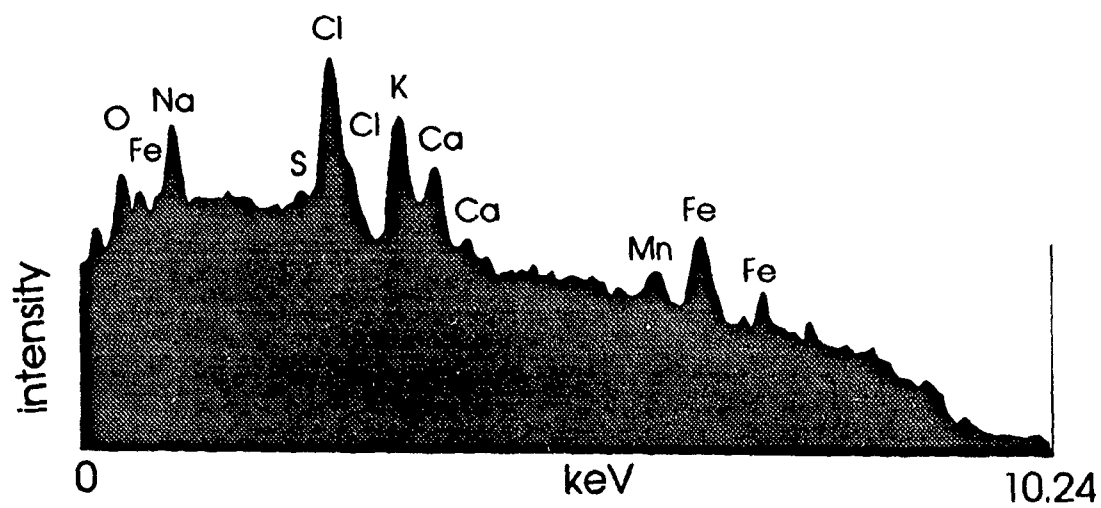


Figure 18. Scanning electron microscope images of minerals in opened fluid inclusions in cassiterite from Takua Pit Thong.

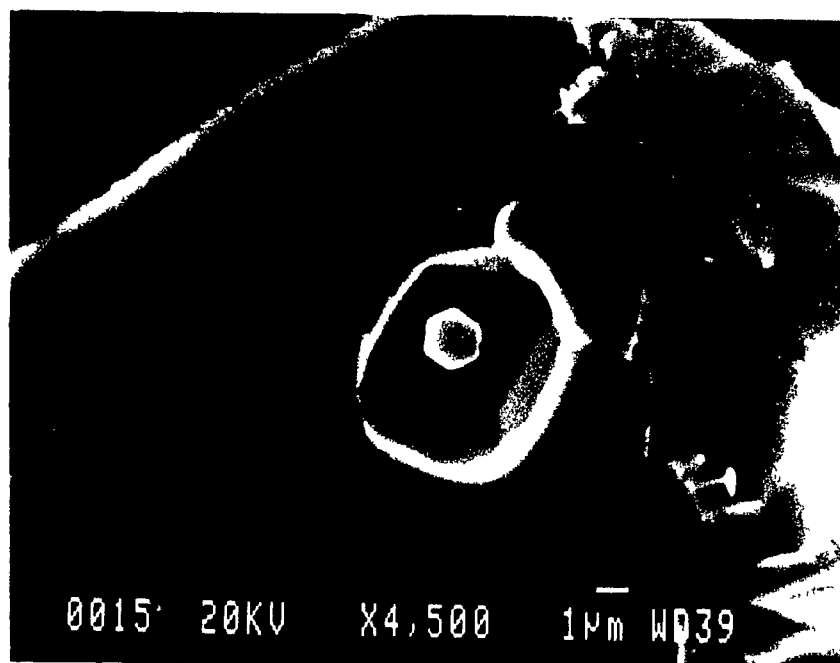
a) quartz

b) magnetite (tabular) and wolframite (rod-shaped)

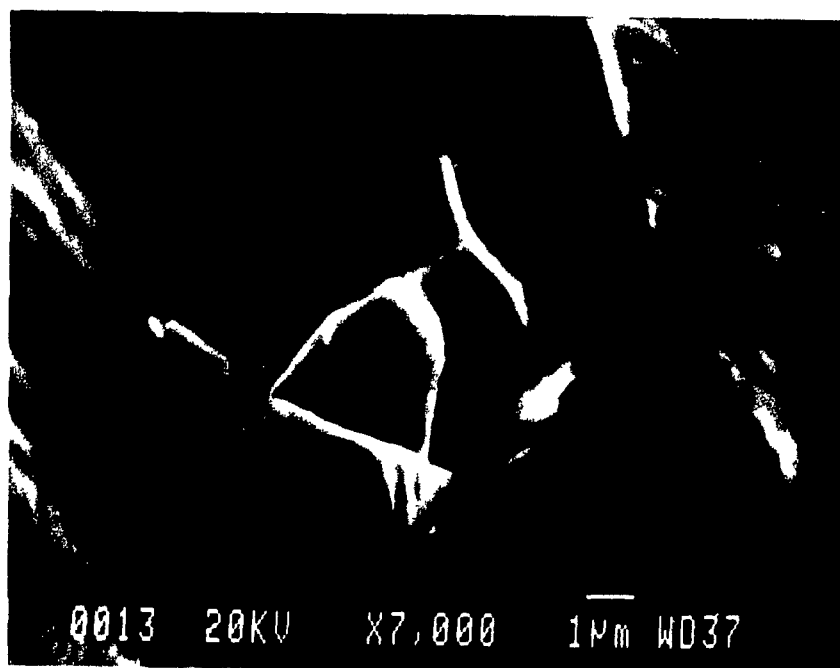
c) unknown Si-Fe-Cl mineral

d) spectrum of the unknown Si-Fe-Cl mineral

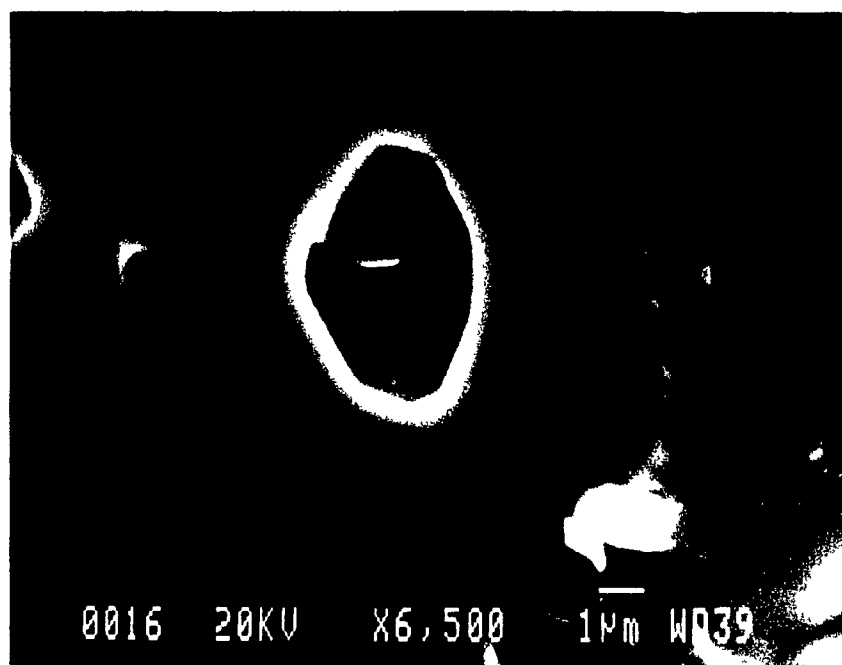
a



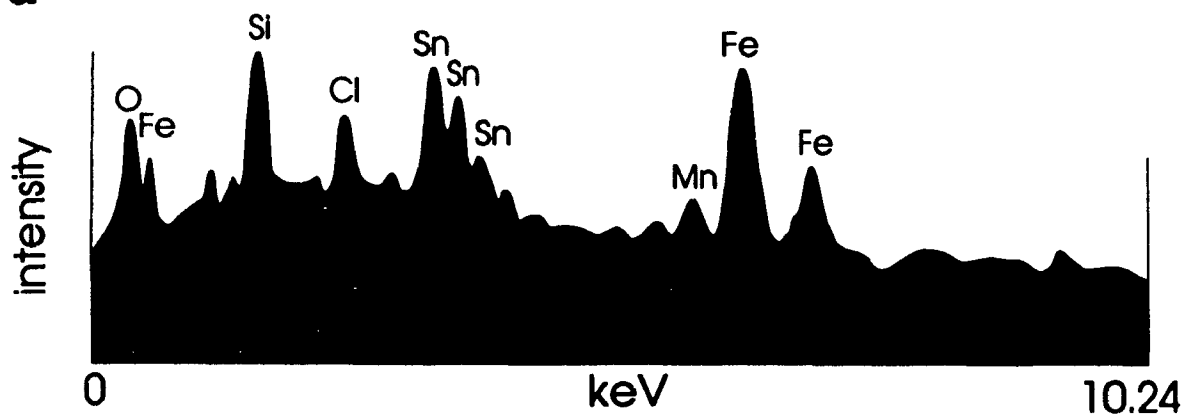
b



c



d



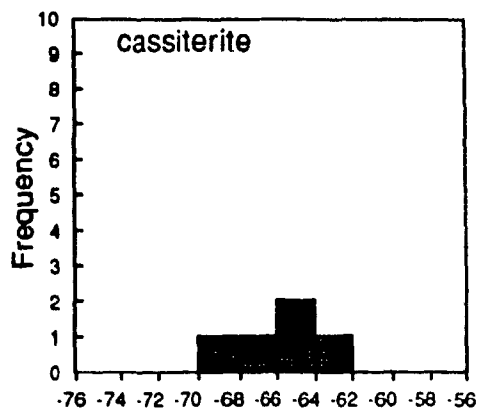
Aqueous-carbonic and Carbonic Inclusions

Aqueous-carbonic fluid inclusions (type 3, L-L-V) are present in secondary planes in quartz and fluorite where they are generally associated with vapour-rich carbonic inclusions. They are not present in cassiterite. Of the inclusions of this type analyzed, ice was observed to melt only in fluorite, at a temperature ranging from -14.7° to -14.9°C , which corresponds to a salinity of 18 wt % NaCl eq. (Appendix VII). By contrast, clathrates were observed to melt only in quartz, typically at a temperature of approximately $+10^{\circ}\text{C}$, which suggests low salinities. The carbonic ice in type 3 inclusions in both quartz and fluorite typically melts at temperatures of -57° to -58°C , indicating that it is CO_2 -rich (Fig. 19). The homogenization temperatures of the carbonic phases are however, highly variable, -20.6° to $+27.0^{\circ}\text{C}$ in quartz and $+1.9^{\circ}$ to $+19.1^{\circ}\text{C}$ in fluorite (Appendix VII).

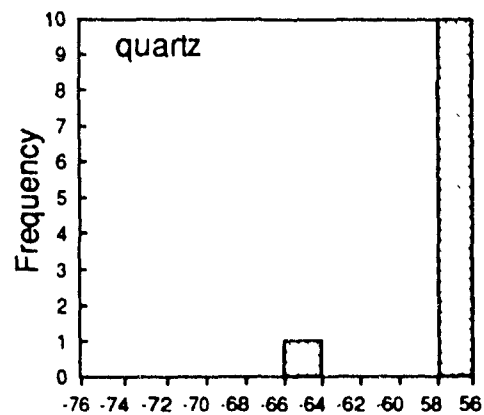
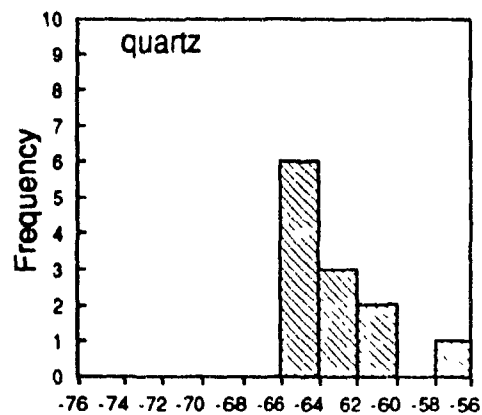
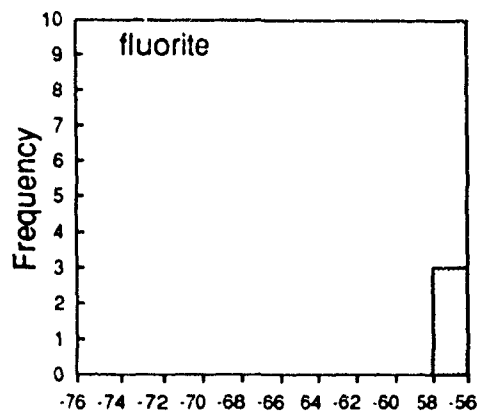
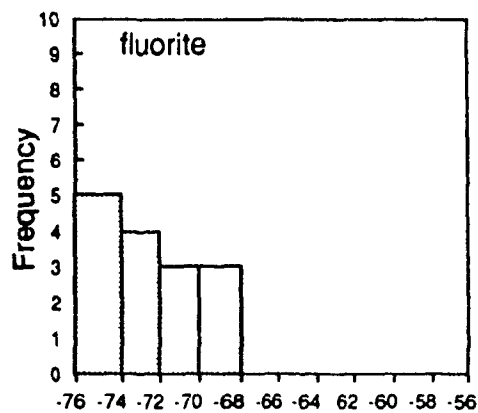
Carbonic and carbonic-aqueous inclusions are much more common than aqueous-carbonic inclusions. Primary vapour-only inclusions are the dominant type 2 inclusion in cassiterite whereas secondary vapour-rich (V-L) inclusions are the dominant type in quartz and fluorite. Clathrate melting was recorded in some carbonic inclusions (only a single visible vapour phase) indicating that these inclusions are carbonic-aqueous. Melting of the carbonic ice occurred at temperatures ranging from -63° to -70°C in cassiterite, -60° to -65°C in quartz, and -68° to -76°C in fluorite (Fig. 19). The homogenization of the carbonic liquid and

Figure 19. Histograms showing the distribution of the carbonic melting temperatures of type 2 (V and V-L) and type 3 (L-L-V) inclusions from the Takua Pit Thong deposits.

V-L-L inclusions



L-L-V inclusions



Tmelt CO₂

Tmelt CO₂

vapour could not be observed in cassiterite owing either to the darkness of these inclusions or their low density. Type 2 inclusions in quartz and fluorite display a wide range of carbonic phase homogenization and clathrate melting temperatures. Figure 20 compares carbonic ice melting and homogenization temperatures. There is an apparent correlation between these two parameters, i.e., lower melting temperatures correspond to lower homogenization temperatures. This trend suggests that there is a continuous range of carbonic compositions. Although there is a wide range of homogenization temperatures, inclusions along a single secondary plane have consistent ($\pm 0.5^\circ\text{C}$) homogenization temperatures. Figure 21 is a plot of clathrate versus carbonic ice melting temperatures. There is significant variation of both parameters which suggests that the compositions of both the aqueous and carbonic portions of the inclusions are highly variable. The X_{CH_4} of the type 2 inclusions in fluorite cannot be determined with the existing phase equilibria data, but, given the low melting and homogenization temperatures, must have a value of greater than 0.7 (Arai *et al.*, 1971). The X_{CH_4} of vapour-rich inclusions in quartz ranges from approximately 0.2 to 0.6. All the aqueous-carbonic (L-L-V) inclusions in fluorite have X_{CH_4} values of < 0.05 and the single analysis from quartz is approximately 0.5.

FLUID EVOLUTION

The primary type 1 inclusions hosted by cassiterite most probably represent

Figure 20. Carbonic liquid-vapour homogenization temperatures versus carbonic melting temperatures of type 2 (V and V-L) inclusions from the Takua Pit Thong deposits.

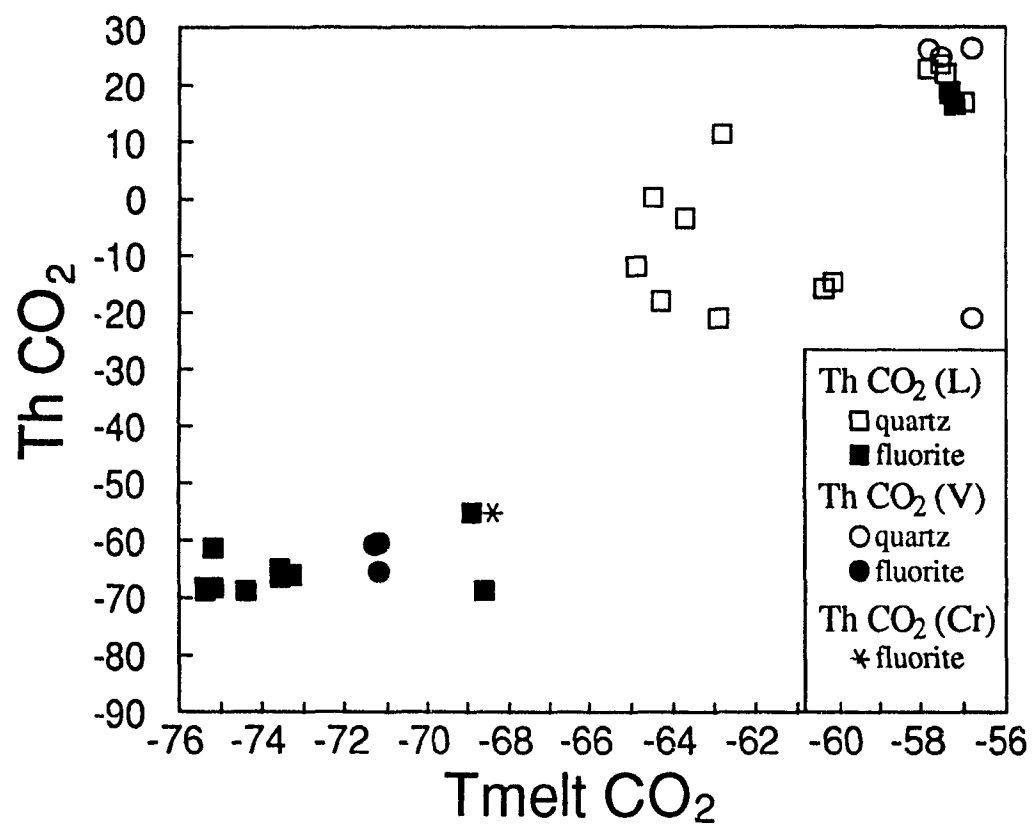
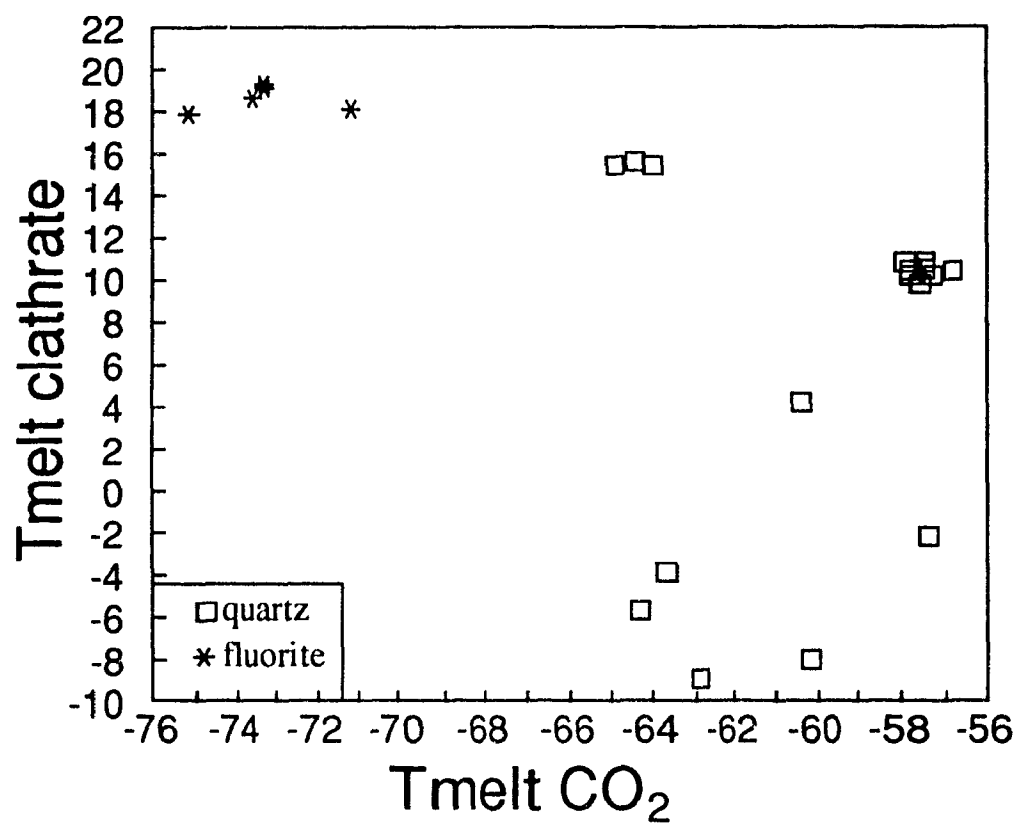


Figure 21. Clathrate melting temperatures versus carbonic ice melting temperatures of V-L inclusions from the Takua Pit Thong deposits.



1 trapped mineralizing fluids. The high iron content of these inclusions is consistent with the iron-enrichment observed in the silicate alteration associated with cassiterite. However, it is also apparent that these fluid inclusions are very complex, containing high concentrations of Fe, Mn, and Ca chlorides, and probably also significant W and Bi. They are therefore probably of magmatic or an evolved origin (external fluid compositions modified by exchange with the wallrock). It is interesting that the primary fluid inclusions at Takua Pit Thong are compositionally similar to the pseudosecondary high salinity inclusions in cassiterite at Nong Sua. Future analyses of the primary inclusions in the cassiterite at Kow Daeng, that predates Takua Pit Thong cassiterite, will have bearing on the interpretation of the fluid evolution at Takua Pit Thong, i.e., if they are of lower salinity the fluid evolution could be similar to that at Nong Sua, whereas, if their composition is similar to that at Takua Pit Thong, a magmatic origin is likely for the primary aqueous inclusions in cassiterite at both Takua Pit Thong and Kow Daeng.

The fact that halite-bearing and low salinity L-V inclusions are typically secondary indicates that they were trapped after the primary inclusions in cassiterite. The most probable source of the hypersaline fluids is formational or metamorphic waters, as at Nong Sua and Pilok, but an origin by boiling of the moderately saline fluids (primary fluid inclusions in cassiterite) is also a possibility. The low temperature-salinity fluid inclusions most probably represent an influx of late meteoric waters (as at Pilok).

The wide range of carbonic compositions suggests that the carbonic and aqueous-carbonic fluid inclusions do not have a magmatic source. The depletion of carbon in the metasediments around the ore deposits, particularly in the calc-silicate hornfels, indicates that the most probable source of carbonic and aqueous-carbonic fluids was the metasedimentary rocks. The fact that carbonic (vapour-rich) fluid inclusions are enriched in CH_4 and that aqueous-carbonic (liquid-rich) inclusions are enriched in CO_2 , suggests that they coexisted as immiscible fluids (Ramboz *et al.*, 1982). This does not necessarily indicate that one fluid became immiscible (i.e. boiled); it is also possible that vapour-rich and liquid-rich fluids were both produced by devolatilization reactions in the metasediments. If the carbonic fluids are not magmatic, then some pressure correction is required for the primary inclusions in cassiterite (they were not boiling), and thus the temperature of mineralization was probably $\geq 500^\circ\text{C}$.

CONCLUSIONS

Although the data on the deposits in the Takua Pit Thong area are preliminary some tentative conclusions can be drawn. The evolution of mineralization at Takua Pit Thong and Kow Daeng are constrained by the timing of different alteration-mineralization events. From these constraints it is apparent that there was an early Sn-Nb-Ta event at Kow Daeng followed by a Sn-Cu-Zn-Be-F-S event at Takua Pit Thong. Since pegmatite and skarn mineralization at Kow

Daeng preceded sulphide replacement mineralization at Takua Pit Thong, it is probable that the temperature of cassiterite deposition at Kow Daeng was at least as high as at Takua Pit Thong. The early "pegmatite" and skarn mineralization may have formed at the early vapour-saturated stage of the source granite, whereas the later sulphide replacement mineralization was probably deposited by very late magmatic fluids or from fluids that coexisted with the source granite at early sub-solidus conditions. Both early and late fluids are similar in at least one respect. Iron enrichment trends in the silicate minerals associated with cassiterite are observed in both deposits. At Takua Pit Thong, cassiterite is interpreted to have precipitated at $\geq 500^{\circ}\text{C}$ from a moderately high salinity aqueous fluid. The carbonic fluids at Takua Pit Thong are interpreted to have had a metamorphic source and the presence of liquid-rich (and CO_2 -rich) and vapour-rich (and CH_4 -rich) fluid inclusions suggests that two immiscible aqueous-carbonic fluids coexisted.

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CHAPTER 7

THE ORIGIN OF CONTRASTING MINERALIZATION STYLES

INTRODUCTION

One of the principal objectives of this study was to investigate the mechanisms by which Sn-W deposits form. Three mechanisms: direct crystallization from melt (magmatic); precipitation from magmatic aqueous fluids (orthomagmatic) and; hydrothermal remobilization were proposed. The Nong Sua and Pilok deposits were selected as representatives of magmatic-dominated and hydrothermal-dominated systems, respectively. The study of aplite-pegmatite-hosted mineralization at Nong Sua demonstrated that magmatic and orthomagmatic processes controlled cassiterite-wolframite deposition. The vein-stockwork mineralization at Pilok, by contrast, probably formed as a result of hydrothermal remobilization of Sn-W from the host granite. In this chapter the two deposits are contrasted and, where possible, compared to other deposits in western Thailand (including Takua Pit Thong, for which preliminary data are provided in Chapter 6), in an effort to explain the Sn-W metallogeny of the western Thai province. This province is then compared to other Sn-W belts to determine whether the conclusions of this study have more global application.

CHARACTERISTICS OF THE WESTERN THAI Sn-W PROVINCE

Magmatic evolution and the depth of emplacement

Tin was concentrated by magmatic processes (differentiation) at both Nong Sua and Pilok. In the case of Nong Sua, the tin concentration in the melt was high enough for cassiterite to be saturated (Chapter 2). There is no evidence to support direct crystallization of cassiterite at Pilok, but the distribution of tin in the granitic rocks suggests magmatic enrichment followed by hydrothermal redistribution (Lehmann and Mahawat, 1989). It is significant that magmatic cassiterite at Nong Sua crystallized from melt at >90% fractionation (Chapter 2), whereas pegmatite pods at Pilok are typically unmineralized. This may simply reflect the small size of the Pilok pegmatite pods, i.e., given the same magmatic evolution as Nong Sua, the amount of cassiterite crystallized in these pods would have been negligible. An alternative explanation for the absence of cassiterite in the pegmatite pods at Pilok is that they were emplaced at shallower depth (cf. Černý, 1982). The restriction of the cassiterite mineralization to veins-stockworks is also consistent with the observation of Varlamoff (1978), based on occurrences of cassiterite mineralization in various metallogenic provinces that, at depth, tin tends to be concentrated in pegmatites and, at shallower levels, in vein-stockwork deposits.

Lithostatic pressure during emplacement of the Nong Sua pegmatite is estimated to have been ≤ 3.8 kbar, corresponding to a depth of >10 km (Chapter 4). The

lithostatic pressure at Pilok is constrained by the occurrence of two alkali feldspar granite to have been >1.5 kbar and from andalusite in the contact metamorphic aureole to have been <3.8 kbar. Minimum pressures, based on fluid inclusion data (probably less than lithostatic), suggest that the mineralization was deposited at >0.35 to 1.0 kbar (Chapter 5). It therefore follows that the depth of emplacement of the Pilok granites was probably somewhat higher than at Nong Sua, i.e., at or above 10 km. These depth estimates are somewhat deeper than those of Varlamoff (1978), but the relative level of emplacement of pegmatite versus stockwork deposits is similar. The major Sr-W ($\pm$ Ta, Nb) deposits and mining camps of the western Thai belt, most of which were visited by the author, are shown in Figure 1 and Table 1. Pegmatite and stockwork deposits are present over the length of the belt, but the relative importance of the deposit types varies amongst mining camps. This suggests that the deposits were emplaced at the same general level, i.e., within a few vertical kilometres of each other, but that local variations in depth may have been sufficient to change the mineralization style.

The west-central region (Pilok to Hub Kapong; Fig. 1) is characterized by vein and stockwork deposits that are commonly boron-rich. Pegmatite deposits are rare, but minor Sn-Ta-Nb pegmatites that locally contain lepidolite are present at Som Porn (Appendix VIII). The granites in the Suan Pung area (also west-central) may be relatively fluorine-rich since fluorite is abundant at Takua Pit Thong (Chapter 6) and Chat Chai (Appendix VIII), and granite-associated fluorite deposits occur

Figure 1. Location of some of the major primary Sn-W deposits and strike-slip faults of the western Thai province. The abbreviations are TPFZ Three Pagodas fault zone, RFZ Ranong fault zone and KMFZ Klong Marui fault zone.

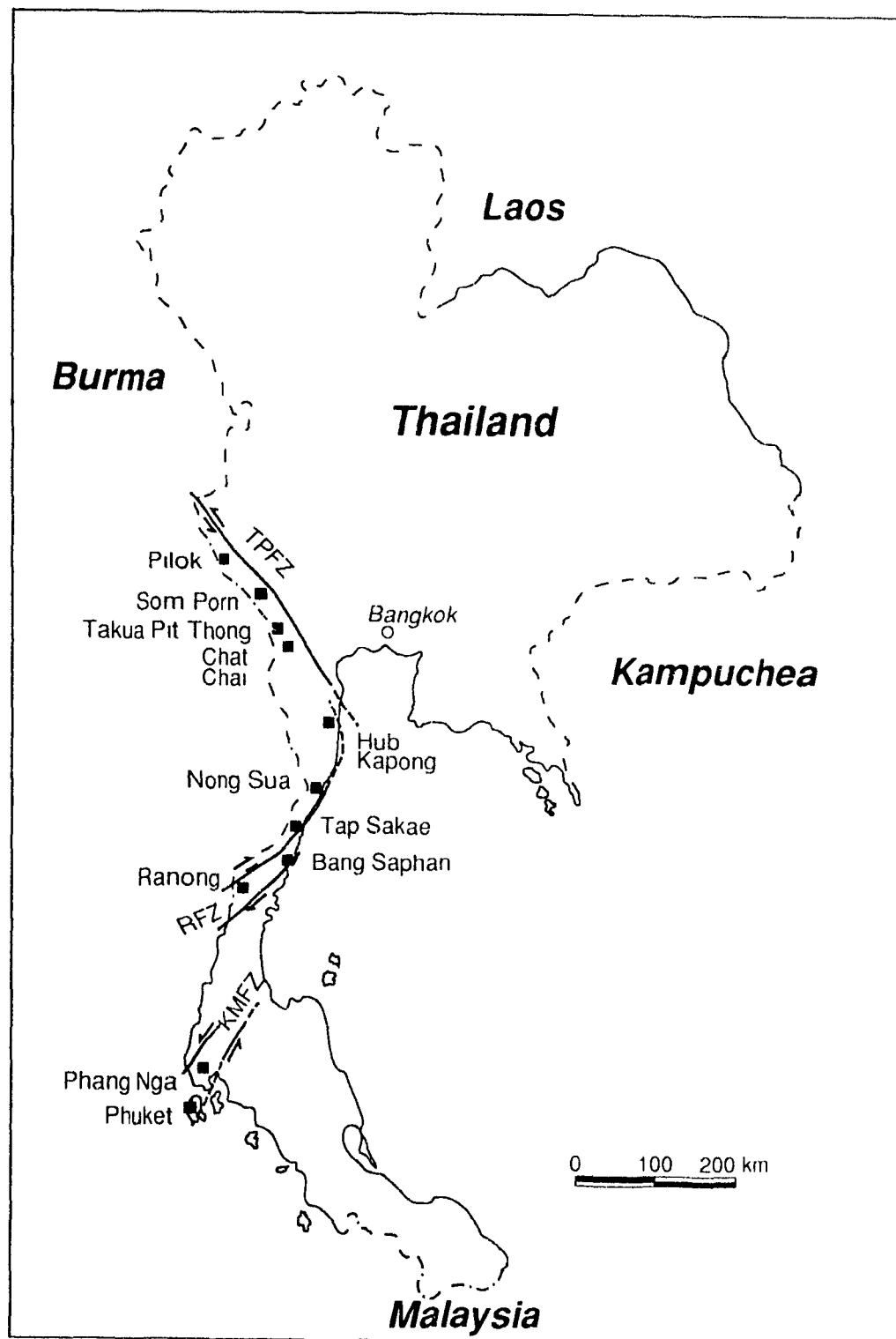


TABLE 1
Some of the Sn-W Deposits in Western Thailand

Deposit	Mineralization	Comments	Reference
Pilok	stockwork-vein	boron enriched	this study, Chapter 5
Som Porn	stockwork, pegmatite disseminated	minor lepidolite	this study, Appendix VIII
Takua Pit Thong	sulphide replacement, "pegmatite", skarn, breccia	fluorine enriched	this study, Chapter 6
Chat Chai	stockwork	fluorine enriched	this study, Appendix VIII
Ban Tha Lao (Hub Kapong)	vein	boron enriched	this study, Appendix VIII
Nong Sua	pegmatite	boron enriched	this study, Chapters 2,3 & 4
Thai Prasit (Tap Sakae)	"pegmatite"	quartz-muscovite $\pm$ feldspar dykes	this study, Appendix VIII
Visate (Tap Sakae)	pegmatite	boron enriched	Charusiri (1989)
Mae Toranee (Bang Saphan)	pegmatite	K-feldspar rich	Charusiri (1989)
Haad Som Pan (Ranong)	disseminated, stockwork	boron enriched	Charusiri (1989)
Khao Chai (Ranong)	vein, stockwork	boron enriched	Charusiri (1989)
Nok Hook (Phang Nga)	disseminated	boron enriched	Charusiri (1989), Pollard <u>et al.</u> (1991)
Juti (Phang Nga)	pegmatite	boron enriched	Charusiri (1989)
Ngan Tawee (Phang Nga)	pegmatite	boron enriched	Charusiri (1989)
Bang-I-Tum (Phang Nga)	pegmatite	boron enriched, minor lithium	Garson <u>et al.</u> (1975) Charusiri (1989)
Reung Kiet (Pnang Nga)	pegmatite	lithium enriched	Garson <u>et al.</u> (1975) Charusiri (1989)

TABLE 1 (continued)

Deposit	Mineralization	Comments	Reference
Ban Nguan (Phuket)	pegmatite	lithium enriched	Garson <i>et al.</i> (1975) Charusiri (1989)
Tor Soong (Phuket)	pegmatite	boron enriched	Charusiri (1989)
Chao Fa	pegmatite	boron enriched	Charusiri (1989)
Khao Kow	disseminated, stockwork	boron enriched	Charusiri (1989)

elsewhere in this area (Fig. 1 in Chapter 6). In the south-central region the tin deposits are typically pegmatite-hosted (Nong Sua; Tap Sakae and Bang Saphan areas; Fig. 1) and most of these pegmatites are boron-rich (Table 1). If Nong Sua is representative of the pegmatites, they are anomalously low in fluorine (~500 ppm in pegmatite versus ~1500 ppm for the regional granites; Appendix I). In the Ranong area disseminated, stockwork and vein-hosted cassiterite-wolframite deposits are reported to be associated with tourmaline-rich leucocratic granite (Charusiri, 1989). The Phang Nga and Phuket areas are somewhat different from the rest of the western Thai belt in containing a number of lepidolite-bearing pegmatites that host cassiterite (Garson et al., 1975; Gocht and Pluhar, 1982; Charusiri, 1989). However, abundant cassiterite is also present in tourmaline-muscovite pegmatites (i.e., similar to Nong Sua) and in stockworks and disseminations in granite (Garson et al., 1975; Pollard et al., 1991). The B, F, and Li contents of each mining camp are different. The variable amounts of these elements could have given rise to differences in the solidus temperatures of the associated plutons and consequently, differences in the levels of emplacement (cf. Manning and Pichavant, 1988).

The granites at Phuket and Pilok have normative compositions that are close to those of the 1 kbar hapiogranite minimum composition (Putthapiban, 1984; Lehmann and Mahawat, 1989). The composition of aplite at Nong Sua is similar to that of the granites at Phuket and Pilok in this respect (Chapter 2). However,

considering that the granites may not have been water saturated and that the haplogranite minimum does not include the affects of Ca, Fe, B, F, Li and excess Al, this method is not reliable for estimating pressure. Furthermore, no volcanic rocks are present in the western Thai belt, as might be expected for magmas emplaced at a depth of ≤ 4 km. All the major mining camps lie close to major strike-slip faults (Fig. 1), hence differential uplift may have exposed different crustal levels. The data from Nong Sua and Pilok suggest that the depth of granite emplacement was approximately 10 km ($\pm \sim 3$ km?), discussed above. Some support for this depth estimate is provided by the contact metamorphic assemblage at Phuket, biotite-andalusite-cordierite that locally includes sillimanite (Garson et al., 1975). It is not clear whether the muscovite and chlorite in these rocks also constitute part of this assemblage, but if so, the pressure of contact metamorphism is restricted to have been approximately 3 kbar (van Bosse and Williams-Jones, 1988).

Fluid evolution

Table 2 summarizes the fluid inclusion and stable isotopic data, from the Nong Sua and Pilok deposits. The fluid inclusion data for Takua Pit Thong are preliminary and thus will be discussed later, along with data from other deposits in western Thailand. It is apparent that fluids of similar composition are present at Nong Sua and Pilok. This is not surprising as both deposits are associated with

TABLE 2
Comparison of Fluid Inclusion and Stable Isotope Data

Incl. Isotope	Nong Sua	Pilok
aqueous (L)	absent	late stage, <1 wt% NaCl eq.
aqueous (L-V)	early orthomagmatic $T_h \sim 300^\circ\text{C}$, ~3 wt% NaCl eq., later external $T_h \sim 400^\circ\text{C}$, ~20 wt% NaCl eq.	older $T_h \sim 300^\circ\text{C}$, 10 wt % NaCl eq., younger $T_h \sim 200^\circ\text{C}$, 2 wt % NaCl eq.
aqueous (L-V-H)	late, 33 to 36 wt% NaCl eq., significant Ca, Fe, Mn, K	late, 32 to 35 wt% NaCl eq., significant bivalent cations probable
aqueous-carbonic (L-L-V)	indeterminate origin, X_{CH_4} 0.0 to 0.15, T_h 300° to 400°C , 0 to 20 wt% NaCl eq.	indeterminate origin, X_{CH_4} 0.0 to 0.2, T_h 280° to 320°C , <6 wt% NaCl eq.
aqueous-carbonic (V-L-L)	indeterminate origin, X_{CH_4} 0.0 to 0.25, T_h 300° to 400°C , 0 to 15 wt% NaCl	indeterminate origin, X_{CH_4} 0.0 to 0.3, T_h 280° to 400°C , <6 wt% NaCl
$\delta^{18}\text{O}$ qtz	12.3 to 12.7 per mil magmatic, 11.5 to 12.9 per mil hydrothermal	10.8 and 10.9 per mil pegmatite, 9.1 and 9.4 per mil wolf-peg, 8.9 to 12.4 per mil greisen
$\delta^{18}\text{O}$ mu	9.8 to 10.2 per mil magmatic, 8.0 to 9.4 per mil hydrothermal	8.2 and 8.9 per mil pegmatite, 7.1 and 9.3 per mil wolf-peg, 6.1 to 9.0 per mil greisen
$\delta^{18}\text{O}$ cas	4.9 to 6.4 per mil disseminated and vein	3.4 to 3.7 per mil greisen and vein
δD mu	-91 to -106 per mil magmatic, -85 to -88 per mil hydrothermal	-100 and -114 per mil magmatic, -108 to -134 per mil greisen

abbreviations are qtz quartz, mu muscovite, cas cassiterite, wolf-peg wolframite-bearing pegmatite

granitic rocks of similar origin that were emplaced into the same metasedimentary sequence. At Nong Sua the orthomagmatic fluid is interpreted to have had low salinity (Chapter 4). At Pilok, the composition of the orthomagmatic fluid is unknown, but the style and timing of the alteration and mineralization provide some clues as to the nature of this fluid. The earliest fluid event recognized at Pilok is the "magmatic metasomatism" that affected the biotite and leucocratic granites. The involvement of fluids is indicated by myrmekite, granophyre and local replacement textures (Chapter 5). It is reasonable to assume that this event reflects the interaction between orthomagmatic fluids and the host granite, and it is therefore significant that there was no associated mineralization. Since the partitioning of tin between aqueous fluids and melt is strongly dependent on the chlorinity of the aqueous fluid (Taylor, 1988), it follows that the orthomagmatic fluid must have had low to moderate salinity, i.e., if a hypersaline orthomagmatic fluid was involved, it would have been tin-rich, and early, high temperature cassiterite mineralization associated with feldspathic alteration would probably have occurred.

Some halite-saturated fluid inclusions (32 to 35 wt % NaCl eq.) with high concentrations of bivalent cations occur at Pilok, but they are rare and have been observed only along secondary planes in quartz. By contrast, systems that are interpreted to have had orthomagmatic hypersaline fluids typically contain abundant, primary, halite-saturated (and commonly sylvite-saturated) fluid

inclusions (cf. Kelly and Turneaure, 1970; Charoy and Pollard, 1989). As noted above the orthomagmatic fluid at Nong Sua was of low salinity, but rare, secondary, halite-saturated fluid inclusions (33 to 36 wt % NaCl eq. with significant concentrations of Ca and other bivalent cations) are also present in this deposit. The high salinity fluids were interpreted to reflect the influx of formational or metamorphic fluids (Chapter 4). However, the high concentrations of Fe and Mn chlorides interpreted for the external fluid may be the result of fluid-wallrock interaction, with the Fe and Mn originating by exchange with the aplite-pegmatite (for Ca-Mg). It is probable that the high salinity fluids at Pilok are of similar origin to those at Nong Sua. The presence of this fluid at both deposits therefore is probably related to the metasedimentary sequence into which they were emplaced.

The early fluid evolution at Nong Sua is interpreted to have involved three component mixing between aqueous orthomagmatic fluids, external saline aqueous fluids and external aqueous-carbonic fluids (Chapter 4). The early fluid evolution at Pilok is less well constrained, but there is evidence of mixing between aqueous and aqueous-carbonic fluids (Chapter 5). One difference between the fluid evolution of Nong Sua and Pilok is the influx of late, low temperature-salinity fluids of probable meteoric origin in the latter deposit. The lack of these late fluids, the lack of pervasive alteration and the restricted range of isotopic compositions at Nong Sua, suggest a relatively closed-system. The pervasive

alteration at Pilok indicates a relatively open system, typical of a stockwork. However, the variability of the isotopic data suggests that the fluid evolution was more complex than simple magmatic-meteoric mixing.

The majority of the granites and deposits in the western Thai belt were emplaced in the Kaeng Krachan Group, or in the south, the stratigraphically equivalent Phuket Group. At Pilok the mineralization is associated with granite-metasedimentary (Kaeng Krachan Group) contacts (cf. Fig. 1 in Chapter 5). This relationship also typifies mineralization elsewhere in the west-central region (cf. Fig. 1 in Chapter 6). The granite-metasedimentary relationship at the Nong Sua aplite-pegmatite is less certain, but the pegmatite deposits in the Tap Sakae and Bang Saphan areas <100 km to the south are hosted by the Kaeng Krachan Group (Charusiri, 1989; and this study, Appendix VIII). In the Phuket-Phang Nga area, most of the mineralized pegmatites are hosted by the Phuket Group metasediments (Garson *et al.*, 1975). These relationships suggest that fluid-metasediment interaction or metamorphic fluids may have been fundamental in controlling mineralization. One of the most striking similarities between the fluid inclusions at Nong Sua and Pilok is the composition and homogenization temperature of the aqueous-carbonic inclusions (Table 2). In both cases a metamorphic or evolved (meteoric or magmatic) origin is proposed for these fluids, involving different fO_2 conditions to account for the differences in the CH_4/CO_2 ratios (Chapters 4 & 5). The participation of metamorphic fluids is consistent with the association of

mineralization with granite-metasedimentary contacts. These fluids may therefore have controlled fO_2 and hence played a major role in the precipitation of cassiterite (Chapters 4 & 5). This differs from models that propose that magmatic fluids need to be "ponded" at the apices of granite intrusions to form tin mineralization, e.g., (Jackson et al., 1989), or other granite-related mineralization, e.g., the Henderson molybdenum deposit (White et al. 1981). Such "ponding" may have also occurred in western Thailand, but intrusion apices are commonly not the site of mineralization.

The little additional fluid inclusion data that are available for other Sn-W deposits in the western Thai belt are from Zaw and Thet (1983), Zaw (1984), Charusiri (1989) and this study (Chapter 6 and Appendix VIII). These data are summarized in Table 3 (note that for each of the deposits studied by Charusiri, 1989, only one sample was analyzed). Most of these deposits contain aqueous and aqueous-carbonic fluid inclusions. The latter are interpreted to have had an evolved (meteoric or magmatic fluid exchange with the metasediments) or metamorphic origin, similar to those at Nong Sua and Pilok. Minor hypersaline fluid inclusions are present at some deposits but, like Nong Sua and Pilok, they do not appear to be directly related to mineralization and are probably formational or metamorphic fluids. Possible exceptions are the primary fluid inclusions hosted by cassiterite at Takua Pit Thong (Chapter 6). These inclusions (two-phase L-V) homogenize at 400° to 490°C, are 22 to 26 wt % NaCl eq., and contain high concentrations of Fe,

TABLE 3
FLUID INCLUSIONS IN OTHER WESTERN THAI DEPOSITS

Deposit	Aqueous Inclusions	Aqueous-Carbonic Inclusions	Reference
Mawchit (Burma) vein-hosted	Th 120° to 270°C, 3.8 to 8.7 wt % NaCl eq.	none reported	Zaw and Thet (1983)
Tavoy Township (Burma), vein-hosted	Th 140° to 270°C, 1.8 to 5.0 wt % NaCl eq.	none reported	Zaw (1984)
Vistae pegmatite Tap Sakae	Th 288° to 423°C, 2.5 to 12 wt % NaCl eq.	none reported	Charusiri (1989)
Tor Soong pegmatite Phuket	Th 265° to 325°C, 0 to 3 wt % NaCl eq.	none reported	Charusiri (1989)
Khao Kow disseminated Phuket	Th 278° to 294°C, 0.6 to 3 wt % NaCl eq.	present but not analyzed	Charusiri (1989)
Takua Pit Thong sulphide replacement	Th 130° to 490°C, 0 to 26 wt % NaCl eq., primary L-V inclusions in cassiterite more restricted Th 400° to 490°C, 22 to 26 wt % NaCl eq., some L-V-H inclusions 30 to 38 wt % NaCl eq.	liquid-rich and vapour-rich, wide range in X_{CH_4} 0.0 to 1.0	this study, Chapter 6
Som Pom, stockwork & disseminated	L-V and L-V-H reported, none analyzed	vapour-rich reported, none analyzed	this study, Appendix VIII
Chat Chai stockwork	L-V and L reported, none analyzed	liquid-rich and vapour-rich reported, none analyzed	this study, Appendix VIII
Ban Tha Lao Hub Kapong	L-V and L reported, none analyzed	liquid-rich and vapour-rich reported, none analyzed	this study, Appendix VIII
Thai Prasit pegmatite Tap Sakae	L-V reported, none analyzed	vapour-rich reported, none analyzed	this study, Appendix VIII

Mn, and Ca chlorides (Chapter 6). They may represent an external fluid that has undergone exchange with granite (to account for the Fe and Mn) as at Nong Sua or, alternatively, they may represent a moderately high salinity orthomagmatic fluid. The aqueous-carbonic fluid inclusions at Takua Pit Thong also contrast with those present elsewhere in western Thailand by having a much greater range of compositions. Some aqueous-carbonic inclusions have X_{CH_4} values of >0.7 (Chapter 6). This may be related to the fact that the hosts are calc-silicate hornfels and marble at Takua Pit Thong, whereas the other deposits are hosted by pelitic rocks. A feature that Takua Pit Thong has in common with some other deposits, including Pilok, is the presence of late, low temperature-salinity fluids of probable meteoric origin.

In summary, the key features of the western Thai belt are a moderately deep level of emplacement and the occurrence of low to moderately saline aqueous and aqueous-carbonic fluid inclusions. The latter are interpreted to be metamorphic in origin or meteoric-magmatic fluids that have undergone exchange with metamorphic rocks.

COMPARISON OF THE WESTERN THAI BELT TO OTHER Sn-W PROVINCES

Table 4 summarizes the principal characteristics of most of major the Sn-W provinces around the world. Although pressure or depth of emplacement is

generally difficult to estimate, it is apparent from Table 4 that there is a strong correlation between the depth of emplacement and fluid composition. Where tin deposits are associated with high-level intrusions, indicated by the presence of coeval volcanic rocks, cassiterite is invariably associated with, or was preceded by, high temperature hypersaline fluids e.g., Bolivia (Kelly and Turneaure, 1970), Erzgebirge (Thomas, 1979), New South Wales (Eadington, 1983) and Queensland (Charoy and Pollard, 1989). By contrast, deeper plutonic belts lack coeval volcanism and typically contain at least some mineralized pegmatites, e.g., the Portuguese Hercynian (Gouanvic and Gagny, 1987), the Spanish Hercynian (Robles *et al.*, 1991), Malaysia (Hosking, 1973) and western Thailand (this study). The Sn-W deposits in plutonic belts are commonly dominated by aqueous-carbonic and/or low to moderately saline aqueous fluids, e.g., the Portuguese, French and Spanish Hercynian (Kelly and Rye, 1979; Bril, 1982; Mangas and Arribas, 1987), Malaysia (Schwartz and Askury, 1989), Indonesia (Schwartz and Surjono, 1990) and western Thailand deposits (this study).

A potential explanation for the contrasting fluid evolution of sub-volcanic and plutonic systems is the effect of pressure on the fluid-melt partitioning of chlorine. At pressures less than approximately 1 kbar the first aliquot of water exsolved from a vapour-saturated melt has a relatively low salinity. Successive aliquots of water contain increasingly higher salt concentrations. At pressures greater than approximately 2 kbar chlorine exhibits the opposite behaviour; the first aliquot has

TABLE 4
CHARACTERISTICS OF OTHER Sn-W (W-Sn) PROVINCES

Region	Deposit Types	Carbonic and Aqueous-Carbonic Inclusions	Aqueous Inclusions	Depth or Pressure	Other Features	Reference
Bolivia	stockwork, breccia	rare or absent	early high temperature hypersaline, later low temperature-salinity fluids, some boiling	associated volcanic rocks, P est <500 bars	B-rich	Kelly and Turneure (1970), Turneure (1971), Grant <i>et al</i> (1973, 1980), Sugaki <i>et al</i> (1988)
Erzgebirge Germany	stockwork	rare or absent	early hypersaline, later low temperature-salinity fluids	associated volcanic rocks, P est <500 to 2000 bars	F-rich and Li-rich	Thomas (1979, 1982), Tischendorf and Föster (1990), Thomas <i>et al</i> (1991)
Queensland Australia	stockwork	rare or absent	early high temperature hypersaline, later low temperature-salinity	associated volcanic rocks, P est <1000 bars	F-rich	Witt (1988), Charoy and Pollard (1989)
New South Wales Australia	vein, disseminated	rare or absent	early high temperature hypersaline, later low temperature-salinity	associated volcanic rocks, P est <700 bars	F-rich	Eadington (1983), Sun and Eadington (1987)
Tasmania Australia	sulphide replacement, skarn, stockwork	generally rare, abundant at some deposits, H ₂ O-CO ₂ and hydrocarbons at Aberfoyle, some fluid immiscibility	early high temperature-salinity at some deposits, moderate temperature-salinity at some deposits, some boiling	no associated volcanic rocks but intrusions are high-level, P est 200-500 to <2000 bars	generally F-rich	Collins (1981), Kwak and Askins (1981), Patterson <i>et al</i> (1981), Solomon (1981), Hoffmann <i>et al</i> (1988), Wright and Kwak (1989)
Cornwall England	vein, breccia	rare or absent	early high temperature moderate salinity, also early hypersaline?, late low temperature-salinity, also some late Ca-Mg-rich fluids	no associated volcanic rocks, depth est. 3-6 km	some deposits B-rich, other deposits Li-F-rich	Charoy (1979), Jackson <i>et al</i> . (1977, 1982, 1989), Alderton and Rankin (1983), Rankin and Alderton (1983, 1985), Shepherd <i>et al</i> . (1985), Bottrell and Yardley (1988)

TABLE 4 (continued)

Region	Deposit Types	Carbonic and Aqueous-Carbonic Inclusions	Aqueous Inclusions	Depth or Pressure	Other Features	Reference
French Hercynian	vein	early, low salinity H ₂ O-CO ₂ -CH ₄ -N ₂ , moderate temperature	later low salinity fluids	no associated volcanic rocks, fluid pressure est. ~700 bars	B-rich	Bril (1982), Bril and Ramboz (1982), Ramboz <i>et al.</i> (1985), Dubessy <i>et al.</i> (1987)
Spanish-Portugese Hercynian	vein, pegmatite, disseminated	early, low salinity H ₂ O-CO ₂ -CH ₄ -N ₂ -H ₂ S, moderate to high temperature	later low salinity fluids, some late Ca-Mg-rich fluids	no associated volcanic rocks, fluid pressure est. 100-3250 bars	some deposits are Li or B rich	Kelly and Rye (1979), Bussink <i>et al.</i> (1984), Noronha (1984), Campbell <i>et al.</i> (1988), Mangas and Arribas (1987, 1988a, 1988b, 1991)
Main Range Malaysia	vein, replacement, pegmatite	early carbonic CO ₂ , moderate temperature	later low salinity fluids, rare hypersaline fluids	no associated volcanic rocks, fluid P est. ~1 kbar	some B-rich	Schwartz and Askury (1989)
Indonesia	disseminated stockwork	some early CO ₂ -CH ₄ , moderate temperature, or absent	some later low to moderately saline, or only low to moderately saline	no associated volcanic rocks, fluid P est. 700-2000 bars	some F-rich	Schwartz and Surjono (1990, 1991)

the highest salinity and the salinity decreases with successive aliquots (Candela, 1989; Cline and Bodnar, 1991). The partitioning behaviour of chlorine has been used to explain the association of high temperature hypersaline fluid inclusions with high-level porphyry copper deposits (op. cit.). This also seems to be a reasonable explanation for the association of high temperature hypersaline fluid inclusions with sub-volcanic tin deposits. Boiling is interpreted to have occurred in many of the sub-volcanic systems, e.g., Bolivia (Kelly and Turneaure, 1970) but, like porphyry copper systems, boiling is not required for hypersaline fluids to have been produced.

In the plutonic systems the pressure at which vapour saturation occurred may also explain the low to moderate salinity of the aqueous fluid inclusions, as the late magmatic fluids in these systems would have decreasingly lower salinity. The common occurrence of aqueous-carbonic inclusions indicates that either the magmatic fluid is aqueous-carbonic or that metamorphic fluids were involved. Although the solubility of CO_2 in granitic melts is pressure dependent, the change of solubility from 3 to 1 kbar is negligible (Dingwell, 1986; Stolper *et al.*, 1987). Therefore the lack of carbonic fluids in the high-level intrusions argues against a magmatic origin for the CO_2 . In several of the plutonic systems a metamorphic origin for the carbonic fluids has been proposed (including this study) and in one case (Aberfoyle, Tasmania) the presence of hydrocarbons in fluid inclusions strongly supports a metasedimentary source (Hoffmann *et al.* 1988).

It therefore seems reasonable to propose that, in most plutonic systems, the aqueous carbonic fluid inclusions have a metamorphic origin (or fluids that underwent exchange with metamorphic rocks). The reason why aqueous-carbonic inclusions are rare in sub-volcanic systems is probably also related to the depth of emplacement. The sub-volcanic plutons are emplaced in relatively cool rocks. At 6 wt % NaCl, <500 bars and <300°C the maximum X_{CO_2} of external aqueous fluids is <0.05 (Bowers and Helgeson, 1983). If the high salinity orthomagmatic fluids circulate outside the pluton the amount of CO_2 that could be dissolved into this fluid would be much less, and any carbonic fluids produced would have a high X_{CO_2} and low density, and hence might "boil off" and not be trapped as fluid inclusions. By contrast, the higher pressure and lower salinity of plutonic systems allows for the formation of supercritical aqueous-carbonic fluids or, if below the two-phase solvus, relatively dense aqueous-carbonic fluids with moderately high X_{CO_2} compositions (Bowers and Helgeson, 1983).

It is interesting that the deposits in Cornwall and Tasmania have some of the fluid inclusion characteristics of both sub-volcanic and plutonic environments (Table 4). In both cases the depth of emplacement is interpreted to be relatively shallow, but coeval volcanism is lacking. These regions therefore seem to be intermediate between sub-volcanic and plutonic environments both in terms of depth of emplacement and fluid composition.

Further support for a relationship between depth of emplacement and fluid

composition can be found by examining other granite-related deposits. The low oxygen fugacities interpreted for ilmenite-series granites are thought to be related to a carbonaceous metasedimentary source (Ishihara, 1981). It follows that if the aqueous-carbonic inclusions associated with these granites are magmatic, they also originated from the same carbonaceous metasediments (e.g., Schwartz and Asbury, 1989). However, molybdenum deposits, associated with magnetite-series granites, have depth-fluid composition relationships similar to Sn-W deposits. Deposits emplaced at high levels, e.g., Henderson, contain abundant hypersaline fluid inclusions (White et al. 1981), whereas those emplaced at a deeper level, e.g., Trout Lake, are dominated by aqueous-carbonic fluid inclusions (Linnen and Williams-Jones, 1990). The similarities of fluid compositions associated with both ilmenite-series and magnetite-series granites argues against a magmatic origin for the aqueous-carbonic fluid inclusions.

In conclusion, the depth of emplacement was one of the principal factors that controlled mineralization style in the western Thai Sn-W province. The depth of emplacement also seems to have been important in other Sn-W provinces and in controlling granite-related mineralization in general.

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CHAPTER 8

CONCLUSIONS

Textural relationships between disseminated cassiterite and feldspar at Nong Sua indicate that cassiterite crystallized from melt at or prior to vapour saturation. The pegmatite evolved from early plagioclase-tourmaline-muscovite-rich units with peraluminous compositions to late K-feldspar-rich units that are more alkalic. The covariance of Sn with K_2O indicates that the tin concentration of the melt increased with fractionation. By using the tin contents of K-feldspar-rich samples, experimental tin solubility, and empirical mineral-melt partition data, it has been shown that the melt contained on the order of 700 ppm Sn, which is consistent with the magmatic crystallization of cassiterite. Mineralization continued after vapour saturation to form vein-hosted cassiterite and wolframite, and contact-style cassiterite. In the case of the former, the timing of mineralization is constrained by crosscutting relationships to have been at or near solidus conditions of the pegmatite.

The mineralogy of the Nong Sua pegmatite supports the above interpretation. Primary magmatic tourmaline compositions overlap with those of hydrothermal origin and, from tourmaline intergrowths with disseminated and vein cassiterite, a penecontemporaneous timing of these minerals is established. Both tourmaline and cassiterite therefore must have crystallized from melt and aqueous fluid; late Ca-Mg

enrichment in tourmaline records the involvement of external fluids. Elevated tin abundances in the cores of garnet crystals that are defined by fluid inclusions support a late-stage, high tin concentration in the melt. The restriction of Ti-Fe-Nb-Ta oxides to aplite-pegmatite suggests that these oxides have a magmatic origin, and their common occurrence in disseminated cassiterite and absence in vein cassiterite also supports crystallization of cassiterite from melt and later fluids. The composition of muscovite has been modified and indicates water-rock interaction at subsolidus conditions. However, the incipient nature of sericite alteration suggests low water-rock ratios.

Vapour saturation of the melt at Nong Sua occurred at approximately 650°C and caused the pressure to increase to 3.8 kbar. The orthomagmatic fluid had a composition of 3 wt. % NaCl eq., and as a consequence of this composition, tin was partitioned in favour of the melt. The fluid overpressure caused brittle failure in the aplite units. Fluids migrated into the low pressure fracture zones to form quartz-tourmaline veins and cassiterite-wolframite mineralization. This event can be modeled by an isothermal decompression to 2.7 kbar, followed by cooling. In both the vein and disseminated mineralization three component mixing occurred between low salinity orthomagmatic fluids, external high salinity fluids enriched in Ca, Fe, and Mn, and external carbonic fluids. Cassiterite deposition is interpreted to have been controlled by increasing fO_2 , whereas wolframite deposition resulted from the mixing of W-rich with Fe-Mn-rich fluids. In both cases cooling may also

have contributed to mineralization. The carbonic fluids were probably derived from metasedimentary rocks and the high salinity fluid is interpreted to have had a metamorphic or evolved meteoric origin; oxygen and hydrogen isotopic data are consistent with either origin.

The W-Sn-polymetallic mineralization at Pilok was closely associated with the intrusion of a leucocratic granite. Early, pervasive fluids infiltrated biotite, biotite-muscovite and leucocratic granite which resulted in a "two-phase variant" overprint of primary magmatic textures and possible re-melting of the granitic rocks. This event was followed by tourmalinization. A contact metamorphic-metasomatic aureole was developed in the pelitic metasediments, coeval with the early fluid evolution of the granites. The bulk of the wolframite mineralization post-dated tourmalinization and pre-dated greisen alteration, but the temperature and pressure of this event is poorly constrained. Cassiterite-sulphide mineralization is associated with greisen alteration of leucocratic granite and is interpreted to have been deposited from aqueous-carbonic fluids at 300° to 450°C. The aqueous-carbonic fluid inclusions are considered to have originated either from metamorphism of carbonaceous pelite or from interaction between magmatic or meteoric fluids with metamorphic rocks. It is proposed that low fO_2 aqueous-carbonic fluids remobilized tin and tungsten from granitic rocks and deposited cassiterite and wolframite in response to increasing fO_2 , pH, or decreasing temperature.

The evolution of mineralization at Takua Pit Thong and Kow Daeng is constrained by the timing of different alteration-mineralization events. From these constraints it is apparent that there was an early "pegmatite" and skarn Sn-Nb-Ta event at Kow Daeng followed by a sulphide replacement Sn-Cu-Zn-Be-F-S event at Takua Pit Thong. Iron enrichment trends in the silicate minerals associated with cassiterite are observed in both deposits. At Takua Pit Thong, cassiterite precipitated at $\geq 500^{\circ}\text{C}$ from a moderately high salinity aqueous fluid. The carbonic fluids at Takua Pit Thong are interpreted to have had a metamorphic source and the presence of liquid-rich (and CO_2 -rich) and vapour-rich (and CH_4 -rich) fluid inclusions suggests that two immiscible aqueous-carbonic fluids coexisted.

The depth of emplacement was one of the principal factors that controlled mineralization style in the western Thai Sn-W province. The depth of emplacement also seems to have been important in other Sn-W provinces and in controlling granite-related mineralization in general.

CONTRIBUTIONS TO ORIGINAL KNOWLEDGE

1) This is the first investigation in which both the magmatic and hydrothermal evolution of a pegmatite-hosted Sn-W deposit have been characterized, and one of only a few studies of this class of Sn-W deposit.

2) Although magmatic crystallization of cassiterite has been previously proposed, this is the first time that partition and solubility data have been used to

demonstrate the viability of the magmatic model.

3) This is one of a small number of investigations that have characterized the compositions of magmatic and hydrothermal tourmaline from the same deposit; the data suggest that the mineral-melt and fluid mineral partition coefficients for tourmaline are similar.

4) This study indicates that Sn-W mineralization in a single metallogenic belt can form by direct crystallization from melt, precipitation from orthomagmatic fluids and by remobilization.

5) It has been shown that the depth of emplacement is a major control of mineralization style and that fluid composition correlates with emplacement depth. This is the first study to apply fluid-melt chlorine partition data to the genetic interpretation of Sn deposits. Application of these parameters provides a new tool for the classification of Sn-W metallogenic provinces.

6) This is the first comprehensive study of the origin of some of the mineralization in one of the world's most important tin belts.

RECOMMENDATIONS FOR FUTURE WORK

This study has characterized the evolution of a few representative deposit types in western Thailand, but many problems remain unresolved. The depth of emplacement of most of the deposits is poorly constrained. Only a single pegmatite deposit was investigated; other pegmatites, in particular the lepidolite-bearing type,

should be investigated to further characterize this style of mineralization. Other pegmatite deposits, such as those in the Portugese and Spanish Hercynian could provide a valuable contrast to the Thai examples, as could the relationships between pegmatite and stockwork and vein deposits in these belts. The origin of the contrasting styles of mineralization in the Takua Pit Thong area requires additional research. The sulphide replacement deposits at Takua Pit Thong are hosted by marble, but the marble at Som Porn hosts stockwork mineralization; the differences in the mineralization at these deposits are unexplained. New experimental data on fluid-melt, mineral-melt and fluid-mineral partition of Sn would greatly aid future modelling of these systems, as would data on the partitioning behaviour of W which is even less well understood.

APPENDIX I

WHOLE-ROCK DATA FROM NONG SUA

Abbreviations are apl-peg aplite in the banded pegmatite unit, BP fine-grained banded tourmaline-rich pegmatite cgF peg coarse-grained pegmatite in the banded unit. Regional granites are BGr biotite granite and H-BGr hornblende-biotite granite

Sample Type	PK-1 aplite	PK-3c aplite	PK-12b aplite	PK-25 aplite	PK-38c aplite	PK-38d aplite	PK-42b aplite	PK-44 aplite	PK-58 aplite	PK-16a cgl'
S102	75.25	75.15	73.71	75.97	72.99	75.51	74.55	76.11	75.28	80.23
TiO2	0.05	0.05	0.03	0.04	0.05	0.06	0.04	0.04	0.05	0.04
Al2O3	15.06	15.00	15.28	14.54	15.19	14.74	14.76	14.54	14.77	11.66
Fe2O3	0.40	0.32	0.26	0.40	0.61	0.80	0.56	0.53	0.66	0.21
MnO	0.09	0.04	0.01	0.08	0.05	0.02	0.08	0.11	0.14	0.01
MgO	0.01	0.01	0.06	0.01	0.01	0.01	0.01	0.02	0.01	0.01
CaO	0.60	0.71	0.79	0.72	0.26	1.40	0.63	0.60	0.77	0.44
Na2O	3.61	3.85	4.43	3.97	2.59	5.69	3.58	3.99	3.98	3.03
K2O	4.51	4.30	4.69	3.83	7.91	0.99	5.17	3.40	3.68	3.89
P2O5	0.03	0.05	0.05	0.03	0.08	0.06	0.06	0.04	0.06	0.04
LOI	0.76	0.77	0.46	0.64	0.33	0.48	0.56	0.70	0.60	0.43
Total	100.36	100.25	99.76	100.20	100.03	99.77	99.99	100.55	99.98	99.97

Sample	PK-1	PK-3c	PK-12b	PK-25	PK-38c	PK-38d	PK-42b	PK-44	PK-58	PK-16a
Nb ppm	23	21	27	35	15	34	16	28	42	28
Zr ppm	20	17	20	28	9	34	23	22	35	3
Y ppm	17	16	13	21	5	39	16	26	29	5
Sr ppm	14	12	15	8	13	12	14	11	7	7
Rb ppm	404	431	421	377	680	119	480	342	336	336
Pb ppm	34	32	36	24	47	25	44	27	20	26
Th ppm	5	7	9	6	8	13	5	8	8	5
U ppm	5	5	5	5	5	5	5	5	5	5
F ppm	300	460	180	440	270	480	350	520	460	250
Li ppm	17	23	11	21	8	40	17	25	32	17
Na ppm	6	10	5	18	6	5	7	10	7	7
Ca ppm	116	95	173	14	1190	2	327	2	385	290
Mn ppm	23	13	19	14	95	53	106	10	20	8
V ppm	19	13	< 10	12	< 10	< 10	< 10	10	< 10	< 10
Cr2O3 ppm	< 15	< 15	< 15	< 15	< 15	18	< 15	< 15	< 15	< 15
Ni ppm	10	13	< 10	14	15	13	12	10	11	< 10
BaO ppm	< 10	< 10	< 10	< 10	46	40	41	< 10	< 10	20
q	35.78	34.91	28.34	36.73	27.29	35.79	32.75	38.67	36.56	46.87
or	26.65	25.41	27.72	22.63	46.75	5.85	30.55	20.09	21.75	22.99
ab	30.55	32.58	37.49	33.59	21.92	48.15	30.29	33.76	33.68	25.64
an	2.76	3.18	3.61	3.40	0.80	6.53	2.71	2.75	3.40	1.93
c	3.23	2.85	1.59	2.62	2.07	1.91	2.28	3.29	2.99	1.76
mt	0.16		0.00	0.13			0.15	0.22	0.32	
il	0.09	0.07	0.02	0.08	0.05	0.03	0.07	0.08	0.09	0.02
hem	0.29	0.32	0.26	0.31	0.61	0.80	0.46	0.38	0.44	0.21
ap	0.08	0.12	0.11	0.06	0.17	0.15	0.15	0.08	0.15	0.09
ru		0.01	0.02		0.02	0.05				0.03
hy	0.02	0.02	0.15	0.02	0.02	0.02	0.02	0.05	0.02	0.02

Sample	PK-17a	PK-17d	PK-19a	PK-19c	PK-23c	PK-57	PK-2b	PK-17c	PK-17e	PK-19b
Type	Graphic	cgF peg	Graphic	cgF peg	cgF peg	Graphic	QFT peg	QFT peg	QFT peg	QFT peg
SiO2	72.19	68.22	74.84	71.13	80.98	73.59	77.28	78.96	75.82	79.37
TiO2	0.02	0.02	0.02	0.02	0.02	0.02	0.12	0.06	0.08	0.04
Al2O3	15.09	17.38	13.61	16.04	12.19	14.58	13.76	13.47	14.68	13.06
Fe2O3	0.14	0.11	0.24	0.19	0.17	0.13	2.26	1.00	1.16	0.58
MnO	0.01	0.01	0.01	0.01	0.01	0.01	0.07	0.02	0.02	0.01
MgO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.09	0.01
CaO	0.04	0.05	0.04	0.12	0.90	0.14	0.65	0.77	0.71	0.91
Na2O	1.97	2.45	1.96	2.71	5.28	2.57	4.09	4.83	4.30	5.17
K2O	10.08	11.60	8.86	9.62	0.45	8.73	0.76	0.54	2.71	0.64
P2O5	0.05	0.08	0.03	0.05	0.03	0.05	0.02	0.02	0.05	0.02
LOI	0.21	0.21	0.22	0.25	0.27	0.18	0.58	0.63	0.44	0.35
Total	99.77	100.11	99.86	100.10	100.28	99.99	99.59	100.36	100.08	100.17

Sample	PK-17a	PK-17d	PK-19a	PK-19c	PK-23c	PK-57	PK-2b	PK-17c	PK-17e	PK-19b
Nb ppm	3	3	3	3	24	3	95	28	47	37
Zr ppm	3	3	3	3	4	3	45	4	5	10
Y ppm	5	5	5	5	6	5	10	14	16	16
Sr ppm	13	12	11	9	11	15	14	18	16	11
Rb ppm	825	995	702	811	37	678	68	47	215	52
Pb ppm	60	69	44	50	13	58	13	15	22	12
Th ppm	5	5	5	5	5	5	10	5	9	6
U ppm	5	5	5	5	5	5	5	5	5	5
F ppm	70	60	82	90	98	96	900	390	440	200
Li ppm	5	4	4	14	21	5	33	35	24	20
Ta ppm	5	3	5	5	3	3	21	5	16	8
B ppm	120	68	435	347	215	45	8504	2762	3838	1427
Sn ppm	23	42	16	34	4	27	3	29	13	11
V ppm	13	< 10	< 10	16	< 10	< 10	22	< 10	< 10	< 10
Cr2O3 pp	26	< 15	76	< 15	< 15	< 15	< 15	< 15	19	< 15
Ni ppm	< 10	11	< 10	13	< 10	13	10	10	10	< 10
BaO ppm	< 10	32	< 10	< 10	10	54	< 10	41	< 10	26
q	22.19	9.67	29.50	18.41	46.68	25.04	49.23	47.17	38.91	44.94
or	59.57	68.55	52.36	56.85	2.66	51.59	4.49	3.19	16.02	3.78
ab	16.67	20.73	16.59	22.93	44.68	21.75	34.61	40.87	36.39	43.75
an			0.02	0.28	4.28	0.39	3.08	3.66	3.22	4.38
c	0.98	0.89	0.79	1.07	1.45	0.76	5.08	3.60	3.49	2.26
mt										
tl	0.02	0.02	0.02	0.02	0.02	0.02	0.14	0.05	0.05	0.03
hem	0.14	0.11	0.24	0.19	0.17	0.13	2.26	1.00	1.16	0.58
ap	0.11	0.18	0.06	0.11	0.07	0.11	0.05	0.06	0.11	0.05
ru	0.01		0.01	0.01	0.00	0.00	0.05	0.03	0.05	0.03
hy	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.07	0.22	0.02

Sample Type	PK-31 apl-peg	PK-38b BP	PK-41b BP	PK-23a BP	PK-17b BP	PK-26a BGr	PK-27a BGr	PK-28 H-BGr	PK-29 H-BGr	PK-30 H-BGr
SiO2	76.70	75.94	78.26	75.75	75.54	70.18	72.33	71.87	69.55	74.03
TiO2	0.02	0.06	0.07	0.05	0.06	0.51	0.19	0.56	0.37	0.21
Al2O3	14.85	14.81	13.89	14.67	14.59	15.05	14.78	13.34	15.14	13.79
Fe2O3	0.27	0.78	0.77	0.80	0.80	2.94	1.73	3.26	2.26	1.60
MnO	0.01	0.02	0.02	0.02	0.02	0.07	0.05	0.09	0.06	0.07
MgO	0.01	0.03	0.01	0.01	0.01	0.87	0.46	0.96	0.65	0.29
CaO	1.39	0.90	0.64	0.63	0.86	1.82	1.25	3.01	2.46	1.54
Na2O	6.37	4.84	4.51	4.29	4.04	3.17	2.80	3.19	3.09	3.28
K2O	0.42	2.34	1.59	3.75	3.80	5.14	5.85	3.26	5.52	5.08
P2O5	0.05	0.05	0.03	0.05	0.03	0.17	0.05	0.24	0.16	0.07
LOI	0.29	0.36	0.61	0.35	0.48	0.35	0.55	0.46	0.58	0.36
Total	100.38	100.13	100.40	100.36	100.21	100.28	100.15	100.28	100.91	100.36
Sample	PK-31	PK-38b	PK-41b	PK-23a	PK-17b	PK-26a	PK-27a	PK-28	PK-29	PK-30
Nb ppm	35	47	40	37	31	25	11	35	20	16
Zr ppm	28	15	7	17	13	179	95	325	206	117
Y ppm	23	11	12	12	9	23	18	16	11	<5
Sr ppm	8	17	13	16	15	108	80	249	358	162
Rb ppm	376	184	160	282	305	317	414	361	380	587
Pb ppm	24	23	10	27	24	47	80	40	42	62
Th ppm	8	10	9	15	11	40	36	67	46	47
U ppm	5	5	5	5	5	5	5	5	5	5
F ppm	140	340	480	300	420	1300	540	1400	1050	1700
Li ppm	14	17	17	16	21	101	81	149	85	82
Ta ppm	10	14	9	15	17	6	6	4	1	4
B ppm	369	1770	1543	2222	2268	202	21	107	37	76
Sn ppm	12	18	4	7	36	3	1	1	1	3
V ppm	< 10	16	16	< 10	< 10	38	28	27	19	19
Cr2O3 ppm	20	< 15	< 15	< 15	19	< 15	690	79	< 15	< 15
Ni ppm	< 10	< 10	< 10	< 10	< 10	12	19	13	15	38
BaO ppm	42	17	12	< 10	42	484	503	253	654	209
q	35.20	36.99	44.65	35.21	35.72	27.64	30.46	33.83	24.67	31.98
or	2.48	13.83	9.40	22.16	22.46	30.38	34.57	19.27	32.62	30.02
ab	53.90	40.96	38.16	36.30	34.19	26.40	23.69	26.99	26.15	27.75
an	6.55	4.15	2.96	2.82	4.07	7.92	5.81	12.45	11.14	7.16
c	1.52	2.79	3.67	2.52	2.34	1.45	1.71			0.27
mt										
il	0.02	0.05	0.05	0.04	0.04	0.15	0.11	0.19	0.13	0.15
hem	0.27	0.78	0.77	0.80	0.80	2.94	1.73	3.26	2.26	1.60
ap	0.12	0.11	0.08	0.11	0.07	0.39	0.14	0.56	0.37	0.17
ru	0.01	0.04	0.05	0.03	0.04	0.43	0.13	0.21	0.30	0.13
hy	0.02	0.07	0.02	0.02	0.02	2.17	1.15	2.39	1.62	0.72

APPENDIX II

ELECTRON MICROPROBE DATA FROM NONG SUA

IIa FELDSPAR DATA

The abbreviations used are: Anal no. analysis number, descrpt description, tw twinned, untw untwinned, perth perthite, mcr microcline, rpl replacement, myr myrmekite, v vein, c core, r rim, m middle, sg same grain as, cont in contact with, cass cassiterite, wolf wolframite, ovgr overgrowth, ser sericitized, gar garnet, pl plagioclase, ex exsolution, incl inclusion.

APLITE PLAGIOCLASE

Sample Anal no. descript	PK-1a 1 myr	PK-1a 1 myr	PK-1a 2 myr	PK-1a 4 myr	PK-1a 5 plag	PK-1a 7 plag	PK-3c 1 plag	PK-3c 4 plag cont 3	PK-3c 5 plag c	PK-3c 7 plag incl
Fe2O3	0.03	0.00	0.00	0.00	0.00	0.02	0.00	0.04	0.00	0.00
Na2O	10.36	10.36	11.07	10.15	10.77	10.09	11.01	10.63	10.63	11.22
SiO2	66.39	67.25	68.57	66.97	68.5	66.57	67.6	66.91	66.86	68.28
Al2O3	21.09	21.35	20.06	21.51	20.35	21.65	20.48	20.91	20.78	19.56
MgO	0.00	0.01	0.00	0.02	0.00	0.01	0.00	0.00	0.00	0.00
K2O	0.14	0.16	0.11	0.12	0.12	0.17	0.10	0.10	0.09	0.15
CaO	2.06	1.97	0.8	2.18	1.17	2.47	1.46	1.82	2.07	0.54
BaO	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.03	0.05
Tot.	100.07	101.1	100.61	100.95	100.92	100.97	100.64	100.41	100.46	99.81
Fe	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.001	0.000	0.000
Na	0.881	0.871	0.932	0.854	0.904	0.851	0.929	0.900	0.901	0.952
Si	2.912	2.916	2.977	2.908	2.966	2.895	2.944	2.922	2.922	2.989
Al	1.090	1.091	1.026	1.101	1.038	1.110	1.051	1.077	1.070	1.009
Mg	0.000	0.000	0.00	0.001	0.000	0.001	0.000	0.000	0.000	0.000
K	0.008	0.009	0.006	0.007	0.006	0.009	0.006	0.005	0.005	0.008
Ca	0.097	0.091	0.037	0.102	0.054	0.115	0.068	0.085	0.097	0.025
Ba	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.001
Xab	0.893	0.897	0.956	0.888	0.937	0.873	0.927	0.909	0.898	0.966
Xan	0.099	0.094	0.038	0.105	0.056	0.118	0.068	0.086	0.097	0.026
Xor	0.008	0.009	0.006	0.007	0.007	0.009	0.006	0.005	0.005	0.008
Sample Anal no. descript	PK-3c 9 plag repl	PK-3c 12 plag cont 10	PK-3c 14 plag cont gar	PK-3c 15 plag r	PK-3c 16 plag c sq 15	PK-9c 1 myr r	PK-9c 2 myr c sq 1	PK-9c 3 plag sq 1	PK-9c 4 myr	PK-9c 5 plag
Fe2O3	0	0	0	0	0.05	0	0.03	0.02	0.04	0.0
Na2O	10.76	11.36	11.34	10.82	10.21	11.26	11.02	10.99	11.1	10.75
SiO2	67.27	68.22	67.58	66.91	66.11	67.72	67.26	67.47	66.97	67.78
Al2O3	21.07	20.02	19.39	20.7	21.44	19.94	20.02	19.81	20.41	20.14
MgO	0	0	0	0	0	0	0	0.01	0	0.0
K2O	0.11	0.15	0.11	0.1	0.13	0.14	0.09	0.09	0.09	0.07
CaO	1.73	0.91	0.68	1.91	2.31	0.84	0.9	0.73	1.74	1.06
BaO	0	0.13	0	0.07	0	0	0	0	0.04	0.0
Tot.	100.94	100.78	99.09	100.49	100.25	99.9	99.31	99.14	99.88	99.3
Fe	0	0	0	0	0.002	0	0.001	0.001	0.001	0.0
Na	0.906	0.958	0.97	0.917	0.867	0.956	0.941	0.939	0.945	0.917
Si	2.922	2.9665	2.983	2.924	2.896	2.967	2.962	2.974	2.94	2.961
Al	1.079	1.026	1.009	1.066	1.107	1.03	1.039	1.029	1.056	1.045
Mg	0	0	0	0	0	0	0	0.001	0	0.0
K	0.006	0.008	0.006	0.006	0.007	0.008	0.005	0.005	0.005	0.004
Ca	0.081	0.043	0.032	0.089	0.108	0.039	0.042	0.034	0.058	0.05
Ba	0	0.002	0	0.001	0	0	0	0	0.001	0.0
Xab	0.913	0.95	0.962	0.906	0.882	0.953	0.952	0.96	0.937	0.945
Xan	0.081	0.042	0.032	0.088	0.11	0.039	0.043	0.035	0.058	0.051
Xor	0.006	0.008	0.006	0.006	0.007	0.008	0.005	0.005	0.005	0.004
Sample Anal no. descript	PK-9c 6 myr c	PK-9c 7 myr r	PK-25 1 plag c sq 1	PK-25 2 plag r overgr 2	PK-25 3 untw plag c	PK-25 8 untw r sq 8	PK-25 9 untw plag cont 12	PK-25 13 plag r	PK-31c 1 plag c	PK-31c 2 plag r
Fe2O3	0.03	0.01	0	0	0	0	0.01	0	0.02	0.0
Na2O	10.58	11.03	9.93	11.1	10.6	10.23	10.3	9.88	10.32	10.42
SiO2	66.97	67.47	66.12	68.19	67.21	66.71	66.2	66.24	66.82	66.65
Al2O3	20.49	20.03	21.04	19.67	20.37	20.85	20.66	20.94	20.79	20.91
MgO	0	0	0.01	0	0	0	0	0	0	0.0
K2O	0.13	0.12	0.16	0.08	0.12	0.15	0.1	0.18	0.09	0.07
CaO	1.34	0.78	1.96	0.5	1.16	1.84	1.7	1.81	1.69	1.84
BaO	0.01	0.05	0	0.06	0.04	0	0.02	0.05	0.03	0.0
Tot.	99.55	99.49	99.23	99.6	99.5	99.77	98.99	99.08	99.75	99.89
Fe	0.001	0.0005	0	0	0	0	0	0	0.001	0.0
Na	0.902	0.94	0.85	0.943	0.903	0.87	0.884	0.846	0.878	0.886
Si	2.944	2.966	2.918	2.989	2.953	2.928	2.929	2.925	2.932	2.923
Al	1.062	1.0375	1.094	1.016	1.055	1.079	1.077	1.09	1.075	1.081
Mg	0	0	0.001	0	0	0	0	0	0	0.0
K	0.007	0.0065	0.009	0.004	0.007	0.008	0.006	0.01	0.005	0.004
Ca	0.063	0.037	0.093	0.023	0.055	0.086	0.0805	0.0855	0.08	0.087
Ba	0	0.001	0	0.001	0.001	0	0	0.001	0.0005	0.0
Xab	0.928	0.956	0.893	0.972	0.936	0.902	0.911	0.899	0.912	0.907
Xan	0.065	0.037	0.097	0.024	0.057	0.0895	0.083	0.091	0.083	0.089
Xor	0.007	0.007	0.01	0.004	0.007	0.0085	0.006	0.011	0.005	0.004

Sample Anal no. descript	PK-31c 3 plag m sq 1,2	PK-31c 7 plag ex perth 6	PK-31c 8 plag	PK-31c 9 plag r	PK-31c 10 plag c sg 9	PK-31c 11 plag c	PK-31c 12 plag r sg 11	PK-31c 13 plag r	PK-31c 14 plag c sg 13	PK-31c 15 plag r
Fe2O3	0	0.01	0	0	0	0	0.03	0.01	0	0.0
Na2O	10.24	11.22	10.2	10.14	10.19	10.11	10.14	10.03	9.97	9.88
SiO2	66.4	68.79	66.38	66.13	66.46	66.53	66.07	65.84	65.59	65.22
Al2O3	21.05	19.42	20.95	21.09	21.07	20.79	20.76	21.2	21.12	21.1
MgO	0	0	0	0	0	0	0.01	0	0	0.0
K2O	0.11	0.15	0.1	0.08	0.15	0.1	0.09	0.1	0.07	0.08
CaO	2.07	0.31	1.9	2.06	2.07	1.88	1.77	2.11	2.2	2.18
BaO	0.02	0.07	0.01	0.05	0	0	0	0.04	0.04	0.05
Tot.	99.89	99.97	99.54	99.55	99.94	99.41	98.86	99.33	98.97	98.52
Fe	0	0	0	0	0	0	0.001	0	0	0.0
Na	0.8715	0.95	0.87	0.866	0.867	0.863	0.87	0.858	0.856	0.853
Si	2.915	3.003	2.921	2.912	2.915	2.929	2.925	2.906	2.905	2.903
Al	1.089	0.999	1.0865	1.095	1.089	1.079	1.083	1.103	1.103	1.107
Mg	0	0	0	0	0	0	0.001	0	0	0.0
K	0.006	0.008	0.006	0.004	0.0085	0.006	0.005	0.006	0.004	0.005
Ca	0.097	0.0145	0.09	0.097	0.097	0.089	0.084	0.1	0.104	0.104
Ba	0	0.001	0	0.001	0	0	0	0.001	0.001	0.001
Xab	0.894	0.977	0.901	0.895	0.891	0.901	0.907	0.89	0.888	0.887
Xan	0.1	0.015	0.093	0.1	0.1	0.093	0.088	0.104	0.108	0.108
Xor	0.006	0.008	0.006	0.004	0.009	0.006	0.005	0.006	0.004	0.005
Sample Anal no. descript	PK-31c 16 plag c sg 15	PK-38c 1 skeletal plag	PK-38c 2 skeletal plag	PK-42a 1 skeletal plag	PK-42a 2 skeletal plag	PK-42a 4 plag r	PK-42a 5 replplag c sg 4	PK-42a 3 replmyr r	PK-42a 4 myr m sg 2	PK-42a 5 myr c sg 2
Fe2O3	0.03	0	0.04	0.04	0	0.01	0.04	0.06	0	0.02
Na2O	10.06	11.25	11.23	11.14	10.67	11.37	11.3	10.38	10.02	10.1
SiO2	65.95	68.74	67.85	67.67	67.67	68.51	68.48	66.25	65.98	66.14
Al2O3	20.91	19.79	19.65	20.02	19.99	19.58	19.72	20.82	21.35	21.5
MgO	0	0.01	0	0	0	0	0	0.01	0	0.0
K2O	0.09	0.13	0.11	0.13	0.77	0.08	0.1	0.06	0.07	0.17
CaO	2.04	0.45	0.41	0.65	0.69	0.53	0.56	1.74	2.19	2.21
BaO	0.01	0	0	0.02	0	0	0	0	0	0.0
Tot.	99.09	100.37	99.29	99.67	99.79	100.08	100.19	99.32	99.6	100.14
Fe	0.001	0	0.001	0.001	0	0	0.001	0.002	0	0.001
Na	0.862	0.948	0.957	0.947	0.908	0.962	0.955	0.888	0.855	0.858
Si	2.916	2.989	2.984	2.969	2.97	2.99	2.986	2.922	2.903	2.898
Al	1.09	1.014	1.019	1.035	1.034	1.007	1.013	1.082	1.107	1.11
Mg	0	0.001	0	0	0	0	0	0.001	0	0.0
K	0.005	0.007	0.64	0.007	0.043	0.004	0.005	0.004	0.004	0.01
Ca	0.097	0.021	0.019	0.031	0.032	0.025	0.026	0.082	0.103	0.104
Ba	0	0	0	0	0	0	0	0	0	0.0
Xab	0.895	0.971	0.974	0.961	0.923	0.971	0.968	0.917	0.889	0.883
Xan	0.1	0.021	0.02	0.031	0.033	0.025	0.027	0.084	0.107	0.107
Xor	0.005	0.008	0.006	0.008	0.044	0.004	0.005	0.004	0.004	0.01
Sample Anal no. descript	PK-42a 6 myr cont pl sg 2	PK-42a 7 plag sg 2	PK-42a 8 plag sg 2	PK-42a 9 myr r sg 2	PK-42a 14 myr r	PK-42a 2 myr	PK-42a 3 plag sg 2	PK-42a 6 plag c	PK-42a 7 plag r	PK-42a 8 myr
Fe2O3	0	0	0.04	0	0	0	0	0.01	0.05	0.0
Na2O	10.63	9.95	9.48	10.88	9.88	10.12	10.13	10.12	10.22	10.34
SiO2	68.03	66.23	66.44	68.72	65.87	66.42	66.16	66.01	66.64	67.17
Al2O3	20.3	21.45	21.75	19.94	21.38	21.28	21.44	21.47	21.25	20.97
MgO	0	0	0.01	0	0	0.01	0	0	0	0.01
K2O	0.11	0.19	0.29	0.1	0.17	0.17	0.1	0.17	0.16	0.12
CaO	1.14	2.34	2.5	0.68	2.21	2.34	2.32	2.44	2.48	2.1
BaO	0.04	0.05	0.02	0	0.01	0.06	0	0	0	0.07
Tot.	100.27	100.21	100.53	100.32	99.51	100.41	100.15	100.21	100.81	100.77
Fe	0	0	0.001	0	0	0	0	0	0.002	0.0
Na	0.898	0.845	0.802	0.917	0.844	0.858	0.86	0.86	0.863	0.872
Si	2.964	2.9	2.897	2.987	2.902	2.904	2.899	2.893	2.904	2.9235
Al	1.043	1.107	1.118	1.022	1.1	1.097	1.107	1.109	1.092	1.0755
Mg	0	0	0.001	0	0	0.001	0	0	0	0.0
K	0.006	0.01	0.016	0.006	0.01	0.009	0.006	0.01	0.009	0.007
Ca	0.053	0.11	0.117	0.032	0.104	0.11	0.109	0.1145	0.116	0.098
Ba	0.001	0.001	0	0	0	0.001	0	0	0	0.001
Xab	0.938	0.875	0.858	0.961	0.881	0.879	0.883	0.874	0.874	0.893
Xan	0.055	0.114	0.125	0.033	0.109	0.112	0.112	0.116	0.117	0.1
Xor	0.007	0.011	0.017	0.006	0.01	0.009	0.006	0.01	0.009	0.007

Sample Anal no. descrpt	PK-43b 1 myr c	PK-43b 2 myr r sg 1	PK-43b 3 myr m sg 1	PK-43b 4 myr c	PK-43b 5 myr r sg 4	PK-43b 6 plag c	PK-43b 7 plag r sg 6	PK-43b 8 plag c	PK-43b 9 plag	PK-58 2 plag
Fe2O3	0	0	0.03	0	0.04	0.04	0	0.03	0	0.0
Na2O	10.41	10.28	10.27	10.3	10.29	10.03	10.16	9.93	10.21	10.54
SiO2	66.15	66.7	65.58	65.89	65.88	64.9	65.25	64.73	65.51	66.94
Al2O3	20.96	20.83	21.17	21.08	21.04	21.77	21.28	21.79	21.53	20.58
MgO	0.01	0	0.01	0	0	0	0.01	0.01	0	0.01
K2O	0.19	0.22	0.13	0.17	0.19	0.21	0.16	0.11	0.2	0.07
CaO	2.01	1.88	2.11	2.14	2.09	2.84	2.4	2.7	2.39	1.54
BaO	0.08	0.02	0.01	0	0	0	0.01	0.02	0	0
Tot.	99.81	99.93	99.3	99.58	99.52	99.79	99.26	99.31	99.85	99.67
Fe	0	0	0.001	0	0.001	0.0015	0	0.001	0	0.0
Na	0.889	0.874	0.88	0.88	0.88	0.858	0.872	0.852	0.872	0.897
Si	2.911	2.926	2.9	2.905	2.906	2.865	2.889	2.867	2.884	2.94
Al	1.087	1.077	1.103	1.095	1.094	1.1325	1.11	1.137	1.1175	1.066
Mg	0.001	0	0	0	0	0	0.001	0.001	0	0
K	0.011	0.012	0.007	0.009	0.011	0.012	0.009	0.006	0.011	0.004
Ca	0.0945	0.088	0.1	0.101	0.099	0.134	0.114	0.128	0.113	0.073
Ba	0.001	0	0	0	0	0	0	0	0	0
Xab	0.894	0.897	0.891	0.889	0.889	0.855	0.877	0.864	0.875	0.921
Xan	0.095	0.09	0.101	0.102	0.1	0.134	0.114	0.13	0.113	0.075
Xor	0.011	0.013	0.008	0.009	0.011	0.012	0.009	0.006	0.012	0.004
Sample Anal no. descrpt	PK-58 4 plag	PK-58 1 myr r	PK-58 2 myr c sg 1	PK-58 3 myr m sg 1	PK-58 4 myr r sg 1	PK-58 5 myr c sg 1	PK-58 6 myr r sg 1	PK-58 4 myr cont ksp 3	PK-58 6 myr c sg 4	PK-58 7 myr r sg 4
Fe2O3	0	0	0	0	0.01	0.01	0.02	0	0.02	0.03
Na2O	10.3	11.29	10.57	10.42	10.21	10.15	10.77	10.56	10.2	10.51
SiO2	66.42	68.29	66.89	66.33	66.33	66.59	67.58	67.71	66.67	67.74
Al2O3	20.75	19.55	20.56	20.58	20.73	20.85	20.74	20.25	21.32	20.14
MgO	0	0	0	0	0	0.01	0	0	0	0.02
K2O	0.09	0.07	0.1	0.12	0.13	0.18	0.15	0.12	0.24	0.15
CaO	1.87	0.6	1.8	2.03	2.01	2.25	1.41	1.47	2.28	1.43
BaO	0	0	0.02	0.01	0.02	0	0.05	0	0.01	0.09
Tot.	99.91	99.8	99.94	99.49	99.43	100.04	100.21	100.11	100.75	100.14
Fe	0	0	0	0	0	0	0.0005	0	0.001	0.001
Na	0.881	0.958	0.899	0.891	0.872	0.863	0.913	0.894	0.862	0.893
Si	2.926	2.989	2.934	2.925	2.924	2.92	2.954	2.958	2.906	2.961
Al	1.078	1.008	1.063	1.07	1.077	1.078	1.0425	1.043	1.095	1.038
Mg	0	0	0	0	0	0.001	0	0	0	0.001
K	0.005	0.004	0.0055	0.0065	0.007	0.01	0.008	0.007	0.014	0.008
Ca	0.089	0.028	0.085	0.096	0.095	0.106	0.066	0.069	0.106	0.067
Ba	0	0	0	0	0	0	0.001	0	0	0.002
Xab	0.904	0.967	0.909	0.897	0.895	0.882	0.925	0.922	0.878	0.922
Xan	0.091	0.028	0.086	0.096	0.097	0.108	0.067	0.071	0.108	0.069
Xor	0.005	0.004	0.005	0.007	0.007	0.01	0.008	0.007	0.014	0.009
Sample Anal no. descrpt	PK-58 8 myr c sg 4	PK-58 11 plag c	PK-58 12 plag r sg 11							
Fe2O3	0.01	0	0.04							
Na2O	10.22	10.2	10.21							
SiO2	66.62	66.65	66.76							
Al2O3	21.21	21.04	21.28							
MgO	0	0	0.01							
K2O	0.14	0.25	0.18							
CaO	2.37	2.32	2.33							
BaO	0.01	0.1	0							
Tot.	100.58	100.56	100.81							
Fe	0	0	0.001							
Na	0.865	0.864	0.862							
Si	2.908	2.912	2.907							
Al	1.091	1.0835	1.092							
Mg	0	0	0.001							
K	0.008	0.014	0.01							
Ca	0.111	0.109	0.109							
Ba	0	0.002	0							
Xab	0.879	0.876	0.879							
Xan	0.113	0.11	0.111							
Xor	0.008	0.014	0.01							

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Sample anal # descript	PK-17c 1 myr r	PK-17c 2 myr c sq 1	PK-17c 3 plag sg 1	PK-17c 4 plag sg 1	PK-17c 5 plag c	PK-17c 6 plag r sg 5	PK-17c 7 plag c	PK-17d 1 ragged repl	PK-17d 2 ragged repl sg 1	PK-17d 3 ragged repl sg 1
Fe2O3	0	0.02	0.02	0	0.01	0	0	0	0.02	0.08
Na2O	11.28	10.64	11.03	9.85	10.12	10	9.98	11.18	11.44	11.24
SiO2	67.99	66.33	67.82	65.75	65.6	65.54	65.52	68.04	68.91	68.41
Al2O3	19.59	20.41	19.94	21.27	21.46	21.03	21.59	20.16	19.44	19.14
MgO	0.01	0	0	0.01	0	0	0	0	0	0
K2O	0.1	0.13	0.2	0.26	0.06	0.05	0.18	0.11	0.09	0.2
CaO	0.54	1.35	0.67	2.24	2.24	2	2.46	0.91	0.53	0.14
BaO	0	0	0.05	0.04	0.02	0.06	0.02	0	0	0.02
Tot.	99.51	98.88	99.74	99.43	99.51	98.69	99.76	100.4	100.63	99.23
Fe	0	0.001	0.001	0	0	0	0	0	0.001	0.003
Na	0.96	0.914	0.938	0.843	0.865	0.861	0.852	0.945	0.963	0.958
Si	2.985	2.938	2.973	2.902	2.893	2.91	2.885	2.9645	2.991	3.008
Al	1.014	1.065	1.03	1.1065	1.115	1.101	1.121	1.035	1.005	0.992
Mg	0.001	0	0	0.001	0	0	0	0	0	0
K	0.006	0.007	0.011	0.015	0.003	0.003	0.01	0.006	0.005	0.011
Ca	0.025	0.064	0.031	0.106	0.106	0.095	0.116	0.042	0.0245	0.0065
Ba	0	0	0.001	0.001	0	0.001	0	0	0	0
Xab	0.969	0.927	0.957	0.875	0.888	0.898	0.871	0.951	0.97	0.982
Xan	0.025	0.065	0.032	0.11	0.109	0.099	0.119	0.043	0.025	0.007
Xor	0.006	0.008	0.011	0.015	0.003	0.003	0.01	0.006	0.005	0.011
Sample anal # descript	PK-17d 4 ragged repl sg 1	PK-17d 6 carlsbad tw	PK-17d 7 carlsbad tw	PK-17d 8 carlsbad tw	PK-17d 9 ragged repl	PK-17e 1 plag r	PK-17e 2 plag c sg 1	PK-17e 3 plag c sg 1	PK-17e 4 plag r	PK-17e 5 plag c sg 4
Fe2O3	0.06	0.04	0	0.03	0.03	0.06	0.03	0.01	0.03	0
Na2O	11.41	11.66	11.53	11.33	11.52	10.87	11.31	10.56	10.47	10.17
SiO2	68.92	68.45	69.75	69.27	68.96	66.71	67.86	66	65.93	66.01
Al2O3	19.37	19.51	19.72	19.8	19.72	20.64	20.03	21.19	20.96	21.36
MgO	0	0.02	0	0.01	0	0	0	0	0	0
K2O	0.05	0.2	0.11	0.18	0.12	0.07	0.03	0.07	0.24	0.35
CaO	0.18	0.06	0.11	0.31	0.39	1.58	0.6	2.31	2.05	2.36
BaO	0.03	0.04	0	0.01	0.08	0.13	0	0.01	0	0
Tot.	100.01	99.98	101.23	100.94	100.82	100.05	99.86	100.16	99.68	100.24
Fe	0.002	0.001	0	0.001	0.001	0.02	0.001	0.0005	0.001	0
Na	0.964	0.988	0.963	0.95	0.968	0.925	0.96	0.899	0.895	0.864
Si	3.005	2.993	3.004	2.995	2.99	2.927	2.97	2.897	2.907	2.895
Al	0.996	1.005	1.001	1.009	1.008	1.067	1.033	1.096	1.089	1.104
Mg	0	0.0015	0	0.001	0	0	0	0	0	0
K	0.003	0.011	0.006	0.01	0.007	0.004	0.002	0.004	0.0135	0.0195
Ca	0.008	0.003	0.005	0.014	0.018	0.074	0.028	0.109	0.097	0.111
Ba	0.0005	0.001	0	0	0.001	0.002	0	0	0	0
Xab	0.989	0.986	0.989	0.975	0.975	0.922	0.97	0.888	0.89	0.869
Xan	0.008	0.003	0.005	0.015	0.018	0.074	0.029	0.108	0.096	0.111
Xor	0.003	0.011	0.006	0.01	0.007	0.004	0.002	0.004	0.014	0.02
Sample anal # descript	PK-17e 6 plag	PK-17e 7 plag c	PK-17e 8 plag r sg 7	PK-31a 3 plag c	PK-31a 4 plag r sg 3	PK-31a 4 plag sg 3	PK-31a 4 incl repl sg 3	PK-31a 9 untw r	PK-31a 10 untw c sg 9	PK-31a 11 untw m sg 9
Fe2O3	0	0.02	0.01	0	0	0.03	0	0	0	0
Na2O	10.48	10.13	10.64	10.27	10	11.1	11.04	11.05	10.22	10.08
SiO2	66.25	65.5	65.91	66.09	65.9	67.97	68.28	68.46	66.42	66.44
Al2O3	21.39	21.26	20.95	21.1	21.11	19.9	19.94	19.74	21.09	21.14
MgO	0	0	0.01	0	0	0	0.01	0	0	0
K2O	0.11	0.28	0.15	0.12	0.07	0.13	0.08	0.07	0.08	0.06
CaO	2.28	2.51	2	2.03	2.08	0.57	0.54	0.41	1.85	1.73
BaO	0.04	0	0	0	0.02	0.01	0	0	0	0.04
Tot.	100.54	99.7	99.68	99.61	99.19	99.7	99.89	99.74	99.66	99.48
Fe	0	0.001	0	0	0	0.001	0	0	0	0
Na	0.888	0.867	0.91	0.876	0.857	0.943	0.936	0.936	0.87	0.855
Si	2.896	2.89	2.906	2.91	2.911	2.987	2.983	2.992	2.919	2.922
Al	1.102	1.106	1.0885	1.095	1.099	1.027	1.026	1.017	1.092	1.095
Mg	0	0	0.001	0	0	0	0	0	0	0
K	0.006	0.016	0.009	0.007	0.004	0.008	0.005	0.004	0.004	0.004
Ca	0.107	0.1185	0.094	0.096	0.099	0.027	0.025	0.02	0.087	0.082
Ba	0.001	0	0	0	0	0	0	0	0	0.001
Xab	0.887	0.866	0.898	0.895	0.893	0.964	0.969	0.976	0.906	0.91
Xan	0.107	0.118	0.093	0.098	0.103	0.028	0.026	0.02	0.09	0.086
Xor	0.006	0.016	0.009	0.007	0.004	0.008	0.005	0.004	0.004	0.004

Sample anal # descript	PK-31a 1 cont perth 3	PK-31a 2 ragged repl 3	PK-31a 4 untw	PK-31a 5 untw	PK-31a 6 untw	PK-31a 1 myr c	PK-31a 2 myr r sg 1	PK-31b 1 plag m myr ?	PK-31b 2 plag r sg 1	PK-31b 3 plagr
Fe203	0.02	0	0	0.03	0	0	0.03	0.01	0	0
Na2O	10.43	11.22	10.26	10.15	10.66	10.13	10.07	11.02	9.71	10.01
SiO2	66.46	67.93	65.99	65.09	66.45	66.47	66.15	68.56	70.91	66.19
Al2O3	20.46	19.73	20.91	21.44	20.76	20.82	21.19	19.85	18.48	21.09
MgO	0	0	0	0.01	0	0	0	0.01	0	0
K2O	0.22	0.07	0.13	0.1	0.06	0.1	0.08	0.19	0.08	0.12
CaO	1.73	0.63	2.17	2.22	1.77	1.85	1.98	0.41	1.14	2.07
BaO	0	0.03	0.05	0	0.01	0	0.03	0.02	0	0.02
Tot.	99.32	99.61	99.52	99.03	99.71	99.39	99.32	100.07	100.32	99.5
Fe	0.001	0	0	0.001	0	0	0.001	0	0	0
Na	0.893	0.954	0.878	0.872	0.909	0.865	0.86	0.931	0.814	0.855
Si	2.934	2.98	2.911	2.886	2.923	2.928	2.912	2.989	3.063	2.914
Al	1.064	1.02	1.087	1.12	1.076	1.082	1.1	1.02	0.9411	1.094
Mg	0	0	0	0	0	0	0	0	0	0
K	0.012	0.004	0.0075	0.0055	0.003	0.005	0.004	0.01	0.004	0.007
Ca	0.082	0.03	0.103	0.106	0.084	0.087	0.093	0.02	0.053	0.098
Ba	0	0.0005	0.001	0	0	0	0	0	0	0
Xab	0.905	0.966	0.889	0.887	0.913	0.904	0.898	0.969	0.935	0.891
Xan	0.083	0.03	0.104	0.107	0.084	0.091	0.098	0.02	0.061	0.102
Xor	0.013	0.004	0.007	0.006	0.003	0.005	0.004	0.011	0.004	0.007
Sample anal # descript	PK-31b 4 plag c sg 3	PK-33a 1 plag r	PK-33a 2 plag m	PK-33a 4 plag m	PK-33b 1 plag r	PK-33b 2 plag m sg 1	PK-33b 3 plag c sg 1	PK-33b 4 plag r sg 1	PK-33b 5 plag r	PK-33b 6 plagm sg 5
Fe203	0	0	0.01	0.01	0.07	0.04	0.01	0.01	0.04	0
Na2O	10.07	10.25	10.51	10.12	10.78	10.62	10.43	10.94	10.78	10.46
SiO2	65.99	65.79	66.09	66.07	67.1	66.02	66.06	67.28	66.95	66.17
Al2O3	21.34	21.53	21.2	21.45	20.83	21.51	21.63	0.2069	20.73	21.76
MgO	0	0	0	0	0	0	0	0	0	0
K2O	0.27	0.23	0.09	0.24	0.23	0.2	0.2	0.35	0.08	0.09
CaO	2.2	2.27	1.98	2.21	1.2	2.16	2.19	0.8	1.53	2.19
BaO	0.05	0	0	0	0.03	0	0	0	0	0
Tot.	99.92	100.08	99.89	100.1	100.23	100.55	100.52	100.08	100.11	100.68
Fe	0	0	0	0	0.002	0.001	0	0	0	0
Na	0.858	0.873	0.896	0.861	0.9135	0.901	0.884	0.928	0.915	0.885
Si	2.9	2.889	2.904	2.897	2.933	2.888	2.888	2.9435	2.931	2.887
Al	1.1055	1.114	1.098	1.109	1.073	1.109	1.115	1.067	1.0695	1.119
Mg	0	0	0	0	0	0	0	0	0	0
K	0.015	0.013	0.005	0.013	0.013	0.011	0.011	0.02	0.005	0.052
Ca	0.103	0.107	0.093	0.104	0.056	0.101	0.1025	0.037	0.072	0.1025
Ba	0.01	0	0	0	0.0005	0	0	0	0	0
Xab	0.879	0.879	0.901	0.88	0.93	0.889	0.886	0.942	0.923	0.891
Xan	0.106	0.108	0.094	0.106	0.057	0.1	0.103	0.038	0.072	0.103
Xor	0.016	0.013	0.005	0.014	0.013	0.011	0.011	0.02	0.005	0.005
Sample anal # descript	PK-33b 7 plag c sg 5	PK-38a 1 repl v c	PK-38a 2 repl v sg 1 r	PK-38a 3 repl v	PK-38a 4 repl v	PK-38a 6 ragged repl	PK-38a 7 carlsbad	PK-43a 8 plag r	PK-43a 9 plag c sg 8	PK-43a 1 plagc
Fe203	0	0	0	0.01	0	0	0.01	0.01	0	0
Na2O	10.48	11.47	11.51	11.42	11.67	11.56	11.59	10.17	9.92	10.24
SiO2	66.5	68.33	68.32	68.09	68.91	68.61	68.62	65.77	67.57	66.05
Al2O3	21.54	19.5	19.43	19.52	19.63	19.57	19.53	21.03	20.67	21.24
MgO	0	0	0	0.01	0	0	0	0	0	0
K2O	0.05	0.11	0.09	0.12	0.08	0.07	0.13	0.2	0.26	0.21
CaO	2.14	0.31	0.35	0.44	0.28	0.35	0.36	2.52	2.49	2.19
BaO	0.11	0.02	0.01	0	0	0.01	0.01	0	0.02	0.08
Tot.	100.82	99.75	99.71	99.6	100.56	100.18	100.25	99.7	98.94	100.02
Fe	0	0	0	0	0	0	0	0	0	0
Na	0.885	0.974	0.978	0.972	0.983	0.978	0.98	0.869	0.854	0.872
Si	2.897	2.992	2.993	2.987	2.993	2.992	2.992	2.9	2.912	2.902
Al	1.106	1.007	1.003	1.009	1.005	1.006	1.004	1.093	1.082	1.1
Mg	0	0	0	0.001	0	0	0	0	0	0
K	0.003	0.006	0.005	0.007	0.004	0.004	0.007	0.011	0.015	0.012
Ca	0.1	0.015	0.016	0.021	0.013	0.016	0.017	0.119	0.1185	0.103
Ba	0.002	0	0	0	0	0	0	0	0	0.001
Xab	0.896	0.979	0.979	0.973	0.983	0.98	0.976	0.87	0.865	0.884
Xan	0.101	0.015	0.016	0.021	0.013	0.016	0.017	0.119	0.12	0.105
Xor	0.003	0.006	0.005	0.007	0.004	0.004	0.007	0.011	0.015	0.012

Sample anal # descript	PK-43a 2 myr c	PK-43a 3 myr r sg 1	PK-43a 4 myr m sg 1	PK-43a 5 myr r sg 1	PK-43a 1 myr	PK-57 3 incl sg 1	PK-57 4 cont	PK-57 6 ex? repl?	PK-57 7 incl	PK-57 8 r sg 7
Fe2O3	0.01	0	0	0	0.05	0	0.04	0.04	0.02	0.01
Na2O	10.08	10.81	10.04	9.33	10.33	11.02	11.73	11.56	11071	11.4
SiO2	65.99	67.45	65.91	68.2	66.03	67.57	67.8	67.89	68.46	67.97
Al2O3	20.91	19.99	20.77	18.97	21.04	19.41	19.64	19.23	19.4	19.22
MgO	0	0	0	0	0.02	0.01	0.01	0	0	0
K2O	0.18	0.12	0.15	0.14	0.14	0.11	0.09	0.12	0.17	0.1
CaO	2.38	1.19	2.17	2.04	2.08	0.85	0.61	0.44	0.55	0.51
BaO	0.01	0.03	0	0	0.01	0.07	0	0.03	0	0.08
Tot.	99.56	99.59	99.03	98.68	99.7	99.04	99.93	99.29	100.3	99.3
Fe	0	0	0	0	0.002	0	0.001	0.001	0.001	0.0005
Na	0.862	0.9205	0.862	0.798	0.8815	0.9435	0.997	0.987	0.991	0.9737
Si	2.91	2.963	2.918	3.01	2.907	2.9835	2.972	2.991	2.988	2.994
Al	1.087	1.035	1.084	0.987	1.092	1.01	1.015	0.998	0.998	0.998
Mg	0	0	0	0	0.001	0.001	0.001	0	0	0
K	0.01	0.007	0.009	0.008	0.008	0.006	0.005	0.007	0.009	0.005
Ca	0.112	0.056	0.103	0.097	0.098	0.04	0.029	0.0205	0.026	0.024
Ba	0	0.0005	0	0	0	0.001	0	0.0005	0	0.001
Xab	0.876	0.936	0.886	0.884	0.893	0.953	0.968	0.973	0.966	0.971
Xan	0.114	0.057	0.106	0.107	0.099	0.041	0.028	0.02	0.025	0.024
Xor	0.01	0.007	0.009	0.009	0.008	0.006	0.005	0.007	0.009	0.005
Sample anal # descript	PK-57 2 ex? repl?									
Fe2O3	0									
Na2O	11.52									
SiO2	67.95									
Al2O3	19.52									
MgO	0									
K2O	0.09									
CaO	0.48									
BaO	0.07									
Tot.	99.63									
Fe	0									
Na	0.9805									
Si	2.984									
Al	1.01									
Mg	0									
K	0.005									
Ca	0.022									
Ba	0.001									
Xab	0.973									
Xan	0.022									
Xor	0.005									

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Sample anal # descript	PK-36 8 repl cass	PK-36 12 cont 11	PK-36 1 sg 8	PK-36 2 sg 7	PK-36 3 myr sg 12	PK-36 4 repl myr cass	PK-36 5 repl myr cass	PK-36 7 repl cass	PK-36 9 myr	PK-36 5 incl cass
Fe2O3	0	0.03	0.01	0	0	0	0	0.01	0	0.04
Na2O	9.48	9.36	10.13	10.19	10.26	9.85	9.66	9.42	9.62	11.06
SiO2	65.33	65	65.84	65.81	65.7	65.9	65.06	64.91	65.68	68.46
Al2O3	21.11	21.34	21.68	21.46	21.94	21.4	20.98	21.43	21.21	19.51
MgO	0	0	0	0	0	0	0	0	0	0.01
K2O	0.19	0.21	0.2	0.12	0.15	0.11	0.1	0.18	0.24	0.12
CaO	2.54	2.64	2.62	2.4	2.73	2.44	2.2	2.28	2.19	0.6
BaO	0	0.03	0	0	0.05	0	0.023	0.05	0	0
Tot.	98.68	98.62	100.47	99.99	100.84	99.69	98.03	98.29	98.95	99.8
Fe	0	0.001	0	0	0	0	0	0	0	0.001
Na	0.817	0.807	0.86	0.868	0.869	0.84	0.837	0.814	0.826	0.938
Si	2.902	2.892	2.882	2.891	2.869	2.899	2.907	2.894	2.908	2.994
Al	1.107	1.119	1.0118	1.111	1.129	1.109	1.105	1.126	1.107	1.006
Mg	0	0	0	0	0	0	0	0	0	0.001
K	0.011	0.012	0.011	0.007	0.0085	0.006	0.006	0.01	0.013	0.0065
Ca	0.121	0.126	0.123	0.113	0.1275	0.115	0.105	0.109	0.104	0.028
Ba	0	0.0005	0	0	0.001	0	0	0.001	0	0
Xab	0.861	0.854	0.865	0.879	0.865	0.874	0.883	0.872	0.876	0.964
Xan	0.127	0.133	0.124	0.115	0.127	0.1195	0.111	0.117	0.11	0.007
Xor	0.011	0.013	0.011	0.007	0.008	0.006	0.006	0.011	0.014	0.029

Sample anal # descript	PK-36 6 incl cass	PK-36 7 incl cass	PK-36 8 incl cass	PK-36 4 myr traverse	PK-36 6 myr traverse	PK-36 7 myr traverse	PK-36 1 myr traverse	PK-36 2 myr traverse	PK-36 3 myr traverse	PK-36 4 myr trav.
Fe2O3	0	0.04	0.03	0	0.04	0.07	0.01	0	0.04	0
Na2O	10.64	9.9	10.54	10.71	10.37	9.82	10.77	10.32	10.01	10.09
SiO2	67.73	65.99	67.09	65.17	63.65	63.48	67.28	66.41	66.09	65.88
Al2O3	19.67	20.89	20.09	20.7	21.53	21.78	20.84	21.42	21.76	21.54
MgO	0	0	0.01	0	0	0	0	0	0.01	0.02
K2O	0.09	0.12	0.1	0.13	0.13	0.26	0.18	0.18	0.16	0.15
CaO	1.19	2.63	1.55	1.42	2.31	2.68	1.57	2.33	2.52	2.4
BaO	0	0	0	0	0	0.08	0	0	0.02	0.01
Tot.	99.32	99.57	99.41	98.14	98.03	98.17	100.66	100.66	100.62	100.09
Fe	0	0.001	0.001	0	0.001	0.002	0	0	0.001	0
Na	0.907	0.846	0.8995	0.928	0.904	0.855	0.91	0.8735	0.847	0.8585
Si	2.979	2.9095	2.954	2.914	2.86	2.851	2.93	2.898	2.885	2.8905
Al	1.02	1.085	1.042	1.091	1.14	1.153	1.07	1.102	1.12	1.114
Mg	0	0	0.0001	0	0	0	0	0	0.001	0.001
K	0.005	0.007	0.006	0.007	0.008	0.015	0.01	0.01	0.009	0.008
Ca	0.056	0.124	0.073	0.068	0.111	0.129	0.073	0.109	0.118	0.113
Ba	0	0	0	0	0	0.001	0	0	0	0
Xab	0.937	0.866	0.919	0.925	0.884	0.856	0.916	0.88	0.87	0.876
Xan	0.058	0.127	0.006	0.068	0.109	0.129	0.074	0.11	0.121	0.115
Xor	0.005	0.007	0.075	0.007	0.007	0.015	0.01	0.01	0.009	0.009
Sample anal # descript	PK-47a 1 myr r	PK-47a 2 myr c sg 1	PK-47a 4 myr r	PK-47a 5 myr c sg 4	PK-47a 6 plag	PK-47a 7 plag c	PK-47a 8 plag r sg 7	PK-47a 9 plag c	PK-52 1 plag cass	PK-52 2 repl sg 1
Fe2O3	0.04	0.02	0.06	0	0.04	0.01	0.06	0	0	0
Na2O	10.94	10.82	10.76	11.17	11.05	11.25	11.2	11.36	10.02	9.97
SiO2	68.74	68.74	68.84	69.22	69.48	69.11	69.58	69.17	65.86	65.28
Al2O3	19.62	20.04	19.61	19.56	19.54	19.5	19.23	19.2	21.14	21.53
MgO	0	0	0	0	0.01	0	0.02	0	0	0
K2O	0.09	0.11	0.11	0.08	0.09	0.06	0.06	0.09	0.09	0.08
CaO	0.62	0.88	0.72	0.49	0.51	0.45	0.24	0.26	2.18	2.56
BaO	0	0	0.09	0.03	0	0.02	0	0.03	0	0.12
Tot.	100.05	100.62	100.19	100.57	100.72	100.41	100.39	100.11	99.29	99.54
Fe	0.001	0.001	0.002	0	0.001	0.0005	0.002	0	0	0
Na	0.924	0.91	0.908	0.939	0.927	0.947	0.942	0.96	0.858	0.854
Si	2.996	2.981	2.997	3.0015	3.006	3.002	3.019	3.013	2.9075	2.883
Al	1.008	1.025	1.006	1.001	0.997	0.998	0.983	0.986	1.1	1.121
Mg	0	0	0	0	0.001	0	0.001	0	0	0
K	0.005	0.006	0.006	0.005	0.005	0.003	0.003	0.005	0.005	0.004
Ca	0.029	0.041	0.034	0.023	0.023	0.021	0.011	0.012	0.103	0.121
Ba	0	0	0.002	0.0005	0	0	0	0.0005	0	0.002
Xab	0.965	0.951	0.958	0.972	0.97	0.975	0.985	0.983	0.888	0.872
Xan	0.03	0.043	0.035	0.024	0.025	0.021	0.012	0.012	0.107	0.124
Xor	0.005	0.006	0.006	0.005	0.005	0.004	0.003	0.005	0.005	0.004
Sample anal # descript	PK-52 5 cont cass	PK-52 6 plag c	PK-52 7 plag r sg 6	PK-52 9 plag	PK-59 1 plag c	PK-59 2 cont cass	PK-59 3 cont cass	PK-59 5 repl cass	PK-59 6 repl cass	PK-59 7 cont cass
Fe2O3	0.05	0.05	0	0	0	0.05	0.05	0.04	0.05	0
Na2O	10.02	10.36	10.02	10.26	10.3	9.92	9.93	10.11	9.95	9.68
SiO2	65.27	66.04	65.24	65.62	66.03	65.9	65.01	65.72	65.88	65.72
Al2O3	21.6	21.46	21.65	21.67	21.87	21.95	21.15	21.71	21.73	21.83
MgO	0	0	0	0	0	0	0.03	0	0	0
K2O	0.18	0.22	0.15	0.13	0.19	0.11	0.4	0.1	0.17	0.11
CaO	2.53	2.36	2.49	2.39	2.53	2.68	2.15	2.45	2.44	2.76
BaO	0.05	0	0	0	0	0.05	0	0.08	0	0
Tot.	99.7	100.49	99.56	100.07	100.91	100.67	98.73	100.22	100.22	100.11
Fe	0.002	0.002	0	0	0	0.0015	0.002	0.0015	0.002	0
Na	0.857	0.879	0.8579	0.8735	0.87	0.84	0.857	0.86	0.845	0.823
Si	2.879	2.89	2.8794	2.882	2.878	2.877	2.894	2.882	2.886	2.881
Al	1.123	1.107	1.126	1.122	1.123	1.1295	1.11	1.122	1.122	1.128
Mg	0	0	0	0	0	0	0.002	0	0	0
K	0.01	0.012	0.009	0.007	0.0105	0.006	0.023	0.006	0.01	0.006
Ca	0.12	0.1105	0.118	0.112	0.118	0.125	0.102	0.115	0.115	0.13
Ba	0.001	0	0	0	0	0.001	0	0.001	0	0
Xab	0.869	0.877	0.872	0.8795	0.871	0.864	0.872	0.877	0.872	0.858
Xan	0.121	0.11	0.12	0.113	0.118	0.129	0.104	0.117	0.118	0.135
Xor	0.01	0.012	0.009	0.007	0.011	0.007	0.023	0.006	0.01	0.007

Sample anal # descript	PK-59 8 plag c	PK-59 9 plag r sg 8	PK-59 10 plag c	PK-60b 1 plag incl in 2	PK-60b 1 myr r	PK-60b 2 myr c sg 1	PK-60b 3 myr r	PK-60b 4 myr c sg 3	PK-60b 5 myr c	PK-60b 6 myr c
Fe2O3	0	0	0.08	0.03	0.03	0.01	0	0	0.01	0.02
Na2O	10.13	10.1	10.59	11.55	11.19	11.19	11.22	11.24	11.19	11.3
SiO2	66.24	65.73	66.8	68.25	68.04	68.4	68.84	68.62	68.5	68.69
Al2O3	21.47	21.56	20.39	19.17	19.96	20.07	19.96	19.83	20.35	19.99
MgO	0	0	0.02	0	0	0	0	0	0	0.01
K2O	0.1	0.11	0.28	0.12	0.08	0.07	0.09	0.09	0.09	0.13
CaO	2.32	2.52	0.92	0.3	0.76	0.78	0.66	0.56	0.84	0.37
BaO	0.08	0	0.11	0	0.04	0.06	0.09	0	0	0.04
Tot.	100.32	100.02	99.18	99.4	100.1	100.59	100.85	100.35	100.97	100.54
Fe	0	0	0.003	0.001	0.001	0	0	0	0	0.001
Na	0.859	0.86	0.906	0.984	0.948	0.943	0.942	0.948	0.939	0.951
Si	2.898	2.887	2.949	3	2.972	2.973	2.983	2.986	2.965	2.983
Al	1.107	1.116	1.061	0.993	1.028	1.028	1.019	1.017	1.038	1.023
Mg	0	0	0.001	0	0	0	0	0	0	0
K	0.005	0.006	0.016	0.007	0.004	0.004	0.005	0.006	0.005	0.007
Ca	0.109	0.119	0.043	0.014	0.035	0.0365	0.031	0.026	0.039	0.017
Ba	0.001	0	0.002	0	0.001	0.0015	0	0	0	0.001
Xab	0.883	0.873	0.939	0.979	0.96	0.959	0.964	0.968	0.956	0.975
Xan	0.112	0.121	0.045	0.014	0.036	0.037	0.031	0.027	0.04	0.018
Xor	0.005	0.006	0.016	0.007	0.004	0.004	0.005	0.005	0.005	0.007
Sample anal # descript	PK-63 1 cont cass	PK-63 2 plag c sg 1	PK-63 3 repl cass	PK-63 4 myr	PK-63 5 myr	PK-63 6 plag	PK-65 2 plag r cloudy	PK-65 3 plag r cloudy	PK-65 5 incl cass	PK-65 6 cloudy cont cass
Fe2O3	0	0	0	0.01	0	0.03	0.06	0.05	0.03	0.04
Na2O	11.41	11.24	11.41	11.2	10.91	11.46	10.64	10.81	10.26	10.49
SiO2	69.51	69.55	69.14	69.27	67.23	69.12	68.07	68.27	66.74	66.96
Al2O3	19.63	19.49	19.6	19.67	19.92	19.47	20.1	19.96	20.8	20.63
MgO	0	0.01	0	0	0.01	0.01	0.01	0	0	0.02
K2O	0.08	0.07	0.07	0.08	0.17	0.06	0.24	0.17	0.23	0.3
CaO	0.28	0.21	0.2	0.26	0.32	0.26	0.73	0.54	1.52	0.86
BaO	0	0.07	0	0	0.04	0	0	0.01	0.03	0
Tot.	100.91	100.65	100.45	100.49	98.6	100.4	99.84	99.82	99.61	99.3
Fe	0	0	0.001	0	0	0.001	0.002	0.002	0.001	0.001
Na	0.956	0.944	0.96	0.9415	0.936	0.965	0.9015	0.916	0.875	0.895
Si	3.003	3.011	3.001	3.003	2.976	3.003	2.976	2.983	2.933	2.947
Al	1	0.995	1.0025	1.005	1.039	0.997	1.0355	1.028	1.077	1.07
Mg	0	0.001	0	0	0.0005	0.0001	0	0	0	0.0015
K	0.004	0.004	0.004	0.0045	0.009	0.003	0.013	0.01	0.013	0.017
Ca	0.013	0.01	0.009	0.012	0.015	0.012	0.034	0.0255	0.072	0.041
Ba	0	0.001	0	0	0.001	0	0	0	0.005	0
Xab	0.982	0.986	0.986	0.982	0.974	0.984	0.95	0.963	0.912	0.94
Xan	0.013	0.01	0.01	0.013	0.016	0.012	0.036	0.027	0.075	0.043
Xor	0.004	0.004	0.004	0.005	0.01	0.004	0.014	0.01	0.013	0.018
Sample anal # descript	PK-65 7 plag cloudy									
Fe2O3	0									
Na2O	9.89									
SiO2	66.17									
Al2O3	21.44									
MgO	0									
K2O	0.18									
CaO	2.63									
BaO	0.01									
Tot.	100.33									
Fe	0									
Na	0.839									
Si	2.896									
Al	1.106									
Mg	0									
K	0.01									
Ca	0.123									
Ba	0									
Xab	0.863									
Xan	0.127									
Xor	0.01									

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Sample anal. #	PK-12a 1	PK-12a 2	PK-12a 1	PK-14b 2	PK-14b 3	PK-14b 5	PK-14b 12	PK-35b 1	PK-35b 2	PK-35b 3
descript	plag	plag sg 1	plag sg 1	plag r cont 1	plag c sg 2	plag	plag	plag c	plag r sg 1	plag ser
Fe2O3	0	0	0	0.01	0	0.01	0.03	0.05	0.03	0.03
Na2O	11.16	10.99	11.16	11.06	11.14	10.92	10.94	11.96	11.46	11.6
SiO2	68.1	67.48	68.04	67.97	67.95	68.17	67.69	68.84	68.63	68.79
Al2O3	19.22	19.62	19.34	19.61	19.61	19.9	19.7	19.65	19.7	19.72
MgO	0	0	0.02	0	0.01	0	0	0	0	0
K2O	0.07	0.1	0.08	0.13	0.06	0.09	0.14	0.08	0.07	0.06
CaO	0.06	0.41	0.18	0.55	0.59	0.67	0.6	0.22	0.26	0.35
BaO				0	0.01	0	0.02	0	0.07	0
Tot.	98.62	98.62	99.07	99.34	99.38	99.76	99.13	100.8	100.23	100.55
Fe	0	0	0	0	0	0	0.01	0.0015	0.001	0.001
Na	0.956	0.943	0.955	0.942	0.949	0.926	0.935	1.007	0.9685	0.977
Si	3.008	2.985	3.002	2.987	2.986	2.982	2.982	2.987	2.99	2.9885
Al	1	1.023	1.006	1.016	1.016	1.026	1.023	1.005	1.012	1.01
Mg	0	0	0	0	0	0	0	0	0	0
K	0.004	0.005	0.005	0.0075	0.004	0.005	0.008	0.0045	0.004	0.003
Ca	0.003	0.02	0.008	0.026	0.028	0.032	0.028	0.01	0.012	0.0165
Ba				0	0	0	0	0	0.001	0
Xab	0.993	0.975	0.987	0.966	0.968	0.962	0.963	0.986	0.984	0.98
Xan	0.003	0.02	0.008	0.027	0.028	0.033	0.029	0.01	0.012	0.017
Xor	0.004	0.005	0.005	0.008	0.004	0.005	0.008	0.004	0.004	0.003
Sample anal. #	PK-35b 5	PK-35b 6	PK-47c 1	PK-47c 2	PK-47c 3	PK-47c 4	PK-47c 5	PK-50b 1	PK-50b 3	PK-50b 2
descript	plag ser	plag ser	untw r	untw c sg 1	cont cass	vug	vug	untw repl?	untw	rims cass
Fe2O3	0	0.01	0.04	0.02	0.01	0.04	0.02	0.05	0.01	0
Na2O	11.91	11.57	11.62	11.56	11.57	11.38	11.24	11.17	11.19	11.1
SiO2	68.79	68.49	68.03	68.46	68.55	68.52	68.6	68.56	68.17	69.08
Al2O3	19.65	20.01	19.46	19.34	19.11	19.56	19.29	19.72	19.55	19.23
MgO	0	0	0	0	0	0.01	0	0.01	0	0.01
K2O	0.08	0.04	0.06	0.09	0.07	0.08	0.09	0.09	0.08	0.07
CaO	0.13	0.57	0.29	0.2	0.17	0.3	0.27	0.46	0.34	0.14
BaO	0.04	0	0	0.07	0	0	0	0	0.05	0.03
Tot.	100.61	100.68	99.5	99.73	99.49	99.88	99.52	100.05	99.41	99.66
Fe	0	0	0.001	0.001	0	0.001	0.0005	0.001	0	0
Na	1.0035	0.974	0.99	0.982	0.984	0.964	0.9547	0.945	0.953	0.94
Si	2.989	2.974	2.988	2.999	3.007	2.994	3.006	2.99	2.993	3.018
Al	1.007	1.024	1.0075	0.998	0.988	1.007	0.996	1.014	1.013	0.99
Mg	0	0	0	0	0	0.001	0	0.001	0	0.001
K	0.0045	0.0025	0.003	0.005	0.004	0.005	0.005	0.005	0.005	0.004
Ca	0.006	0.0265	0.013	0.009	0.008	0.014	0.013	0.021	0.016	0.0065
Ba	0.001	0	0	0.001	0	0	0	0	0.001	0.0005
Xab	0.989	0.971	0.983	0.986	0.988	0.981	0.981	0.973	0.979	0.989
Xan	0.006	0.026	0.013	0.009	0.008	0.014	0.013	0.021	0.016	0.007
Xor	0.005	0.003	0.003	0.005	0.004	0.005	0.005	0.005	0.005	0.004
Sample anal. #	PK-60b 2	PK-60b 4	PK-60b 5							
descript	incl in ksp 1	incl cass	incl cass							
Fe2O3	0	0.02	0							
Na2O	11.44	11.12	11.4							
SiO2	68.1	67.52	68.44							
Al2O3	19.51	19.93	19.58							
MgO	0	0	0							
K2O	0.09	0.1	0.08							
CaO	0.36	0.94	0.57							
BaO	0	0	0							
Tot.	99.49	99.64	100.07							
Fe	0	0.001	0							
Na	0.974	0.947	0.965							
Si	2.99	2.965	2.988							
Al	1.009	1.032	1.008							
Mg	0	0	0							
K	0.005	0.006	0.0045							
Ca	0.017	0.044	0.026							
Ba	0	0	0							
Xab	0.978	0.95	0.969							
Xan	0.017	0.044	0.027							
Xor	0.005	0.006	0.004							

APLITE K-FELDSPAR

Sample Anal no. descript ksp	PK-1b 1 graphic	PK-1b 2 graphic	PK-1b 3 graphic	PK-1b 4 graphic	PK-3c 2 ksp cont	PK-3c 3 ksp cont	PK-3c 10 perth	PK-3c 11 perth	PK-3c 17 ksp repl	PK-25 4 untw
					1	gar			15, 16	
Fe2O3	0.01	0	0.01	0.02	0.01	0.04	0	0	0.03	0
Na2O	0.78	0.92	0.95	0.96	0.96	2.22	0.63	0.69	0.38	0.91
SiO2	65.21	65.09	65.25	64.89	65.67	65.71	65.1	65.27	64.87	64.81
Al2O3	18.34	18.45	18.27	18.14	18.28	18.18	18.27	18.22	18	18.26
MgO	0	0	0.01	0	0	0	0	0	0	0.01
K2O	16.5	16.4	16.28	16.16	16.05	14.22	16.65	16.58	16.94	16.29
CaO	0	0	0.01	0	0	0.06	0	0.01	0.02	0
BaO	0	0	0	0	0	0.05	0.13	0	0	0.01
Tot.	100.84	100.85	100.79	100.17	100.97	100.48	100.79	100.78	100.25	100.29
Fe	0	0	0	0.001	0.0005	0.001	0	0	0.001	0
Na	0.0695	0.0815	0.085	0.086	0.085	0.197	0.056	0.062	0.034	0.082
Si	2.996	2.99	2.997	2.998	3.0045	3.005	2.997	3.0005	3.003	2.9935
Al	0.993	0.999	0.989	0.998	0.986	0.98	0.991	0.987	0.982	0.994
Mg	0	0	0.001	0	0	0	0	0	0	0.0005
K	0.967	0.961	0.954	0.952	0.937	0.83	0.978	0.9725	1.001	0.96
Ca	0	0	0.001	0	0	0.003	0	0.001	0.001	0
Ba	0	0	0	0	0	0.001	0.002	0	0	0
Xab	0.067	0.078	0.081	0.083	0.083	0.191	0.054	0.06	0.033	0.079
Xan	0	0	0.001	0	0	0.003	0	0	0.001	0
Xor	0.933	0.922	0.918	0.917	0.917	0.806	0.946	0.94	0.966	0.921
Sample anal. # descript	PK-25 5 untw ksp	PK-25 6 untw ksp sg 5	PK-25 7 untw ksp sg 5	PK-25 10 mcr	PK-25 11 untw ksp sg 10	PK-25 12 ksp	PK-31c 4 ksp repl? 1,2,3	PK-31c 5 ksp repl?	PK-31c 6 perth	PK-38c 3 ksp
Fe2O3	0	0	0	0	0	0	0	0	0	0
Na2O	4.35	1.33	2.43	0.48	0.56	0.57	0.88	0.67	0.74	0.43
SiO2	66.13	64.88	65.21	64.7	65.04	64.87	65.14	65.45	65.11	65.28
Al2O3	18.65	18.31	18.43	18.3	18.25	18.25	18.17	18.2	18.23	18.32
MgO	0	0	0	0	0	0	0	0	0	0
K2O	11.11	15.7	13.98	16.88	16.75	16.82	16.43	16.56	16.35	16.91
CaO	0.03	0	0	0	0	0	0	0	0	0
BaO	0	0.01	0	0	0	0.02	0.03	0	0	0.03
Tot.	100.26	100.22	100.06	100.35	100.59	100.54	100.65	100.88	100.43	100.97
Fe	0	0	0	0	0	0	0	0	0	0
Na	0.382	0.1185	0.2165	0.043	0.05	0.051	0.0785	0.06	0.066	0.038
Si	2.996	2.993	2.993	2.992	2.998	2.9945	2.999	3.004	3	2.998
Al	0.996	0.995	0.997	0.9975	0.991	0.993	0.986	0.985	0.99	0.992
Mg	0	0	0	0	0	0	0	0	0	0
K	0.642	0.924	0.818	0.9955	0.985	0.991	0.965	0.97	0.961	0.991
Ca	0.001	0	0	0	0	0	0	0	0	0
Ba	0	0	0	0	0	0	0.001	0	0	0.005
Xab	0.373	0.114	0.209	0.041	0.048	0.049	0.075	0.058	0.064	0.037
Xan	0.001	0	0	0	0	0	0	0	0	0
Xor	0.626	0.886	0.791	0.959	0.952	0.951	0.925	0.942	0.936	0.963
Sample anal. # descript	PK-38c 4 ksp	PK-42a 3 ksp	PK-42a 6 cg ksp	PK-42a 2 graphic	PK-42a 12 ksp c cont 2	PK-42a 13 ksp r cont 2	PK-52 3 ksp repl cass	PK-52 4 ksp repl sg 3	PK-52 8 ksp	PK-58 3 ksp
Fe2O3	0.01	0	0	0.02	0.01	0	0.01	0.01	0.02	0
Na2O	0.6	0.57	0.75	0.68	0.86	0.62	0.75	0.7	0.69	0.61
SiO2	65.05	65	65.17	64.92	64.87	64.63	65.27	65.36	64.84	63.99
Al2O3	18.38	18.3	18.42	18.39	18.47	18.38	18.25	18.35	18.35	18.16
MgO	0	0	0	0	0	0	0	0	0	0.01
K2O	16.8	16.77	16.53	16.7	16.39	16.82	16.3	16.28	16.19	16.51
CaO	0	0	0	0	0	0	0	0	0	0
BaO	0	0	0.02	0.06	0.05	0.02	0.03	0.11	0.05	0
Tot.	100.84	100.65	100.89	100.79	100.65	100.47	100.6	100.82	100.15	99.31
Fe	0	0	0	0.001	0	0	0	0	0.001	0
Na	0.053	0.051	0.067	0.061	0.077	0.056	0.067	0.063	0.062	0.055
Si	2.992	2.995	2.993	2.989	2.987	2.987	3.0015	3	2.995	2.989
Al	0.9965	0.994	0.997	0.998	1.002	1.001	0.989	0.993	0.999	0.999
Mg	0	0	0	0	0	0	0	0	0	0
K	0.9861	0.986	0.968	0.981	0.963	0.992	0.956	0.953	0.954	0.984
Ca	0	0	0	0	0	0	0	0	0	0
Ba	0	0	0	0.001	0.001	0	0.0005	0.002	0.001	0
Xab	0.051	0.049	0.065	0.059	0.074	0.053	0.066	0.062	0.061	0.053
Xan	0	0	0	0	0	0	0	0	0	0
Xor	0.949	0.951	0.935	0.941	0.926	0.947	0.934	0.938	0.939	0.947

Sample anal. # descript	PK-58 3 ksp r myr 4	PK-58 9 ksp cont myr 4
Fe2O3	0	0.03
Na2O	1.33	0.6
SiO2	65.69	65.38
Al2O3	17.99	18.04
MgO	0	0
K2O	15.53	16.81
CaO	0	0
BaO	0.02	0
Tot.	100.55	100.84
Fe	0	0.001
Na	0.118	0.053
Si	3.0135	3.006
Al	0.9725	0.977
Mg	0	0
K	0.9087	0.986
Ca	0	0
Ba	0	0
Xab	0.115	0.051
Xan	0	0
Xor	0.885	0.949

PEGMATITE K-FELDSPAR

Sample anal. # descript	PK-17d 5 ksp	PK-17d 10 ksp	PK-31a 3 perth	PK-31a 6 perth sg 3	PK-31a 7 ksp	PK-31a 8 perth	PK-31a 3 perth	PK-31b 1 ksp r	PK-31b 2 ksp c sg 1	PK-36 1 ksp
Fe2O3	0.02	0.03	0	0	0.02	0	0.01	0	0.02	0
Na2O	0.81	0.59	0.79	0.62	0.62	0.66	0.97	0.69	0.63	0.69
SiO2	64.62	65.64	65.24	64.95	64.96	64.96	64.31	65.23	64.83	64.9
Al2O3	18.47	18.64	18.56	18.36	18.35	18.31	18.28	18.28	18.18	18.23
MgO	0.01	0	0	0	0	0	0	0.01	0	0.02
K2O	16.34	17.01	15.7	15.96	15.61	15.47	16.31	16.71	16.68	16.57
CaO	0	0	0	0.06	0	0	0	0	0	0
BaO	0.07	0.07	0	0	0.01	0.07	0.03	0	0	0.02
Tot.	100.33	101.97	100.3	99.95	99.57	99.48	99.92	100.93	100.36	100.43
Fe	0.001	0.001	0	0	0	0	0	0	0.001	0
Na	0.072	0.052	0.07	0.056	0.056	0.06	0.088	0.062	0.057	0.062
Si	2.985	2.988	2.998	3	3.004	3.008	2.986	2.997	2.996	2.996
Al	1.006	1	1.058	1	1.001	1	1	0.99	0.991	0.992
Mg	0.001	0	0	0	0	0	0	0.01	0	0.0015
K	0.963	0.988	0.92	0.94	0.921	0.914	0.966	0.979	0.984	0.976
Ca	0	0	0	0.003	0	0	0	0	0	0
Ba	0.001	0.001	0	0	0	0.001	0.0005	0	0	0
Xab	0.07	0.05	0.071	0.056	0.057	0.061	0.083	0.059	0.055	0.059
Xan	0	0	0	0.003	0	0	0	0	0	0
Xor	0.93	0.95	0.929	0.941	0.943	0.939	0.917	0.941	0.945	0.94
Sample anal. # descript	PK-36 2 mcr r	PK-36 3 perth c sg 2	PK-36 6 mcr repl myr 4,5	PK-36 10 perth	PK-36 11 ksp	PK-36 4 ksp sg 11	PK-36 2 ksp incl	PK-36 3 incl cass	PK-36 1 cont cass myr	PK-36 2 cont myr
Fe2O3	0	0	0.02	0	0.01	0.03	0	0.01	0	0
Na2O	0.5	0.56	0.55	1.24	1.02	1.13	0.43	0.53	0.61	0.68
SiO2	64.7	64.98	65.16	64.75	64.48	65.23	64.96	64.92	64.18	64.86
Al2O3	18.18	18.16	18.18	18.08	18.07	18.36	18.47	18.05	18.3	18.51
MgO	0	0.01	0	0.02	0	0	0.02	0.02	0.02	0
K2O	16.73	16.65	16.61	15.5	16.05	16.08	16.75	16.74	16.46	16.44
CaO	0	0	0	0	0	0	0	0	0.02	0
BaO	0.04	0.03	0.02	0.1	0.01	0	0.15	0	0.12	0
Tot.	100.16	100.4	100.55	99.69	99.64	100.82	100.79	100.26	99.71	100.49
Fe	0	0	0.001	0	0	0.001	0	0	0	0
Na	0.045	0.05	0.049	0.111	0.092	0.101	0.039	0.047	0.056	0.061
Si	2.996	3	3.002	3.0205	2.996	2.994	2.9905	3.002	2.987	2.989
Al	0.992	0.988	0.987	0.988	0.989	0.993	1.002	0.984	1.004	1.006
Mg	0	0.001	0	0.001	0	0	0.001	0.001	0.001	0
K	0.989	0.981	0.9765	0.917	0.952	0.941	0.983	0.987	0.977	0.966
Ca	0	0	0	0	0	0	0	0	0.001	0
Ba	0.001	0.001	0	0.002	0	0	0.003	0	0.002	0
Xab	0.044	0.049	0.048	0.108	0.088	0.097	0.038	0.046	0.054	0.059
Xan	0	0	0	0	0	0	0	0	0.001	0
Xor	0.956	0.951	0.952	0.892	0.912	0.903	0.962	0.954	0.945	0.941

Sample anal # descript	PK-36 3 ksp cont myr	PK-38a 5 ksp	PK-38a 8 ksp	PK-43a 1 ksp cont	PK-43a 6 ksp tourcont	PK-43a 7 ksp 2-5 cont	PK-43a 2 ksp 2-5 cont	PK-57 1 ksp myr	PK-57 2 ksp mcr	PK-57 5 ksp sg 1 perth
Fe2O3	0	0.02	0	0.05	0.05	0	0.02	0.03	0	0.01
Na2O	0.69	1.08	1.17	0.69	1.12	1.08	0.65	0.85	0.71	0.87
SiO2	64.9	65.36	65.1	65.21	64.81	65.52	64.81	64.68	64.48	64.86
Al2O3	18.35	18.09	18.3	18.21	17.63	17.73	18.06	18.45	18.24	17.88
MgO	0	0	0	0.01	0	0	0.01	0	0.01	0.01
K2O	16.59	16.04	15.94	16.49	15.74	16	16.59	16.28	16.3	16.43
CaO	0	0	0	0.01	0	0.03	0.02	0.01	0.03	0
BaO	0.045	0	0.02	0.09	0	0.01	0.02	0.03	0	0
Tot.	100.58	100.6	100.52	100.77	99.35	100.36	100.18	100.32	99.77	100.06
Fe	0	0.001	0	0.002	0.002	0	0.0005	0.001	0	0
Na	0.062	0.096	0.104	0.062	0.101	0.096	0.0585	0.076	0.064	0.078
Si	2.992	3.005	2.995	2.999	3.015	3.018	3	2.987	2.993	3.005
Al	0.997	0.98	0.992	0.987	0.967	0.962	0.985	1.004	0.998	0.976
Mg	0	0	0	0.0005	0	0	0.001	0	0.0005	0.001
K	0.976	0.941	0.9355	0.968	0.934	0.94	0.98	0.959	0.965	0.971
Ca	0	0	0	0.0005	0	0.001	0.01	0	0.002	0
Ba	0.001	0	0	0.002	0	0	0	0.0005	0	0
Xab	0.0595	0.093	0.1	0.06	0.097	0.093	0.056	0.073	0.062	0.075
Xan	0	0	0	0.001	0	0.001	0.001	0	0.002	0
Xor	0.9405	0.907	0.9	0.939	0.903	0.906	0.943	0.926	0.936	0.925

Sample anal # descript	PK-57 9 ksp sg 5	PK-57 13 ksp untw	PK-57 14 mcr	PK-60b 2 untw ksp
Fe2O3	0	0.01	0.03	0.02
Na2O	1.25	1.06	0.56	0.7
SiO2	65.02	65.14	65.26	64.17
Al2O3	18.06	18.05	17.9	17.69
MgO	0	0	0	0
K2O	15.66	16	16.71	16
CaO	0	0.02	0.01	0
BaO	0.03	0	0.04	0
Tot.	100.01	100.28	100.51	98.59
Fe	0	0	0.001	0.001
Na	0.112	0.095	0.05	0.064
Si	3.003	3.004	3.01	3.01
Al	0.983	0.981	0.973	0.978
Mg	0	0	0	0
K	0.9225	0.9415	0.983	0.958
Ca	0	0.001	0	0
Ba	0.005	0	0.001	0
Xab	0.108	0.091	0.049	0.063
Xan	0	0.001	0	0
Xor	0.892	0.908	0.951	0.937

VEIN K-FELDSPAR

Sample anal. # descript plag	PK-12a 6 perth?	PK-14b 1 perth	PK-14b 4 ksp	PK-14b 6 untw ksp	PK-14b 7 untw ksp	PK-14b 8 untw ksp	PK-14b 9 untw ksp	PK-14b 10 untw ksp	PK-14b 11 untw ksp	PK-50b 2 ksp cont
Fe2O3	0	0	0	0	r 0	c sg 7 0	c 0.02	r sg 9 0	c sg 9 0	0.01
Na2O	0.84	0.7	0.73	0.71	0.65	0.74	1.25	0.83	0.68	0.52
SiO2	63.3	64.87	65.05	65.04	65.06	65.24	64.8	65.01	64.99	65.15
Al2O3	18.23	18.45	18.29	18.34	18.17	18.21	18.33	18.24	18.32	18.16
MgO	0	0	0	0	0	0.01	0	0	0.01	0
K2O	16.35	16.61	16.53	16.68	16.7	16.6	15.94	16.44	16.64	16.75
CaO	0	0	0	0	0	0	0.02	0	0	0.01
BaO	0	0	0.06	0	0.03	0	0.03	0	0.04	0.13
Tot.	98.96	100.64	100.66	100.77	100.61	100.79	100.38	100.53	100.69	100.73
Fe	0	0	0	0	0	0	0.001	0	0	0
Na	0.0895	0.0625	0.065	0.063	0.058	0.066	0.111	0.074	0.061	0.047
Si	2.977	2.9885	2.996	2.993	2.999	3	2.989	2.996	2.993	3.001
Al	1.011	1.002	0.993	0.995	0.987	0.987	0.996	0.991	0.995	0.986
Mg	0	0	0	0	0	0.0005	0	0	0.001	0
K	0.981	0.976	0.971	0.979	0.982	0.974	0.938	0.966	0.978	0.984
Ca	0	0	0	0	0	0	0.001	0	0	0
Ba	0	0	0.001	0	0.0005	0	0.001	0	0.001	0
Xab	0.072	0.06	0.063	0.061	0.056	0.063	0.106	0.071	0.059	0.0455
Xan	0	0	0	0	0	0	0.001	0	0	0
Xor	0.928	0.94	0.938	0.939	0.944	0.937	0.893	0.929	0.941	0.945

Sample anal. #	PK-50b 4	PK-50b 1	PK-50b 3	PK-54b 1	PK-54b 3	PK-54b 4	PK-54b 5	PK-60b 1	PK-60b 3	PK-60b 6
descript	cont	plagksp	cont	plaguntw	ksp	ksp repl	ksp repl mcr	ksp	incl cass	sg 1
	4		2		wolf	wolf				inclcass
Fe2O3	0	0	0.01	0	0	0	0	0.02	0.02	0
Na2O	0.75	0.42	0.52	0.68	0.69	0.95	0.66	0.72	0.008	0.59
SiO2	65.27	65.52	64.85	64.68	64.84	64.99	64.87	64.48	64.42	64.62
Al2O3	18.31	18.12	17.96	18.14	18.23	18.25	18.39	18.08	18.13	17.94
MgO	0	0	0.01	0	0	0	0	0	0	0.01
K2O	16.42	16.91	16.86	16.34	16.64	15.99	16.47	16.34	16.24	16.72
CaO	0	0	0.02	0	0	0	0	0.01	0	0
BaO	0	0.02	0.04	0	0	0	0.01	0	0.12	0.11
Tot.	100.75	100.99	100.27	99.85	100.4	100.19	100.41	99.65	99.77	100
Fe	0	0	0	0	0	0	0	0.008	0.001	0
Na	0.066	0.037	0.047	0.061	0.062	0.085	0.059	0.065	0.076	0.053
Si	3	3.0075	3.003	2.999	2.995	2.999	2.992	2.9975	2.994	3.001
Al	0.991	0.98	0.98	0.992	0.993	0.993	1	0.991	0.993	0.982
Mg	0	0	0.001	0	0	0	0	0	0	0.001
K	0.962	0.99	0.996	0.966	0.98	0.941	0.969	0.969	0.963	0.99
Ca	0	0	0.001	0	0	0	0	0	0	0
Ba	0	0	0.001	0	0	0	0	0	0.002	0.002
Xab	0.065	0.036	0.045	0.06	0.059	0.083	0.057	0.063	0.073	0.051
Xan	0	0	0.01	0	0	0	0	0	0	0
Xor	0.935	0.964	0.954	0.94	0.941	0.917	0.943	0.937	0.927	0.949
Sample anal. #	PK-60b 7									
descript	ksp									
	incl cass									
Fe2O3	0									
Na2O	0.74									
SiO2	64.46									
Al2O3	18.16									
MgO	0									
K2O	16.53									
CaO	0									
BaO	0.11									
Tot.	100									
Fe	0									
Na	0.067									
Si	2.992									
Al	0.9935									
Mg	0									
K	0.979									
Ca	0									
Ba	0.002									
Xab	0.064									
Xan	0									
Xor	0.936									

APPENDIX IIb
ELECTRON MICROPROBE ANALYSES OF TOURMALINE FROM
NONG SUA

The abbreviations used are: c core, r rim, i intermediate to core-rim, sg same grain as, band occurs as a growth band, lt light colour, gr green, incl cass as an inclusion in cassiterite, blue tourmaline replacement of repl replacement of earlier tourmaline

Aplite

Sample #	PK-31c	PK-31c	PK-31c	PK-31c	PK-31c	PK-31c	PK-38c	PK-38c	PK-38c	PK-38c
Anal. #	1	2	3	4	5	6	1	2	3	6
Desc.	c	r sg 1	c	r sg 3	c	r sg 5	c	r sg 1	c	r sg 3
Colour	brown	brown	lt brown	brown	lt brown	brown	olive	brown	olive	brown
FeO	12.55	12.82	13.51	12.21	13.18	12.99	12.82	12.28	12.78	12.57
MnO	0.52	0.21	0.68	0.20	0.69	0.21	0.20	0.15	0.23	0.18
TiO2	0.52	0.76	0.56	0.93	0.70	0.74	0.43	0.84	0.41	0.95
Na2O	1.82	1.89	1.98	1.89	1.89	1.97	1.79	1.93	1.81	1.91
Al2O3	34.73	33.32	34.01	34.13	34.09	33.60	33.84	34.26	34.30	34.16
SiO2	36.34	36.04	35.54	35.40	35.62	35.14	35.95	35.65	36.03	35.80
F	0.57	0.52	0.49	0.52	0.66	0.50	0.55	0.48	0.45	0.47
MgO	1.09	2.11	0.63	1.83	0.64	1.58	1.93	2.26	1.63	2.24
K2O	0.05	0.06	0.07	0.07	0.04	0.05	0.04	0.07	0.05	0.06
CaO	0.13	0.37	0.28	0.33	0.16	0.33	0.23	0.30	0.22	0.35
SnO2	na	na	na	na	na	na	na	na	na	na
-O=F	0.24	0.22	0.21	0.22	0.28	0.21	0.23	0.20	0.19	0.18
Tot.	88.08	87.88	87.53	87.27	87.38	86.89	87.55	88.01	87.70	88.46
Na	0.582	0.608	0.643	0.611	0.614	0.642	0.577	0.618	0.581	0.609
Ca	0.023	0.066	0.050	0.059	0.029	0.059	0.041	0.053	0.039	0.062
K	0.011	0.013	0.015	0.015	0.009	0.011	0.008	0.015	0.011	0.013
Si	5.997	5.982	5.950	5.898	5.973	5.909	5.980	5.887	5.967	5.886
Al Z	6.003	6.018	6.050	6.102	6.027	6.091	6.020	6.113	6.033	6.114
Al Y	0.752	0.499	0.660	0.600	0.710	0.568	0.615	0.554	0.662	0.505
Fe	1.732	1.779	1.891	1.701	1.848	1.827	1.784	1.696	1.770	1.728
Mg	0.268	0.522	0.157	0.454	0.160	0.396	0.479	0.556	0.402	0.549
Mn	0.073	0.030	0.096	0.028	0.098	0.030	0.028	0.021	0.032	0.025
Ti	0.065	0.095	0.070	0.117	0.088	0.094	0.054	0.104	0.051	0.117
F	0.298	0.273	0.259	0.274	0.350	0.266	0.289	0.251	0.236	0.218
Sample #	PK-38c	PK-38c	PK-38c	PK-43b	PK-43b	PK-43b	PK-43b	PK-58	PK-58	PK-58
Anal. #	7	8	9	1	2	3	4	1	2	3
Desc.	c	c sg 7	r sg 7	c	r sg 1	c	r sg 3	c	r sg 1	c
Colour	olive	brown	olive	blue	brown	blue	brown	brown	brown	brown
FeO	13.03	12.65	12.85	12.17	12.61	11.76	12.80	11.81	12.65	12.77
MnO	0.23	0.19	0.21	0.35	0.24	0.32	0.21	0.13	0.18	0.25
TiO2	0.45	0.80	0.38	0.23	0.60	0.18	0.70	0.52	0.70	0.51
Na2O	1.86	1.92	1.77	1.77	1.98	1.80	1.89	2.00	1.95	2.05
Al2O3	34.59	33.62	35.01	34.34	33.62	34.52	33.23	33.92	34.35	32.90
SiO2	35.77	35.37	36.06	36.59	36.06	36.27	36.08	36.01	35.70	36.64
F	0.55	0.62	0.54	0.38	0.88	0.55	0.60	0.59	0.71	0.45
MgO	1.64	2.49	1.46	2.06	2.20	2.02	1.99	2.53	1.74	2.30
K2O	0.05	0.06	0.04	0.07	0.06	0.05	0.05	0.04	0.05	0.08
CaO	0.21	0.47	0.19	0.30	0.36	0.33	0.34	0.35	0.30	0.29
SnO2	na	na	na	0.00	0.03	0.02	0.01	na	na	na
-O=F	0.23	0.26	0.23	0.16	0.37	0.23	0.25	0.25	0.30	0.19
Tot.	88.17	87.94	88.28	88.10	88.27	87.59	87.64	87.65	88.33	88.06
Na	0.597	0.619	0.565	0.563	0.638	0.577	0.610	0.642	0.627	0.657
Ca	0.037	0.084	0.034	0.053	0.064	0.058	0.061	0.062	0.053	0.051
K	0.011	0.013	0.008	0.015	0.013	0.011	0.011	0.008	0.011	0.017
Si	5.918	5.885	5.938	6.006	5.992	5.995	6.010	5.964	5.925	6.059
Al Z	6.082	6.115	6.062	5.994	6.008	6.005	5.990	6.036	6.075	5.941
Al Y	0.662	0.477	0.733	0.649	0.576	0.721	0.533	0.585	0.644	0.470
Fe	1.803	1.760	1.770	1.671	1.752	1.626	1.783	1.636	1.756	1.766
Mg	0.404	0.618	0.358	0.504	0.545	0.498	0.494	0.625	0.430	0.567
Mn	0.032	0.027	0.029	0.049	0.034	0.045	0.030	0.018	0.025	0.035
Ti	0.056	0.100	0.047	0.028	0.075	0.022	0.088	0.065	0.087	0.063
F	0.288	0.326	0.281	0.197	0.462	0.288	0.316	0.309	0.373	0.235
Sample #	PK-58									
Anal. #	4									
Desc.	r sg 3									
Colour	brown									
FeO	12.00									
MnO	0.16									
TiO2	0.66									
Na2O	1.96									
Al2O3	34.47									
SiO2	35.57									
F	0.57									
MgO	2.41									
K2O	0.06									
CaO	0.43									
SnO2	na									
-O=F	0.24									
Tot.	88.05									

Na	0.628
Ca	0.076
K	0.013
Si	5.875
Al Z	6.125
Al Y	0.585
Fe	1.658
Mg	0.593
Mn	0.022
Ti	0.082
F	0.298

Pegmatite

Sample #	PK-17c	PK-17c	PK-17c	PK-17c	PK-17c	PK-17c	PK-17c	PK-17e	PK-17e
Anal #	1	2	3	5	6	7	8	1	2
Desc.	c	r sg 1	repl	r sg 4	repl	r	c sg 7	c	r sg 1
Colour	olive	brown	blue	brown	blue	brown	lt brown	brown	brown
FeO	12.62	12.46	11.59	12.59	9.88	12.75	12.40	12.82	12.70
MnO	0.18	0.21	0.08	0.21	0.07	0.23	0.19	0.17	0.81
TiO2	0.29	0.76	0.04	0.94	0.05	1.07	0.30	0.66	0.85
Na2O	1.42	1.83	1.85	1.81	1.85	1.86	1.38	1.69	1.94
Al2O3	35.02	34.18	32.74	34.21	32.80	34.39	35.33	34.86	34.96
SiO2	36.97	35.98	37.51	35.69	37.36	35.78	36.95	35.66	35.57
F	0.22	0.59	0.22	0.44	0.32	0.65	0.27	0.47	0.75
MgO	1.11	1.53	3.71	1.65	4.51	1.59	1.22	1.27	0.74
K2O	0.04	0.06	0.02	0.04	0.01	0.05	0.05	0.03	0.04
CaO	0.10	0.38	0.12	0.32	0.19	0.30	0.09	0.19	0.26
SnO2	na	na	na	na	na	na	na	na	na
-O=F	0.09	0.25	0.09	0.19	0.13	0.27	0.11	0.20	0.32
Tot.	87.89	87.72	87.77	87.71	86.90	88.39	88.06	87.61	88.30

Na	0.450	0.589	0.586	0.581	0.589	0.596	0.436	0.543	0.624
Ca	0.018	0.068	0.021	0.057	0.033	0.053	0.016	0.034	0.046
K	0.008	0.013	0.004	0.008	0.002	0.011	0.010	0.006	0.008
Si	6.046	5.968	6.133	5.913	6.130	5.910	6.027	5.910	5.903
Al Z	5.954	6.032	5.867	6.087	5.870	6.090	5.973	6.090	6.097
Al Y	0.796	0.651	0.443	0.594	0.472	0.605	0.819	0.720	0.711
Fe	1.726	1.729	1.585	1.745	1.356	1.761	1.692	1.777	1.763
Mg	0.271	0.378	0.904	0.407	1.103	0.391	0.297	0.314	0.183
Mn	0.025	0.030	0.011	0.029	0.010	0.032	0.026	0.024	0.114
Ti	0.036	0.095	0.005	0.117	0.006	0.133	0.037	0.082	0.106
F	0.114	0.310	0.114	0.231	0.166	0.340	0.139	0.246	0.394

Sample #	PK-17e	PK-33b	PK-33b	PK-33b	PK-33b	PK-33b	PK-33b	PK-43a	PK-43a	PK-43a
Anal #	3	1	2	3	4	5	6	1	2	3
Desc.	band sg 1c	c	c	c	r	r	r	c	c	c
Colour	blue/gr	olive	olive	olive	olive	brown	brown	olive	olive	olive
FeO	13.18	13.31	13.24	13.05	13.48	13.36	13.44	13.16	13.27	12.92
MnO	0.18	0.72	0.16	0.26	0.23	0.23	0.23	0.17	0.19	0.18
TiO2	0.71	0.57	0.36	0.33	0.89	0.92	0.94	0.37	0.38	0.36
Na2O	1.76	1.69	1.57	1.49	1.78	1.87	1.82	1.53	1.61	1.59
Al2O3	34.43	34.67	35.24	35.22	33.93	34.27	34.09	35.42	35.15	35.56
SiO2	35.46	35.98	36.18	36.23	35.49	35.74	35.65	36.30	36.47	36.58
F	0.53	0.32	0.36	0.38	0.40	0.37	0.35	0.32	0.24	0.28
MgO	1.34	1.17	1.09	1.03	1.17	1.61	1.28	1.04	1.04	1.02
K2O	0.05	0.05	0.03	0.03	0.05	0.04	0.03	0.05	0.04	0.03
CaO	0.28	0.17	0.14	0.11	0.25	0.35	0.26	0.13	0.12	0.11
SnO2	na	na	na	na	na	na	na	na	na	na
-O=F	0.22	0.13	0.15	0.16	0.17	0.16	0.15	0.13	0.10	0.12
Tot.	87.69	88.01	88.22	87.97	87.49	88.59	87.93	88.34	88.41	88.51

Na	0.568	0.541	0.500	0.475	0.576	0.597	0.585	0.486	0.511	0.503
Ca	0.050	0.030	0.025	0.019	0.045	0.062	0.046	0.023	0.021	0.019
K	0.011	0.011	0.006	0.006	0.011	0.008	0.006	0.010	0.008	0.006
Si	5.901	5.935	5.943	5.962	5.919	5.883	5.908	5.944	5.965	5.964
Al Z	6.099	6.065	6.057	6.038	6.081	6.117	6.092	6.056	6.035	6.036
Al Y	0.653	0.676	0.765	0.794	0.588	0.531	0.567	0.780	0.740	0.796
Fe	1.834	1.836	1.819	1.796	1.880	1.839	1.863	1.802	1.815	1.762
Mg	0.332	0.288	0.267	0.253	0.291	0.395	0.316	0.254	0.254	0.248
Mn	0.025	0.031	0.022	0.036	0.032	0.032	0.032	0.024	0.026	0.025
Ti	0.089	0.071	0.044	0.041	0.112	0.114	0.117	0.046	0.047	0.044
F	0.279	0.167	0.187	0.198	0.211	0.193	0.183	0.166	0.124	0.144

Sample #	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43b	PK-43b	PK-43b	PK-43b
Anal. #	4	5	6	7	8	5	6	7	8
Desc.	c	c	r	r	r	c	r sg 5	c	r sg 7
Colour	olive	olive	brown	brown	brown	blue	brown	brown	brown
FeO	13.15	12.97	13.22	13.41	12.88	12.26	12.69	12.52	12.86
MnO	0.21	0.21	0.21	0.27	0.17	0.35	0.26	0.21	0.21
TiO2	0.35	0.35	0.99	0.99	0.92	0.39	0.71	0.57	0.94
Na2O	1.48	1.47	1.82	1.79	1.88	1.90	1.83	1.77	1.85
Al2O3	35.67	35.50	34.85	34.78	34.51	32.87	33.72	33.21	33.72
SiO2	36.58	36.16	35.53	35.57	35.38	36.30	35.50	35.76	35.34
F	0.26	0.23	0.31	0.38	0.48	0.65	0.54	0.59	0.61
MgO	1.03	1.07	1.35	1.17	1.77	2.38	1.78	2.21	1.57
K2O	0.04	0.04	0.05	0.05	0.06	0.03	0.05	0.05	0.08
CaO	0.11	0.12	0.34	0.33	0.37	0.35	0.29	0.35	0.30
SnO2	na	na	na	na	na	0.02	0.02	0.00	0.01
-O=F	0.11	0.10	0.13	0.16	0.20	0.27	0.23	0.25	0.26
Tot.	88.77	88.02	88.55	88.58	88.20	87.24	87.15	86.98	87.23
Na	0.467	0.467	0.580	0.571	0.603	0.616	0.594	0.575	0.601
Ca	0.019	0.021	0.060	0.058	0.066	0.063	0.052	0.063	0.054
K	0.008	0.008	0.010	0.011	0.013	0.006	0.011	0.011	0.017
Si	5.950	5.928	5.839	5.856	5.849	6.066	5.940	5.992	5.923
Al Z	6.050	6.072	6.161	6.144	6.151	5.934	6.060	6.008	6.077
Al Y	0.788	0.788	0.590	0.604	0.572	0.541	0.589	0.550	0.584
Fe	1.789	1.778	1.817	1.846	1.781	1.713	1.776	1.754	1.803
Mg	0.250	0.261	0.331	0.287	0.436	0.593	0.444	0.552	0.392
Mn	0.029	0.029	0.029	0.038	0.024	0.050	0.037	0.030	0.030
Ti	0.043	0.043	0.122	0.123	0.114	0.049	0.089	0.072	0.118
F	0.134	0.119	0.161	0.198	0.251	0.344	0.286	0.313	0.323

Mineralized Pegmatite

Sample #	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36b
Anal. #	1	2	3	4	5	6	7	8	9	1
Type	c	r	i	i	c	c	r	r	c	r
Colour	brown	brown	red/brown	red/brown	brown	brown	brown	brown	brown	brown
FeO	9.72	12.62	13.99	9.63	9.28	13.70	13.84	14.22	8.96	13.58
MnO	0.42	0.18	0.28	0.35	0.39	0.12	0.16	0.17	0.31	0.28
TiO2	1.01	0.86	0.53	1.45	0.95	0.27	0.83	0.70	0.68	0.89
Na2O	2.17	1.95	1.96	2.15	2.23	1.77	2.16	1.93	2.21	2.08
Al2O3	31.57	33.10	33.84	31.09	31.74	34.12	33.79	32.08	32.32	32.72
SiO2	36.62	35.56	35.77	36.53	36.77	36.06	35.96	34.86	36.77	35.60
F	0.96	0.88	0.83	0.98	0.99	0.58	0.84	0.66	0.86	0.77
MgO	5.42	2.36	1.43	5.27	5.56	1.10	1.12	1.20	5.53	2.03
K2O	0.08	0.06	0.06	0.07	0.06	0.04	0.07	0.07	0.05	0.07
CaO	0.92	0.43	0.17	0.85	0.95	0.08	0.14	0.16	0.79	0.55
SnO2	0.01	0.04	0.01	0.01	0.01	0.01	0.04	0.04	0.01	0.04
-O=F	0.40	0.37	0.35	0.41	0.42	0.24	0.35	0.28	0.36	0.32
Tot.	88.47	87.67	88.51	87.97	88.52	87.61	88.60	85.80	88.12	88.30
Na	0.693	0.634	0.633	0.691	0.710	0.572	0.697	0.644	0.703	0.675
Ca	0.162	0.077	0.030	0.151	0.167	0.014	0.025	0.030	0.139	0.099
K	0.017	0.013	0.013	0.015	0.013	0.009	0.015	0.015	0.010	0.015
Si	6.031	5.961	5.960	6.051	6.042	6.017	5.983	6.000	6.032	5.956
Al Z	5.969	6.039	6.040	5.949	5.958	5.988	6.017	6.000	5.968	6.044
Al Y	0.159	0.501	0.606	0.121	0.188	0.716	0.610	0.507	0.280	0.408
Fe	1.339	1.769	1.949	1.334	1.275	1.910	1.926	2.047	1.229	1.900
Mg	1.331	0.590	0.355	1.301	1.362	0.273	0.278	0.308	1.352	0.506
Mn	0.059	0.026	0.040	0.049	0.054	0.017	0.023	0.025	0.043	0.040
Ti	0.125	0.108	0.066	0.181	0.117	0.034	0.104	0.091	0.084	0.112
F	0.500	0.467	0.437	0.513	0.514	0.306	0.442	0.359	0.446	0.407
Sample #	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b	PK-36b
Anal. #	2	3	4	5	6	7	8	9	10	11
Type	i sg 1	i sg 1	i sg 1	i sg 1	i sg 1	c sg 1	c sg 1	c sg 1	c sg 1	c sg
Colour	brown	brown	brown	lt brown	lt brown	lt brown	lt brown	lt brown	blue	olive
FeO	14.10	14.34	9.74	9.43	9.86	13.58	13.92	10.18	13.22	13.32
MnO	0.22	0.31	0.32	0.35	0.28	0.27	0.29	0.34	0.30	0.32
TiO2	0.92	0.88	0.72	0.86	1.06	0.50	0.42	0.80	0.29	0.49
Na2O	2.25	1.97	2.05	2.12	2.03	1.98	2.03	2.10	2.03	2.19
Al2O3	33.06	33.38	32.14	31.88	31.95	33.64	32.61	32.34	33.96	32.46
SiO2	35.52	35.97	36.56	36.67	36.57	36.14	36.25	36.69	36.03	36.39
F	0.86	0.73	1.02	1.15	0.94	0.75	0.88	0.88	0.86	0.82
MgO	1.56	1.19	5.22	5.50	5.15	1.47	1.77	4.00	1.58	1.87
K2O	0.12	0.07	0.06	0.05	0.07	0.08	0.06	0.05	0.06	0.08
CaO	0.23	0.19	0.78	0.91	0.98	0.17	0.26	0.47	0.25	0.26
SnO2	0.00	0.01	0.03	0.02	0.02	0.00	0.01	0.02	0.02	0.00
-O=F	0.36	0.31	0.43	0.48	0.40	0.32	0.37	0.37	0.36	0.35
Tot.	88.47	88.70	88.21	88.46	88.50	88.26	88.12	87.49	88.23	87.87

Na	0.730	0.635	0.656	0.678	0.647	0.639	0.660	0.675	0.655	0.711
Ca	0.041	0.034	0.138	0.161	0.173	0.030	0.047	0.084	0.045	0.047
K	0.026	0.015	0.013	0.011	0.015	0.017	0.013	0.011	0.013	0.017
Si	5.948	5.982	6.033	6.045	6.015	6.013	6.075	6.086	6.000	6.096
Al Z	6.052	6.018	5.967	5.955	5.985	5.987	5.925	5.914	6.000	5.904
Al Y	0.472	0.526	0.284	0.239	0.209	0.610	0.515	0.409	0.666	0.504
Fe	1.974	1.995	1.344	1.300	1.356	1.890	1.951	1.412	1.841	1.866
Mg	0.389	0.295	1.284	1.351	1.263	0.365	0.442	0.989	0.392	0.467
Mn	0.031	0.044	0.045	0.049	0.039	0.038	0.041	0.048	0.042	0.045
Ti	0.116	0.110	0.089	0.107	0.131	0.063	0.053	0.100	0.036	0.062
F	0.455	0.384	0.532	0.600	0.489	0.395	0.466	0.462	0.453	0.434

Sample #	PK-36b	PK-36b	PK-36b	PK-59	PK-59	PK-59	PK-59	PK-59	PK-59	PK-59
Anal. #	12	13	14	1	2	3	4	5	6	7
Type	c sg 1	i sg 1	r sg 1	incl	cassincl	cassincl	cassc	c sg 4	i sg 4	i sg
4										
Colour	olive	brown	brown	brown	brown	brown	blue	brown	lt	brown dk brown
FeO	13.19	9.28	10.14	13.83	14.34	13.94	14.80	14.28	13.87	14.34
MnO	0.32	0.31	0.37	0.33	0.13	0.22	0.19	0.12	0.21	0.17
TiO2	0.41	1.06	0.79	0.47	0.72	0.85	0.11	0.72	0.54	0.90
Na2O	2.09	2.07	2.14	2.03	2.26	1.94	2.13	1.95	1.93	2.20
Al2O3	33.63	31.57	32.23	34.10	33.52	32.57	33.75	33.63	33.88	33.18
SiO2	36.19	36.66	36.65	35.94	36.17	36.49	35.74	36.16	35.75	35.98
F	0.57	1.07	0.89	0.55	0.55	0.63	0.68	0.50	0.85	0.48
MgO	1.70	5.24	4.58	1.19	1.11	1.59	0.88	1.03	1.34	1.36
K2O	0.06	0.05	0.07	0.03	0.06	0.07	0.04	0.04	0.05	0.06
CaO	0.24	1.04	0.65	0.16	0.16	0.20	0.12	0.16	0.19	0.25
SnO2	0.02	0.02	0.02	na	na	na	na	na		
-O=F	0.24	0.45	0.37	0.23	0.23	0.27	0.29	0.21	0.36	0.20
Tot.	88.19	87.93	88.15	88.39	88.78	88.23	88.14	88.38	88.25	88.71

Na	0.672	0.664	0.685	0.653	0.726	0.626	0.692	0.627	0.625	0.707
Ca	0.043	0.184	0.115	0.028	0.028	0.036	0.022	0.028	0.034	0.044
K	0.013	0.011	0.015	0.006	0.013	0.015	0.009	0.008	0.011	0.013
Si	6.003	6.068	6.047	5.958	5.989	6.074	5.985	5.997	5.969	5.964
Al Z	5.997	5.932	5.953	6.042	6.011	5.926	6.015	6.003	6.031	6.036
Al Y	0.577	0.226	0.314	0.621	0.530	0.463	0.646	0.570	0.635	0.445
Fe	1.830	1.285	1.399	1.918	1.986	1.940	2.073	1.981	1.937	1.988
Mg	0.420	1.293	1.126	0.294	0.274	0.394	0.220	0.255	0.333	0.336
Mn	0.045	0.043	0.052	0.046	0.018	0.031	0.027	0.017	0.030	0.024
Ti	0.051	0.132	0.098	0.059	0.090	0.106	0.014	0.090	0.068	0.112
F	0.299	0.560	0.464	0.288	0.288	0.332	0.360	0.262	0.449	0.252

Sample #	PK-59	PK-59	PK-59	PK-59	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63
Anal. #	8	9	10	11	1	2	3	4	5	6
Type	i sg 4	r sg 4	c	r sg 10	r	r sg 1	i sg 1	i sg 1	repl	i sg
1										
Colour	blue	brown	blue	brown	brown	brown	blue	brown	blue	brown
FeO	14.60	12.82	15.29	12.86	12.22	13.44	14.62	12.55	9.86	11.54
MnO	0.20	0.53	0.44	0.22	0.19	0.21	0.41	0.33	0.07	0.43
TiO2	0.14	0.71	0.11	0.84	0.88	0.89	0.18	0.40	0.03	0.36
Na2O	1.91	2.05	1.80	1.95	1.97	1.97	1.78	1.89	1.90	1.67
Al2O3	33.68	33.35	33.75	33.51	33.52	32.97	34.48	33.72	33.06	34.52
SiO2	36.37	35.97	36.00	36.16	36.03	36.60	36.08	36.82	37.80	37.11
F	0.92	0.77	0.64	0.65	0.79	0.59	0.67	0.67	0.26	0.45
MgO	1.21	1.92	0.24	2.05	2.64	1.67	0.29	2.08	5.03	2.31
K2O	0.06	0.07	0.06	0.06	0.06	0.06	0.05	0.05	0.01	0.03
CaO	0.15	0.48	0.28	0.37	0.38	0.21	0.20	0.24	0.20	0.19
SnO2	na	na	na	na	0.04	0.01	0.02	0.00	0.00	0.00
-O=F	0.39	0.32	0.27	0.27	0.33	0.25	0.28	0.28	0.11	0.19
Tot.	88.85	88.36	88.34	88.38	88.39	88.36	88.51	88.46	88.12	88.41

Na	0.616	0.661	0.584	0.625	0.632	0.632	0.574	0.603	0.596	0.528
Ca	0.027	0.086	0.050	0.066	0.067	0.037	0.036	0.042	0.035	0.033
K	0.013	0.015	0.013	0.013	0.013	0.013	0.011	0.011	0.002	0.006
Si	6.050	5.982	6.024	5.981	5.962	6.060	6.001	6.063	6.111	6.046
Al Z	5.950	6.018	5.976	6.019	6.038	5.940	5.999	5.937	5.889	5.954
Al Y	0.654	0.518	0.680	0.514	0.499	0.494	0.760	0.607	0.410	0.675
Fe	2.031	1.783	2.140	1.779	1.691	1.861	2.034	1.728	1.333	1.572
Mg	0.300	0.476	0.060	0.505	0.651	0.412	0.072	0.510	1.212	0.561
Mn	0.028	0.075	0.062	0.031	0.027	0.029	0.058	0.046	0.010	0.059
Ti	0.018	0.089	0.014	0.104	0.110	0.111	0.023	0.050	0.004	0.044
F	0.484	0.405	0.339	0.340	0.413	0.309	0.352	0.349	0.133	0.232

Sample #	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63	PK-63
Anal. #	7	8	9	10	11	12	13	14	15	16
Type	c sg 1	i sg 1	i sg 1	r sg 1	r	i sg 11	c sg 11	i sg 11	r sg 11	repl
Colour	blue	brown	green	brown	brown	blue	brown	blue	brown	blue
FeO	14.82	13.92	14.13	12.24	12.60	14.77	12.49	15.20	12.30	10.63
MnO	0.41	0.17	0.16	0.21	0.25	0.46	0.47	0.44	0.19	0.05
TiO2	0.12	0.71	0.54	0.85	0.79	0.27	1.06	0.16	0.88	0.00
Na2O	1.63	2.09	1.83	1.94	1.94	1.79	1.96	1.94	1.93	1.42
Al2O3	34.76	33.22	33.92	33.66	33.13	34.59	31.74	34.21	33.23	34.34
SiO2	36.31	36.60	36.55	35.97	36.35	34.73	35.89	35.71	36.27	37.80
F	0.51	0.64	0.57	0.58	0.75	0.68	0.71	0.75	0.57	0.01
MgO	0.10	0.98	0.83	2.67	2.43	0.21	2.63	0.13	2.64	3.61
K2O	0.04	0.05	0.05	0.05	0.06	0.07	0.05	0.06	0.05	0.00
CaO	0.11	0.17	0.09	0.39	0.36	0.32	0.53	0.29	0.47	0.07
SnO2	0.02	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.02
-O=F	0.21	0.27	0.24	0.24	0.32	0.29	0.30	0.32	0.24	0.00
Tot.	88.62	88.29	88.42	88.31	88.33	87.59	87.22	88.57	88.28	87.94

Na	0.523	0.673	0.587	0.620	0.623	0.586	0.639	0.629	0.617	0.444
Ca	0.020	0.030	0.016	0.069	0.064	0.058	0.096	0.052	0.083	0.012
K	0.008	0.011	0.011	0.011	0.013	0.015	0.011	0.013	0.011	0.000
Si	6.008	6.081	6.048	5.933	6.020	5.863	6.038	5.974	5.984	6.090
Al Z	5.992	5.919	5.952	6.067	5.980	6.137	5.962	6.026	6.016	5.910
Al Y	0.787	0.585	0.663	0.477	0.486	0.746	0.332	0.718	0.446	0.612
Fe	2.051	1.934	1.955	1.688	1.745	2.085	1.757	2.126	1.697	1.437
Mg	0.025	0.243	0.205	0.656	0.600	0.053	0.660	0.032	0.649	0.867
Mn	0.057	0.024	0.022	0.029	0.035	0.066	0.067	0.062	0.027	0.007
Ti	0.015	0.089	0.067	0.105	0.098	0.034	0.134	0.070	0.109	0.000
F	0.267	0.336	0.298	0.303	0.393	0.363	0.378	0.397	0.297	0.005

Sample #	PK-65	PK-65	PK-65	PK-65
Anal. #	4	5	6	7
Type	c	r sg 4	c	r sg 6
Colour	red	brown	brown	brown
FeO	10.80	11.97	12.39	13.18
MnO	0.36	0.14	0.13	0.26
TiO2	1.24	0.84	0.67	0.73
Na2O	2.07	2.04	2.10	1.88
Al2O3	31.29	33.68	32.91	32.03
SiO2	36.65	35.94	36.20	36.26
F	1.02	0.69	0.67	0.64
MgO	4.86	2.82	2.91	2.35
K2O	0.07	0.06	0.06	0.03
CaO	0.80	0.40	0.47	0.62
SnO2	na	na	na	na
-O=F	0.43	0.29	0.28	0.27
Tot.	88.73	88.29	88.25	87.72

Na	0.663	0.653	0.674	0.610
Ca	0.142	0.071	0.083	0.111
K	0.015	0.013	0.013	0.006
Si	6.052	5.935	5.996	6.065
Al Z	5.948	6.065	6.004	5.935
Al Y	0.142	0.489	0.420	0.379
Fe	1.492	1.653	1.716	1.844
Mg	1.196	0.694	0.718	0.586
Mn	0.050	0.020	0.018	0.037
Ti	0.154	0.104	0.083	0.092
F	0.533	0.360	0.351	0.339

Vein

Sample #	PK-48d	PK-48d	PK-48d	PK-48d	PK-48d	PK-48d	PK-50b	PK-50b	PK-50b	PK-50b
Anal. #	1	2	3	4	5	6	1	2	3	4
Type	c	r sg 1	c	r sg 3	c	r sg 5	r	c	i	r
Colour	olive	brown	olive	brown	olive	brown	brown?	brown?	brown?	brown?
FeO	12.36	12.27	12.66	12.22	9.64	12.31	13.82	11.95	11.32	13.59
MnO	0.19	0.19	0.18	0.14	0.29	0.21	0.45	0.21	0.36	0.36
TiO2	0.85	0.84	0.54	0.94	0.29	0.73	0.54	0.87	1.40	0.55
Na2O	2.02	1.98	2.00	1.96	1.95	1.95	2.02	2.09	2.19	1.90
Al2O3	33.86	34.04	33.83	34.04	34.00	33.96	33.71	32.82	31.11	34.07
SiO2	35.26	35.56	36.06	35.74	36.91	35.66	35.46	36.14	36.46	35.35
F	0.66	0.59	0.70	0.54	0.49	0.66	0.69	0.75	0.54	0.59
MgO	2.59	2.41	2.47	2.54	4.31	2.29	1.02	3.32	4.30	1.22
K2O	0.05	0.06	0.05	0.04	0.03	0.04	0.07	0.05	0.07	0.05
CaO	0.52	0.38	0.28	0.37	0.31	0.31	0.38	0.57	0.78	0.45
SnO2	na	na	na	na	na	na	na	na	na	na
-O=F	0.28	0.25	0.29	0.23	0.21	0.28	0.29	0.32	0.23	0.25
Tot.	88.07	88.08	88.46	88.28	88.02	87.84	87.87	88.44	88.31	87.88

Na	0.650	0.635	0.641	0.626	0.615	0.628	0.657	0.670	0.701	0.615
Ca	0.093	0.067	0.050	0.065	0.054	0.055	0.068	0.101	0.138	0.081
K	0.011	0.013	0.011	0.008	0.006	0.008	0.015	0.011	0.015	0.011
Si	5.856	5.886	5.956	5.890	6.006	5.921	5.946	5.974	6.021	5.905
Al Z	6.144	6.114	6.044	6.110	5.994	6.079	6.054	6.026	5.979	6.095
Al Y	0.484	0.527	0.542	0.502	0.527	0.567	0.608	0.368	0.076	0.612
Fe	1.717	1.699	1.749	1.684	1.312	1.709	1.938	1.652	1.563	1.898
Mg	0.641	0.595	0.608	0.624	1.045	0.567	0.255	0.818	1.058	0.304
Mn	0.027	0.027	0.025	0.020	0.040	0.030	0.064	0.029	0.050	0.051
Ti	0.106	0.105	0.067	0.116	0.035	0.091	0.068	0.108	0.174	0.069
F	0.347	0.309	0.366	0.281	0.252	0.347	0.366	0.392	0.282	0.312

Sample #	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b
Anal. #	1	2	3	4	5	6	7	8	9	1
Type	r	i	i	c	c	r	r	i	i	c
Colour	brown	brown	brown	olive	olive	brown	brown	brown	brown	olive
FeO	12.39	12.25	12.36	12.00	12.22	12.17	12.31	11.99	12.65	12.06
MnO	0.19	0.19	0.17	0.11	0.15	0.11	0.14	0.15	0.19	0.13
TiO2	0.94	0.99	1.09	0.47	0.46	1.05	0.97	0.98	1.02	0.49
Na2O	1.81	1.82	1.86	1.80	1.79	1.78	1.76	1.86	1.82	1.78
Al2O3	33.96	34.08	34.09	34.41	34.49	34.39	34.25	33.47	33.24	34.50
SiO2	35.24	35.26	34.62	35.58	35.71	35.38	35.65	35.84	35.33	36.35
F	0.34	0.25	0.43	0.36	0.40	0.50	0.48	0.40	0.36	0.35
MgO	2.00	2.21	2.08	2.18	2.25	2.20	2.05	2.68	2.08	2.25
K2O	0.06	0.05	0.05	0.04	0.05	0.05	0.05	0.05	0.05	0.06
CaO	0.33	0.35	0.44	0.23	0.22	0.40	0.37	0.44	0.33	0.23
SnO2	na	na	na	na	na	na	na	na	na	na
-O=F	0.14	0.11	0.18	0.15	0.17	0.21	0.20	0.17	0.15	0.15
Tot.	87.12	87.34	87.00	87.01	87.57	87.81	87.82	87.69	86.93	88.06

Na	0.585	0.585	0.604	0.580	0.574	0.571	0.564	0.596	0.591	0.566
Ca	0.059	0.062	0.079	0.041	0.039	0.071	0.066	0.078	0.059	0.040
K	0.013	0.011	0.011	0.008	0.011	0.011	0.011	0.011	0.011	0.013
Si	5.872	5.847	5.795	5.909	5.905	5.854	5.895	5.927	5.913	5.960
Al Z	6.128	6.153	6.205	6.091	6.095	6.146	6.105	6.073	6.087	6.040
Al Y	0.540	0.509	0.521	0.644	0.627	0.559	0.571	0.451	0.470	0.626
Fe	1.726	1.699	1.730	1.667	1.690	1.684	1.702	1.658	1.771	1.654
Mg	0.497	0.546	0.519	0.540	0.555	0.543	0.505	0.661	0.519	0.550
Mn	0.027	0.027	0.024	0.015	0.021	0.015	0.020	0.021	0.027	0.018
Ti	0.118	0.123	0.137	0.059	0.057	0.131	0.121	0.122	0.128	0.060
F	0.179	0.131	0.228	0.189	0.209	0.262	0.251	0.209	0.191	0.181

Sample #	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b	PK-54b
Anal. #	2	3	4	5	6	7	8
Type	c sg 1	r sg 1	r sg 1	c	r sg 5	c	r
Colour	olive	brown	brown	brown	brown	olive	brown
FeO	12.22	12.36	12.87	12.43	12.04	11.97	12.57
MnO	0.20	0.18	0.20	0.17	0.13	0.18	0.21
TiO2	0.46	0.93	0.82	0.95	0.99	0.45	0.91
Na2O	1.84	1.77	1.90	1.92	1.78	1.74	1.75
Al2O3	34.59	33.97	34.41	33.37	34.14	34.63	34.07
SiO2	36.20	35.86	35.45	35.79	35.67	36.02	35.65
F	0.39	0.33	0.59	0.56	0.37	0.47	0.46
MgO	2.21	2.15	1.79	2.46	2.22	2.20	1.99
K2O	0.05	0.06	0.05	0.06	0.05	0.04	0.05
CaO	0.24	0.35	0.32	0.46	0.37	0.23	0.35
SnO2	na	na	na	na	na	na	na
-O=F	0.16	0.14	0.25	0.24	0.16	0.20	0.19
Tot.	88.24	87.81	88.14	87.93	87.58	87.72	87.82

Na	0.585	0.566	0.610	0.617	0.570	0.556	0.562
Ca	0.042	0.062	0.057	0.082	0.066	0.041	0.062
K	0.010	0.013	0.011	0.013	0.011	0.008	0.011
Si	5.937	5.917	5.874	5.936	5.895	5.939	5.904
Al Z	6.063	6.083	6.126	6.064	6.105	6.061	6.096
Al Y	0.623	0.524	0.595	0.459	0.545	0.668	0.554
Fe	1.676	1.706	1.784	1.724	1.664	1.651	1.741
Mg	0.540	0.529	0.442	0.608	0.547	0.541	0.491
Mn	0.028	0.025	0.028	0.024	0.018	0.025	0.029
Ti	0.057	0.115	0.102	0.118	0.123	0.056	0.113
F	0.202	0.172	0.309	0.294	0.193	0.245	0.241

Metasediment at contact

Sample #	PK-65	PK-65	PK-65
Anal. #	1	2	3
Type	metased	metased	metased
Colour	brown	brown	brown
FeO	12.25	10.92	12.23
MnO	0.34	0.25	0.18
TiO2	0.75	0.66	0.91
Na2O	1.95	2.05	1.96
Al2O3	32.95	32.58	32.80
SiO2	36.25	36.65	36.20
F	0.56	0.91	0.62
MgO	2.42	4.05	2.83
K2O	0.06	0.05	0.07
CaO	0.44	0.65	0.47
SnO2	na	na	na
-O=F	0.24	0.38	0.26
Tot.	87.72	88.39	88.01
Na	0.628	0.656	0.630
Ca	0.078	0.115	0.083
K	0.013	0.011	0.015
Si	6.020	6.045	6.000
Al Z	5.980	5.955	6.000
Al Y	0.469	0.378	0.407
Fe	1.701	1.506	1.695
Mg	0.599	0.996	0.699
Mn	0.048	0.035	0.025
Ti	0.094	0.082	0.113
F	0.294	0.475	0.325

APPENDIX IIc
ELECTRON MICROPROBE ANALYSES OF MUSCOVITE FROM
NONG SUA

Aplite

Sample	PK-1a	PK-1a	PK-1a	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c
Anal #	1	2	ave	1	2	3	4	5	6	7
FeO	1.66	1.70	1.68	1.83	1.77	1.84	1.77	2.02	1.57	1.60
MnO	0.06	0.12	0.09	0.06	0.03	0.02	0.05	0.05	0.04	0.09
Na2O	0.54	0.36	0.45	0.55	0.57	0.62	0.41	0.56	0.64	0.56
Al2O3	35.14	34.65	34.90	34.28	34.48	34.30	34.88	33.59	34.67	34.74
SiO2	46.31	46.36	46.34	46.47	46.33	46.22	46.53	46.97	47.14	46.35
MgO	0.42	0.35	0.39	0.43	0.48	0.50	0.44	0.45	0.44	0.51
K2O	11.01	11.52	11.27	11.46	11.34	11.09	11.39	10.99	10.85	11.02
CaO	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.00
TiO2	0.07	0.43	0.25	0.55	0.47	0.10	0.41	0.39	0.39	0.31
F	0.21	0.28	0.25	0.29	0.43	0.45	0.46	0.45	0.36	0.56
-O=F	0.09	0.12	0.10	0.12	0.18	0.19	0.19	0.19	0.15	0.24
Tot.	95.33	95.64	95.49	95.81	95.67	94.89	96.13	95.45	95.96	95.50
K	1.882	1.970	1.926	1.959	1.942	1.915	1.942	1.889	1.842	1.891
Na	0.140	0.094	0.117	0.143	0.148	0.163	0.106	0.146	0.165	0.146
Ca	0.000	0.000	0.000	0.001	0.001	0.000	0.000	0.000	0.001	0.000
Si	6.206	6.216	6.211	6.226	6.221	6.256	6.219	6.328	6.274	6.234
Al IV	1.794	1.784	1.789	1.774	1.779	1.744	1.781	1.672	1.726	1.766
Al VI	3.757	3.691	3.724	3.640	3.678	3.728	3.713	3.661	3.713	3.742
Fe	0.186	0.191	0.188	0.205	0.199	0.208	0.198	0.228	0.175	0.180
Mg	0.084	0.070	0.077	0.086	0.096	0.101	0.088	0.090	0.087	0.102
Mn	0.007	0.014	0.010	0.007	0.003	0.002	0.006	0.006	0.005	0.010
Ti	0.009	0.054	0.032	0.069	0.059	0.013	0.052	0.049	0.049	0.039
F	0.089	0.119	0.104	0.123	0.183	0.193	0.194	0.192	0.152	0.238
Sample	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-25	PK-25	PK-25	PK-25	PK-25
Anal #	8	9	10	11	ave	1	2	3	4	5
FeO	1.67	1.41	1.49	1.34	1.66	1.63	1.84	1.65	1.56	1.70
MnO	0.05	0.05	0.08	0.05	0.05	0.08	0.09	0.05	0.04	0.06
Na2O	0.46	0.62	0.42	0.42	0.53	0.38	0.34	0.33	0.44	0.35
Al2O3	34.60	35.08	35.21	35.33	34.65	35.25	34.67	35.70	35.56	35.06
SiO2	46.87	47.07	46.16	45.88	46.54	46.24	46.91	46.54	46.35	46.59
MgO	0.45	0.44	0.40	0.39	0.45	0.37	0.43	0.34	0.31	0.36
K2O	11.30	10.98	11.44	11.56	11.22	11.47	11.37	11.63	11.36	11.52
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.35	0.20	0.24	0.35	0.34	0.33	0.34	0.40	0.36	0.37
F	0.44	0.46	0.35	0.32	0.42	0.27	0.40	0.32	0.39	0.39
-O=F	0.19	0.19	0.15	0.13	0.17	0.11	0.17	0.13	0.16	0.16
Tot.	96.00	96.11	95.65	95.52	95.70	95.91	96.22	96.32	96.22	96.25
K	1.926	1.864	1.957	1.979	1.919	1.955	1.934	1.977	1.931	1.960
Na	0.119	0.160	0.109	0.109	0.138	0.098	0.088	0.085	0.114	0.091
Ca	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000
Si	6.263	6.265	6.189	6.159	6.240	6.178	6.256	6.197	6.176	6.214
Al IV	1.737	1.735	1.811	1.841	1.760	1.822	1.744	1.803	1.824	1.786
Al VI	3.713	3.768	3.753	3.748	3.714	3.729	3.706	3.721	3.761	3.726
Fe	0.187	0.157	0.167	0.150	0.187	0.182	0.205	0.184	0.174	0.190
Mg	0.090	0.087	0.080	0.078	0.090	0.074	0.085	0.067	0.062	0.072
Mn	0.006	0.006	0.009	0.006	0.006	0.009	0.010	0.006	0.005	0.007
Ti	0.044	0.025	0.030	0.044	0.043	0.041	0.043	0.050	0.045	0.046
F	0.186	0.194	0.148	0.136	0.176	0.114	0.169	0.135	0.164	0.165
Sample	PK-25	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-43b	PK-43b	PK-43b
Anal #	ave	1	2	3	4	5	ave	1	2	3
FeO	1.68	2.19	2.11	2.20	2.13	2.24	2.17	1.46	1.38	1.31
MnO	0.06	0.07	0.01	0.05	0.04	0.07	0.05	0.12	0.07	0.07
Na2O	0.37	0.35	0.36	0.37	0.63	0.43	0.43	0.38	0.47	0.48
Al2O3	35.15	34.52	34.59	34.13	34.62	34.16	34.40	35.20	35.64	35.51
SiO2	46.53	46.79	45.87	45.91	46.25	45.69	46.10	46.94	45.53	46.51
MgO	0.36	0.45	0.41	0.49	0.48	0.51	0.47	0.36	0.28	0.30
K2O	11.47	11.44	11.47	11.56	11.20	11.61	11.46	11.41	11.75	11.45
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.36	0.30	0.22	0.33	0.15	0.31	0.26	0.26	0.33	0.20
F	0.35	0.35	0.29	0.64	0.44	0.57	0.46	0.37	0.35	0.36
-O=F	0.15	0.15	0.12	0.27	0.19	0.24	0.19	0.16	0.15	0.15
Tot.	96.18	96.33	95.21	95.41	95.76	95.35	95.61	96.74	95.16	96.04
K	1.951	1.948	1.976	2.001	1.922	2.012	1.972	1.935	1.932	1.948
Na	0.095	0.091	0.094	0.097	0.164	0.113	0.112	0.098	0.123	0.124
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.204	6.246	6.195	6.230	6.220	6.206	6.220	6.239	6.130	6.202
Al IV	1.796	1.754	1.805	1.770	1.780	1.794	1.780	1.761	1.870	1.798
Al VI	3.729	3.676	3.701	3.689	3.708	3.675	3.690	3.753	3.786	3.783
Fe	0.187	0.244	0.238	0.250	0.240	0.254	0.245	0.162	0.155	0.146
Mg	0.072	0.090	0.083	0.099	0.096	0.103	0.094	0.071	0.056	0.060
Mn	0.007	0.008	0.001	0.006	0.005	0.008	0.005	0.014	0.008	0.008
Ti	0.045	0.038	0.028	0.042	0.019	0.040	0.033	0.032	0.042	0.025
F	0.149	0.148	0.124	0.275	0.187	0.245	0.195	0.156	0.149	0.152

Sample	PK-43b	PK-43b	PK-43b	PK-58	PK-58	PK-58
Anal. #	4	5	ave	1	2	ave
FeO	1.36	1.44	1.39	1.42	1.28	1.35
MnO	0.10	0.09	0.09	0.08	0.07	0.08
Na2O	0.40	0.51	0.45	0.38	0.41	0.40
Al2O3	35.29	35.08	35.34	35.19	35.25	35.22
SiO2	46.06	45.56	46.12	46.42	46.34	46.38
MgO	0.31	0.31	0.31	0.34	0.32	0.33
K2O	11.51	11.35	11.39	11.34	11.32	11.33
CaO	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.41	0.46	0.33	0.32	0.31	0.32
F	0.23	0.38	0.34	0.57	0.08	0.33
-O=F	0.10	0.16	0.14	0.24	0.03	0.14
Tot.	95.58	95.03	95.63	95.83	95.36	95.59
K	1.966	1.956	1.947	1.939	1.929	1.934
Na	0.104	0.134	0.116	0.099	0.106	0.102
Ca	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.166	6.154	6.179	6.223	6.191	6.207
Al IV	1.834	1.846	1.821	1.777	1.809	1.793
Al VI	3.735	3.738	3.759	3.782	3.741	3.762
Fe	0.152	0.163	0.156	0.159	0.143	0.151
Mg	0.062	0.062	0.062	0.068	0.064	0.066
Mn	0.011	0.010	0.010	0.009	0.008	0.009
Ti	0.052	0.058	0.042	0.040	0.039	0.040
F	0.097	0.162	0.143	0.242	0.034	0.138

Pegmatite

Sample	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e	PK-17e
Anal. #	1	2	3	4	5	6	7	8	9	10
FeO	1.14	1.11	1.22	1.17	1.21	1.23	1.21	1.07	1.16	1.13
MnO	0.23	0.12	0.17	0.17	0.03	0.24	0.20	0.24	0.16	0.18
Na2O	0.67	0.67	0.70	0.86	0.81	0.70	0.49	0.85	0.83	0.82
Al2O3	35.31	35.78	35.27	35.91	35.55	35.73	35.61	35.54	35.37	35.53
SiO2	46.51	45.59	46.32	46.30	45.99	45.64	46.17	45.81	46.32	46.03
MgO	0.18	0.20	0.21	0.18	0.23	0.19	0.22	0.19	0.20	0.20
K2O	11.05	11.05	10.88	10.69	10.76	10.89	11.26	10.67	10.85	10.82
CaO	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
TiO2	0.60	0.54	0.58	0.36	0.57	0.58	0.55	0.51	0.60	0.47
F	0.91	1.06	1.12	1.13	0.91	0.29	0.84	0.50	1.14	1.16
-O=F	0.38	0.45	0.47	0.48	0.38	0.12	0.35	0.21	0.48	0.49
Tot.	96.22	95.67	96.00	96.28	95.67	95.38	96.19	95.15	96.15	95.85

K	1.888	1.905	1.869	1.829	1.847	1.861	1.925	1.830	1.861	1.863
Na	0.174	0.176	0.183	0.224	0.211	0.182	0.127	0.222	0.216	0.215
Ca	0.000	0.001	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000
Si	6.230	6.160	6.236	6.210	6.189	6.114	6.187	6.159	6.228	6.212
Al IV	1.770	1.840	1.764	1.790	1.811	1.886	1.813	1.841	1.772	1.788
Al VI	3.804	3.859	3.833	3.887	3.827	3.755	3.812	3.791	3.833	3.863
Fe	0.128	0.125	0.137	0.131	0.136	0.138	0.136	0.120	0.130	0.128
Mg	0.036	0.040	0.042	0.036	0.046	0.038	0.044	0.038	0.040	0.040
Mn	0.026	0.014	0.019	0.019	0.003	0.027	0.023	0.027	0.018	0.021
Ti	0.076	0.069	0.073	0.045	0.072	0.073	0.069	0.064	0.076	0.060
F	0.385	0.453	0.477	0.479	0.387	0.123	0.356	0.213	0.485	0.495

Sample	PK-17e	PK-31b	PK-31b	PK-31b	PK-33	PK-33	PK-33	PK-33	PK-38a	PK-38a
Anal. #	ave	1	2	ave	1	2	3	ave	1	2
FeO	1.17	1.35	1.36	1.36	1.48	1.30	1.25	1.34	1.03	0.98
MnO	0.17	0.14	0.17	0.16	0.11	0.12	0.16	0.13	0.12	0.08
Na2O	0.74	0.76	0.74	0.75	0.88	0.65	0.77	0.77	0.63	0.60
Al2O3	35.56	35.54	35.46	35.50	35.41	35.82	35.72	35.65	35.67	35.96
SiO2	46.07	46.64	46.62	46.63	46.65	46.67	46.55	46.62	45.57	46.00
MgO	0.20	0.22	0.24	0.23	0.24	0.19	0.25	0.23	0.17	0.19
K2O	10.89	11.05	10.82	10.94	10.77	11.09	10.97	10.94	10.89	11.19
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.54	0.55	0.62	0.59	0.54	0.57	0.55	0.55	0.47	0.45
F	0.91	0.67	0.50	0.59	0.37	0.53	0.56	0.49	0.71	0.40
-O=F	0.38	0.28	0.21	0.25	0.16	0.22	0.24	0.20	0.30	0.17
Tot.	95.86	96.64	96.32	96.48	96.28	96.73	96.53	96.52	94.95	95.66

K	1.868	1.874	1.834	1.854	1.824	1.874	1.858	1.852	1.878	1.907
Na	0.193	0.196	0.191	0.193	0.226	0.167	0.198	0.197	0.165	0.155
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.193	6.202	6.195	6.198	6.192	6.182	6.181	6.185	6.161	6.145
Al IV	1.807	1.798	1.805	1.802	1.808	1.818	1.819	1.815	1.839	1.855
Al VI	3.826	3.772	3.748	3.760	3.732	3.775	3.770	3.759	3.844	3.807
Fe	0.131	0.150	0.151	0.151	0.164	0.144	0.139	0.149	0.116	0.109
Mg	0.040	0.044	0.048	0.046	0.047	0.038	0.049	0.045	0.034	0.038
Mn	0.020	0.016	0.019	0.017	0.012	0.013	0.018	0.015	0.014	0.009
Ti	0.068	0.069	0.077	0.073	0.067	0.071	0.069	0.069	0.060	0.057
F	0.385	0.282	0.210	0.246	0.155	0.222	0.235	0.204	0.304	0.169

Sample	PK-38a	PK-38a	PK-38a	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-38d1	PK-43
Anal. #	3	4	ave	1	2	3	4	5	ave	1
FeO	1.07	1.07	1.04	2.64	2.38	2.33	2.21	2.10	2.31	1.42
MnO	0.13	0.15	0.12	0.03	0.06	0.05	0.06	0.03	0.05	0.03
Na2O	0.53	0.58	0.59	0.34	0.36	0.38	0.45	0.53	0.41	0.19
Al2O3	36.00	35.78	35.85	33.34	34.14	33.49	33.84	34.66	33.89	34.99
SiO2	46.18	46.16	45.98	46.43	45.85	46.19	46.02	46.20	46.14	47.13
MgO	0.16	0.17	0.17	0.60	0.51	0.61	0.61	0.55	0.58	0.25
K2O	11.18	11.14	11.10	11.39	11.49	11.44	11.36	11.23	11.38	10.83
CaO	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00
TiO2	0.48	0.47	0.47	0.37	0.31	0.34	0.28	0.15	0.29	0.57
F	0.52	0.30	0.48	0.47	0.40	0.46	0.52	0.56	0.48	0.52
-O=F	0.22	0.13	0.20	0.20	0.17	0.19	0.22	0.24	0.20	0.22
Tot.	96.03	95.70	95.59	95.40	95.32	95.10	95.11	95.79	95.35	96.31
K	1.902	1.896	1.896	1.967	1.985	1.981	1.967	1.929	1.966	1.835
Na	0.137	0.150	0.152	0.089	0.095	0.100	0.118	0.138	0.108	0.203
Ca	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Si	6.158	6.157	6.155	6.287	6.208	6.270	6.247	6.222	6.247	6.260
Al IV	1.842	1.843	1.845	1.713	1.792	1.730	1.753	1.778	1.753	1.740
Al VI	3.816	3.783	3.812	3.607	3.656	3.628	3.662	3.723	3.655	3.738
Fe	0.119	0.119	0.116	0.299	0.269	0.265	0.251	0.237	0.264	0.158
Mg	0.032	0.034	0.034	0.121	0.103	0.123	0.123	0.110	0.116	0.049
Mn	0.015	0.017	0.014	0.003	0.007	0.006	0.007	0.003	0.005	0.003
Ti	0.060	0.059	0.059	0.047	0.039	0.043	0.036	0.019	0.037	0.071
F	0.219	0.127	0.204	0.201	0.171	0.197	0.223	0.239	0.206	0.218

Sample	PK-43	PK-43	PK-43	PK-43	PK-43	PK-43	PK-43	PK-43
Anal. #	2	3	4	5	6	7	8	ave
FeO	1.46	1.54	1.44	1.56	1.54	1.38	1.53	1.48
MnO	0.06	0.07	0.05	0.06	0.09	0.03	0.10	0.06
Na2O	0.74	0.50	0.58	0.54	0.54	0.76	0.35	0.60
Al2O3	35.32	35.03	35.11	35.24	34.85	35.45	34.98	35.12
SiO2	46.86	47.09	46.90	46.64	46.23	46.13	46.77	46.72
MgO	0.25	0.32	0.31	0.30	0.30	0.24	0.30	0.28
K2O	10.67	11.19	11.12	11.26	10.97	11.02	11.55	11.08
CaO	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.62	0.55	0.48	0.61	0.64	0.52	0.60	0.57
F	0.55	0.47	0.37	0.53	0.52	0.51	0.40	0.48
-O=F	0.23	0.20	0.16	0.22	0.22	0.21	0.17	0.20
Tot.	96.29	96.59	96.20	96.51	95.46	95.82	96.40	96.20
K	1.808	1.893	1.885	1.911	1.880	1.881	1.960	1.882
Na	0.191	0.129	0.149	0.139	0.141	0.197	0.090	0.155
Ca	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.224	6.246	6.232	6.204	6.210	6.173	6.221	6.221
Al IV	1.776	1.754	1.768	1.796	1.790	1.827	1.779	1.779
Al VI	3.754	3.722	3.731	3.728	3.727	3.764	3.705	3.734
Fe	0.162	0.171	0.160	0.174	0.173	0.154	0.170	0.165
Mg	0.049	0.063	0.061	0.059	0.060	0.048	0.059	0.056
Mn	0.007	0.008	0.006	0.007	0.010	0.003	0.011	0.007
Ti	0.077	0.069	0.060	0.076	0.081	0.065	0.075	0.072
F	0.231	0.197	0.156	0.223	0.221	0.216	0.168	0.204

Replacement

Sample	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-45	PK-45	PK-45
Anal. #	1	2	3	4	5	6	ave	1	2	3
FeO	2.97	3.56	3.10	4.72	3.96	1.77	3.35	1.84	1.76	1.77
MnO	0.08	0.08	0.07	0.19	0.09	0.11	0.10	0.12	0.07	0.07
Na2O	0.24	0.20	0.20	0.17	0.16	0.55	0.25	0.36	0.61	0.33
Al2O3	31.77	33.39	31.23	31.75	31.97	35.33	32.57	33.77	34.02	34.34
SiO2	47.96	47.30	48.42	46.83	47.52	46.97	47.50	46.30	46.26	47.02
MgO	1.33	0.28	1.48	0.64	0.57	0.52	0.80	0.53	0.58	0.45
K2O	11.54	11.39	11.42	11.23	11.15	10.84	11.26	11.37	10.88	11.48
CaO	0.01	0.01	0.00	0.00	0.01	0.00	0.01	0.01	0.00	0.00
TiO2	0.12	0.25	0.13	0.14	0.21	0.18	0.17	0.81	0.83	0.45
F	0.69	0.21	0.84	0.27	0.26	0.15	0.40	0.22	0.16	0.26
-O=F	0.29	0.09	0.35	0.11	0.11	0.06	0.17	0.09	0.07	0.11
Tot.	96.42	96.58	96.53	95.83	95.78	96.36	96.25	95.20	95.10	96.05
K	1.983	1.941	1.963	1.945	1.919	1.829	1.930	1.951	1.863	1.952
Na	0.063	0.052	0.052	0.045	0.042	0.141	0.066	0.094	0.159	0.085
Ca	0.001	0.001	0.000	0.000	0.001	0.000	0.001	0.001	0.000	0.000
Si	6.459	6.318	6.523	6.359	6.411	6.214	6.380	6.229	6.209	6.268
Al IV	1.541	1.682	1.477	1.641	1.589	1.786	1.620	1.771	1.791	1.732
Al VI	3.502	3.575	3.482	3.440	3.495	3.723	3.537	3.584	3.592	3.663
Fe	0.335	0.398	0.349	0.536	0.447	0.196	0.376	0.207	0.198	0.197
Mg	0.267	0.056	0.297	0.130	0.115	0.103	0.161	0.106	0.116	0.089
Mn	0.009	0.009	0.008	0.022	0.010	0.012	0.012	0.014	0.008	0.008
Ti	0.015	0.031	0.016	0.018	0.027	0.022	0.022	0.102	0.105	0.056
F	0.294	0.089	0.358	0.116	0.111	0.063	0.171	0.094	0.068	0.110

Sample	PK-45	PK-45	PK-45	PK-49	PK-49	PK-49	PK-49	PK-49	PK-49	PK-62
Anal. #	4	5	ave	1	2	3	4	5	ave	1
FeO	1.79	1.85	1.80	2.22	2.53	2.50	1.93	2.35	2.31	1.53
MnO	0.08	0.10	0.09	0.04	0.14	0.12	0.05	0.14	0.10	0.01
Na2O	0.53	0.44	0.45	0.55	0.46	0.30	0.59	0.34	0.45	0.30
Al2O3	33.71	33.35	33.84	33.66	33.25	33.21	34.76	33.79	33.73	35.53
SiO2	46.62	47.01	46.64	46.93	46.28	46.22	45.69	45.71	46.17	45.96
MgO	0.59	0.51	0.53	0.75	0.78	0.79	0.64	0.71	0.73	0.31
K2O	11.14	11.11	11.20	11.17	11.20	11.36	11.02	11.17	11.18	11.64
CaO	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.02
TiO2	0.76	0.75	0.72	0.40	0.32	0.43	0.03	0.29	0.29	0.10
F	0.44	0.28	0.27	0.35	0.51	0.26	0.40	0.51	0.41	0.20
-O=F	0.19	0.12	0.11	0.15	0.21	0.11	0.17	0.21	0.17	0.08
Tot.	95.46	95.26	95.42	95.92	95.26	95.08	94.94	94.80	95.20	95.50
K	1.911	1.903	1.916	1.909	1.939	1.962	1.904	1.941	1.931	1.991
Na	0.138	0.115	0.118	0.143	0.121	0.079	0.155	0.090	0.118	0.078
Ca	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.003
Si	6.268	6.312	6.257	6.286	6.280	6.259	6.187	6.227	6.248	6.162
Al IV	1.732	1.688	1.743	1.714	1.720	1.741	1.813	1.773	1.752	1.838
Al VI	3.610	3.589	3.608	3.600	3.598	3.560	3.735	3.652	3.629	3.776
Fe	0.201	0.208	0.202	0.249	0.287	0.283	0.219	0.268	0.261	0.172
Mg	0.119	0.102	0.106	0.150	0.158	0.159	0.129	0.144	0.148	0.062
Mn	0.009	0.011	0.010	0.005	0.016	0.014	0.006	0.016	0.011	0.001
Ti	0.096	0.095	0.091	0.050	0.041	0.055	0.004	0.037	0.037	0.013
F	0.187	0.119	0.115	0.148	0.219	0.111	0.171	0.220	0.174	0.085

Sample	PK-62	PK-62	PK-62	PK-62b	PK-62b	PK-62b	PK-62b
Anal. #	2	3	ave	1	2	3	ave
FeO	2.59	2.00	2.04	2.01	1.87	1.81	1.90
MnO	0.07	0.03	0.04	0.06	0.08	0.09	0.08
Na2O	0.26	0.22	0.26	0.28	0.31	0.33	0.31
Al2O3	33.17	34.30	34.33	33.88	33.89	33.71	33.83
SiO2	46.88	46.13	46.32	46.46	46.83	46.95	46.75
MgO	0.92	0.59	0.61	0.66	0.67	0.77	0.70
K2O	11.57	11.66	11.62	11.34	11.12	11.28	11.25
CaO	0.01	0.01	0.01	0.00	0.00	0.00	0.00
TiO2	0.05	0.12	0.09	0.22	0.33	0.40	0.32
F	0.30	0.14	0.21	0.34	0.28	0.45	0.36
-O=F	0.13	0.06	0.09	0.14	0.12	0.19	0.15
Tot.	95.69	95.13	95.44	95.12	95.25	95.59	95.32
K	1.989	2.006	1.995	1.953	1.906	1.933	1.931
Na	0.068	0.058	0.068	0.073	0.081	0.086	0.080
Ca	0.001	0.001	0.002	0.000	0.000	0.000	0.000
Si	6.317	6.221	6.233	6.274	6.291	6.306	6.290
Al IV	1.683	1.779	1.767	1.726	1.709	1.694	1.710
Al VI	3.585	3.672	3.678	3.666	3.656	3.643	3.655
Fe	0.292	0.226	0.230	0.227	0.210	0.203	0.213
Mg	0.185	0.119	0.122	0.133	0.134	0.154	0.140
Mn	0.008	0.003	0.004	0.007	0.009	0.010	0.009
Ti	0.006	0.015	0.011	0.028	0.042	0.051	0.040
F	0.128	0.060	0.091	0.145	0.119	0.191	0.152

Secondary, muscovite replacement of apatite-pegmatite feldspar

Sample	PK-12a	PK-31b	PK-31b	PK-31b	PK-31b	PK-31b	PK-33b	PK-33b	PK-33b	PK-33b
Anal. #	1	1	2	3	4	ave	1	2	3	ave
FeO	4.37	1.29	1.53	0.98	1.68	1.37	3.43	2.72	2.57	2.91
MnO	0.11	0.06	0.07	0.01	0.06	0.05	0.12	0.07	0.04	0.08
Na2O	0.22	0.21	0.31	0.20	0.43	0.29	0.30	0.24	0.21	0.25
Al2O3	31.12	36.60	36.30	36.71	35.67	36.32	33.21	33.62	34.03	33.62
SiO2	47.12	46.64	46.63	46.60	46.64	46.63	45.80	46.62	46.80	46.41
MgO	1.05	0.04	0.10	0.01	0.06	0.05	0.43	0.30	0.20	0.31
K2O	11.07	11.51	11.60	11.46	11.51	11.52	11.41	11.26	11.59	11.42
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.01
TiO2	0.16	0.04	0.07	0.02	0.07	0.05	0.30	0.21	0.19	0.23
F	0.48	0.21	0.24	0.13	0.15	0.18	0.36	0.25	0.31	0.31
-O=F	0.20	0.09	0.10	0.05	0.06	0.08	0.15	0.11	0.13	0.13
Tot.	95.49	96.51	96.76	96.08	96.21	96.39	95.22	95.17	95.82	95.40
K	1.925	1.940	1.957	1.934	1.951	1.946	1.983	1.940	1.987	1.970
Na	0.058	0.054	0.079	0.051	0.111	0.074	0.079	0.063	0.055	0.066
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.003	0.000	0.000	0.001
Si	6.423	6.164	6.166	6.166	6.198	6.173	6.239	6.297	6.289	6.275
Al IV	1.577	1.836	1.834	1.834	1.802	1.827	1.761	1.703	1.711	1.725
Al VI	3.423	3.865	3.823	3.890	3.786	3.841	3.570	3.650	3.679	3.633
Fe	0.498	0.143	0.169	0.108	0.187	0.152	0.391	0.307	0.289	0.329
Mg	0.213	0.008	0.020	0.002	0.012	0.010	0.087	0.060	0.040	0.062
Mn	0.013	0.007	0.008	0.001	0.007	0.006	0.014	0.008	0.005	0.009
Ti	0.021	0.005	0.009	0.002	0.009	0.006	0.038	0.027	0.024	0.030
F	0.207	0.088	0.100	0.054	0.063	0.076	0.155	0.107	0.132	0.131

Sample	PK-35b	PK-35b	PK-35b	PK-35b	PK-35b	PK-36	PK-36	PK-36	PK-50b	PK-50b
Anal. #	1	2	3	4	ave	1	2	ave	1	2
FeO	1.96	0.95	0.23	3.61	1.69	3.70	3.61	3.66	3.00	2.93
MnO	0.09	0.03	0.00	0.14	0.07	0.09	0.11	0.10	0.11	0.05
Na2O	0.31	0.38	0.32	0.31	0.33	0.17	0.20	0.19	0.23	0.21
Al2O3	32.95	36.64	37.19	33.86	35.16	32.33	32.99	32.66	32.49	32.41
SiO2	47.82	46.99	47.42	46.43	47.17	47.44	46.96	47.20	46.99	47.11
MgO	0.96	0.08	0.05	0.12	0.30	0.67	0.21	0.44	0.72	0.73
K2O	11.19	11.10	11.00	11.40	11.17	11.43	11.29	11.36	11.52	11.43
CaO	0.04	0.04	0.02	0.00	0.03	0.00	0.00	0.00	0.00	0.00
TiO2	0.06	0.02	0.02	0.02	0.03	0.19	0.29	0.24	0.21	0.18
F	0.34	0.16	0.05	0.21	0.19	0.32	0.31	0.32	0.31	0.57
-O=F	0.14	0.07	0.02	0.09	0.08	0.13	0.13	0.13	0.13	0.24
Tot.	95.57	96.33	96.28	96.01	96.05	96.21	95.84	96.02	95.46	95.38
K	1.914	1.866	1.836	1.959	1.893	1.962	1.942	1.952	1.989	1.980
Na	0.081	0.097	0.081	0.081	0.085	0.044	0.052	0.048	0.060	0.055
Ca	0.006	0.006	0.003	0.000	0.004	0.000	0.000	0.000	0.000	0.000
Si	6.411	6.191	6.204	6.255	6.265	6.383	6.334	6.358	6.359	6.396
Al IV	1.589	1.809	1.796	1.745	1.735	1.617	1.666	1.642	1.641	1.604
Al VI	3.618	3.881	3.938	3.630	3.769	3.510	3.578	3.544	3.541	3.583
Fe	0.220	0.105	0.025	0.407	0.187	0.416	0.407	0.412	0.340	0.333
Mg	0.192	0.016	0.010	0.024	0.060	0.134	0.042	0.088	0.145	0.148
Mn	0.010	0.003	0.000	0.016	0.007	0.010	0.013	0.011	0.013	0.006
Ti	0.008	0.002	0.002	0.003	0.004	0.024	0.037	0.030	0.027	0.023
F	0.144	0.067	0.021	0.089	0.080	0.136	0.132	0.134	0.133	0.245

Sample	PK-50b	PK-50b	PK-50b	PK-42b	PK-62	PK-62	PK-62	PK-62	PK-62b	PK-62b
Anal. #	3	4	ave	1	1	2	3	ave	1	2
FeO	3.58	3.51	3.26	2.70	4.60	2.56	2.78	3.31	3.66	2.87
MnO	0.10	0.14	0.10	0.03	0.11	0.07	0.06	0.08	0.05	0.13
Na2O	0.21	0.23	0.22	0.15	0.21	0.30	0.25	0.25	0.21	0.21
Al2O3	32.67	32.33	32.48	33.37	32.13	34.40	32.23	32.92	32.74	31.50
SiO2	46.33	46.29	46.68	47.34	46.86	45.90	48.10	46.95	47.20	47.70
MgO	0.54	0.63	0.66	0.37	0.41	0.49	0.58	0.49	0.44	0.94
K2O	11.59	11.49	11.51	11.45	11.49	11.71	11.57	11.59	11.16	11.28
CaO	0.00	0.00	0.00	0.02	0.01	0.00	0.06	0.02	0.03	0.01
TiO2	0.14	0.15	0.17	0.07	0.03	0.19	0.01	0.08	0.07	0.24
F	0.32	0.35	0.39	0.30	0.45	0.34	0.34	0.38	0.30	0.48
-O=F	0.13	0.15	0.16	0.13	0.19	0.14	0.14	0.16	0.13	0.20
Tot.	95.36	94.99	95.30	95.67	96.12	95.83	95.84	95.93	95.74	95.11
K	2.012	2.003	1.996	1.964	1.991	2.015	1.985	1.997	1.921	1.953
Na	0.055	0.061	0.058	0.039	0.055	0.078	0.065	0.066	0.055	0.055
Ca	0.000	0.000	0.000	0.003	0.001	0.000	0.009	0.003	0.004	0.001
Si	6.305	6.325	6.346	6.365	6.365	6.190	6.469	6.341	6.369	6.475
Al IV	1.695	1.675	1.654	1.635	1.635	1.810	1.531	1.659	1.631	1.525
Al VI	3.545	3.531	3.550	3.652	3.509	3.658	3.577	3.582	3.575	3.515
Fe	0.407	0.401	0.370	0.304	0.523	0.289	0.313	0.374	0.413	0.320
Mg	0.110	0.128	0.133	0.074	0.083	0.099	0.116	0.099	0.088	0.190
Mn	0.012	0.016	0.012	0.003	0.013	0.008	0.007	0.009	0.006	0.015
Ti	0.018	0.019	0.022	0.009	0.004	0.024	0.001	0.010	0.009	0.031
F	0.138	0.151	0.167	0.128	0.193	0.145	0.145	0.161	0.128	0.206

Sample	PK-62b	PK-62b
Anal. #	3	ave
FeO	2.67	3.05
MnO	0.07	0.08
Na2O	0.20	0.21
Al2O3	32.65	32.30
SiO2	47.38	47.43
MgO	0.67	0.68
K2O	11.03	11.16
CaO	0.01	0.02
TiO2	0.05	0.12
F	0.37	0.38
-O=F	0.16	0.16
Tot.	94.95	95.27
K	1.905	1.926
Na	0.052	0.054
Ca	0.001	0.002
Si	6.414	6.419
Al IV	1.586	1.581
Al VI	3.623	3.571
Fe	0.302	0.345
Mg	0.135	0.138
Mn	0.008	0.010
Ti	0.006	0.015
F	0.158	0.164

Vein, muscovite replacement of vein feldspar

Sample	PK-35b	PK-35b	PK-35b	PK-35b	PK-35b	PK-35b	PK-51	PK-51	PK-51	PK-51
Anal. #	1	2	3	4	5	ave	1	2	3	4
FeO	2.37	2.33	2.14	2.18	2.16	2.24	2.62	2.34	2.41	2.30
MnO	0.09	0.12	0.08	0.08	0.08	0.09	0.10	0.11	0.08	0.12
Na2O	0.41	0.42	0.47	0.45	0.39	0.43	0.27	0.43	0.52	0.23
Al2O3	32.96	33.56	33.60	33.69	33.85	33.53	32.65	33.58	33.66	32.06
SiO2	46.65	46.92	46.09	46.46	46.13	46.45	46.86	46.25	46.38	46.69
MgO	0.76	0.69	0.63	0.66	0.62	0.67	0.73	0.63	0.67	1.17
K2O	11.13	11.31	11.22	11.18	11.12	11.19	11.58	11.39	11.10	11.54
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.49	0.46	0.54	0.59	0.55	0.53	0.48	0.38	0.41	0.59
F	0.72	0.33	0.54	0.32	0.18	0.42	0.11	0.35	0.33	0.41
-O=F	0.30	0.14	0.23	0.13	0.08	0.18	0.05	0.15	0.14	0.17
Tot.	95.27	96.02	95.08	95.47	95.00	95.37	95.37	95.31	95.61	94.95

K	1.928	1.933	1.942	1.919	1.913	1.927	1.992	1.965	1.909	2.001
Na	0.108	0.109	0.124	0.117	0.102	0.112	0.071	0.113	0.136	0.061
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.334	6.286	6.253	6.251	6.221	6.269	6.318	6.254	6.253	6.346
Al IV	1.666	1.714	1.747	1.749	1.779	1.731	1.682	1.746	1.747	1.654
Al VI	3.608	3.585	3.626	3.593	3.602	3.603	3.506	3.606	3.601	3.481
Fe	0.269	0.261	0.243	0.245	0.244	0.252	0.295	0.265	0.272	0.261
Mg	0.154	0.138	0.127	0.132	0.125	0.135	0.147	0.127	0.135	0.237
Mn	0.010	0.014	0.009	0.009	0.009	0.010	0.011	0.013	0.009	0.014
Ti	0.063	0.058	0.069	0.075	0.070	0.067	0.061	0.048	0.052	0.075
F	0.309	0.140	0.232	0.136	0.077	0.178	0.047	0.150	0.141	0.176

Sample	PK-51	PK-51	PK-51	PK-51	PK-60b	PK-60b	PK-60b	PK-60b	PK-60b	PK-60b
Anal. #	5	6	7	ave	1	2	3	4	5	6
FeO	2.52	2.74	3.67	2.66	3.94	3.22	2.57	2.53	4.25	4.57
MnO	0.13	0.11	0.19	0.12	0.11	0.01	0.19	0.10	0.05	0.06
Na2O	0.27	0.30	0.29	0.33	0.17	0.15	0.52	0.36	0.17	0.13
Al2O3	33.07	32.19	30.69	32.56	32.53	32.13	32.83	32.58	32.28	30.99
SiO2	46.10	46.56	46.83	46.52	46.58	46.80	46.18	46.89	45.95	46.17
MgO	0.56	0.82	0.96	0.79	0.19	0.98	0.81	0.89	0.29	0.72
K2O	11.60	11.43	11.36	11.43	11.81	11.65	11.26	11.41	11.61	11.70
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO2	0.57	0.48	0.27	0.45	0.42	0.19	0.74	0.70	0.39	0.31
F	0.53	0.47	1.02	0.46	0.19	0.50	0.32	0.41	0.23	0.39
-O=F	0.22	0.20	0.43	0.19	0.08	0.21	0.13	0.17	0.10	0.16
Tot.	95.12	94.90	94.84	95.16	95.86	95.41	95.30	95.70	95.10	94.88

K	2.014	1.988	2.005	1.982	2.039	2.022	1.945	1.962	2.025	2.057
Na	0.071	0.079	0.078	0.087	0.045	0.040	0.137	0.094	0.045	0.035
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Si	6.276	6.347	6.480	6.324	6.305	6.367	6.252	6.321	6.281	6.363
Al IV	1.724	1.653	1.520	1.676	1.695	1.633	1.748	1.679	1.719	1.637
Al VI	3.581	3.520	3.484	3.540	3.495	3.519	3.491	3.497	3.482	3.397
Fe	0.287	0.312	0.425	0.302	0.446	0.366	0.291	0.285	0.486	0.527
Mg	0.114	0.167	0.198	0.160	0.038	0.199	0.163	0.179	0.059	0.148
Mn	0.015	0.013	0.022	0.014	0.013	0.001	0.022	0.011	0.006	0.007
Ti	0.073	0.062	0.035	0.058	0.053	0.024	0.094	0.089	0.050	0.040
F	0.228	0.203	0.446	0.198	0.081	0.215	0.137	0.175	0.099	0.170

Sample	PK-60b
Anal. #	ave
FeO	3.51
MnO	0.09
Na2O	0.25
Al2O3	32.22
SiO2	46.43
MgO	0.65
K2O	11.57
CaO	0.00
TiO2	0.46
F	0.34
-O=F	0.14
Tot.	95.37

K	2.008
Na	0.066
Ca	0.000
Si	6.315
Al IV	1.685
Al VI	3.480
Fe	0.400
Mg	0.131
Mn	0.010
Ti	0.059
F	0.146

APPENDIX II*d*
**ELECTRON MICROPROBE ANALYSES OF GARNET FROM NONG
SUA**

The abbreviations used are: c core, r rim, m middle, FI fluid inclusions in core, sg same grain, incl inclusion in

Sample	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c
Anal. #	1	2	3	4	5	6	7	8	9	10	11	12	13
Descript.	c FI	r	m	c	c	c	c	r	c FI	r	m	m	m
	sg 1	sg 1	sg 1	sg 1	sg 1	sg 6	sg 6	sg 6	sg 9	sg 9	sg 9	sg 9	sg 9
FeO	24.3	26.39	25.59	25.19	24.84	25.46	25.7	26.76	22.42	26.55	24.3	22.95	22.51
MnO	17.56	15.86	16.72	17.26	17.46	16.85	16.64	15.66	20.17	15.72	18.45	19.77	19.31
TiO2	0.07	0	0	0.06	0.04	0.02	0.01	0.01	0.04	0	0.03	0.09	0.06
Na2O	0.03	0	0.01	0.04	0.05	0	0	0.01	0.04	0.02	0	0.03	0
SiO2	35.9	37.47	36.89	36.81	36.02	37.27	37.39	37.41	35.6	36.63	36.63	36.33	34.96
Al2O3	20.68	20.6	20.72	20.72	20.87	20.54	20.63	20.71	20.73	20.84	20.67	20.61	20.17
MgO	0.47	0.37	0.42	0.42	0.49	0.41	0.41	0.38	0.41	0.35	0.35	0.36	0.36
CaO	0.65	0.61	0.52	0.42	0.66	0.73	0.65	0.62	0.62	0.59	0.53	0.44	0.45
SnO2	0.09	0	0	0	0.1	0.02	0	0	0.11	0	0.01	0.04	0
Total	99.74	101.2	100.8	100.9	100.5	101.3	101.4	101.5	100.1	100.7	100.9	100.6	97.83
Fe	1.679	1.788	1.745	1.718	1.706	1.727	1.740	1.810	1.549	1.814	1.660	1.575	1.592
Mn	1.229	1.088	1.153	1.192	1.213	1.159	1.141	1.073	1.413	1.088	1.276	1.275	1.384
Mg	0.058	0.043	0.049	0.052	0.060	0.049	0.049	0.045	0.050	0.044	0.042	0.044	0.046
Ca	0.058	0.051	0.045	0.036	0.058	0.063	0.055	0.053	0.054	0.052	0.046	0.038	0.040
Na	0.004	0	0.001	0.006	0.008	0	0	0.001	0.006	0.004	0	0.004	0
Al	2.010	1.988	1.991	2.008	1.988	1.991	2.006	2.020	2.044	2.022	2.011	1.987	2.005
Si	2.968	3.036	3.007	3.002	2.958	3.024	3.027	3.025	2.942	2.994	2.991	2.980	2.956
Ti	0.004	0	0	0.004	0.002	0.001	0	0	0.002	0	0.002	0.006	0.004
XAlm	0.555	0.601	0.582	0.573	0.561	0.575	0.582	0.606	0.505	0.605	0.548	0.519	0.519
XSpess	0.406	0.366	0.385	0.397	0.399	0.386	0.381	0.359	0.460	0.362	0.421	0.453	0.451
XGros	0.019	0.017	0.015	0.012	0.019	0.021	0.018	0.018	0.017	0.017	0.015	0.012	0.013
XPyr	0.019	0.014	0.016	0.017	0.019	0.016	0.016	0.015	0.016	0.014	0.013	0.014	0.015

Sample	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c	PK-3c
Anal. #	14	15	16	17	18	19	20	21	22	23	24	25	26
Descript.	m	r	c FI	c	c	c	c	c	c	r	r	m	m
	sg 9	sg 9	c FI	sg 16	sg 16	sg 16	sg 16	sg 16	sg 16	sg 16	sg 16	sg 16	sg 16
FeO	24.88	26.63	11.14	16.28	15.38	14.56	14.22	26.78	26.67	26.35	26.47	25.88	23.65
MnO	17.88	16.12	31.38	26.91	26.85	27.73	27.67	16.24	16.15	16.11	15.9	16.53	18.82
TiO2	0.02	0.02	0.04	0	0.15	0.06	0.14	0.04	0	0.01	0	0.02	0.03
Na2O	0.01	0	0	0	0.04	0.01	0.02	0	0	0.01	0	0	0
SiO2	36.64	36.53	36.06	35.75	34.7	35.19	34.9	36.57	36.53	37.03	37.08	36.52	36.72
Al2O3	20.66	20.69	20.59	20.75	20.71	20.56	20.29	20.68	20.82	20.71	20.48	20.61	20.72
MgO	0.34	0.37	0.12	0.06	0.28	0.25	0.25	0.35	0.35	0.39	0.37	0.36	0.32
CaO	0.63	0.62	0.3	0.19	0.68	0.57	0.7	0.6	0.59	0.57	0.54	0.62	0.59
SnO2	0	0	0	0.02	0.17	0.11	0.16	0	0	0.01	0	0.02	0
Total	101.0	100.9	99.63	99.94	98.95	99.04	98.35	101.2	101.1	101.1	100.8	100.5	100.8
Fe	1.699	1.819	0.770	1.177	1.079	1.018	1.002	1.827	1.819	1.791	1.805	1.774	1.614
Mn	1.236	1.116	2.201	1.888	1.907	1.963	1.973	1.123	1.116	1.109	1.099	1.148	1.302
Mg	0.040	0.046	0.016	0.008	0.034	0.032	0.032	0.042	0.042	0.047	0.043	0.044	0.038
Ca	0.054	0.054	0.026	0.016	0.060	0.050	0.064	0.052	0.052	0.049	0.047	0.054	0.052
Na	0.002	0	0	0	0.006	0.002	0.004	0	0	0.001	0	0	0
Al	1.987	1.971	1.990	1.996	1.997	1.989	2.008	2.002	1.971	1.949	1.991	1.982	1.968
Si	2.991	2.986	2.985	2.961	2.908	2.941	2.939	2.982	2.982	3.011	3.024	2.994	2.998
Ti	0.002	0.002	0.002	0	0.010	0.004	0.008	0.002	0	0	0	0.002	0.002
XAlm	0.560	0.599	0.255	0.370	0.350	0.332	0.326	0.600	0.600	0.597	0.607	0.587	0.536
XSpess	0.408	0.367	0.730	0.621	0.619	0.640	0.642	0.368	0.368	0.37	0.366	0.380	0.433
XGros	0.017	0.017	0.008	0.005	0.019	0.016	0.020	0.017	0.017	0.016	0.016	0.017	0.017
XPyr	0.013	0.015	0.005	0.002	0.011	0.010	0.010	0.013	0.013	0.016	0.014	0.014	0.012

Sample	PK-3c	PK-3c	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a
Anal. #	27	28	1	2	3	4	5	6	7	8	9	10	11
Descript.	m	m	r	r	r	c	c	c	r	c	m	m	m
	sg 1	sg 1	sg 1	sg 1	sg 3	sg 3	sg 3	sg 3	sg 7	sg 7	sg 7	sg 7	sg 7
FeO	25.7	23.75	23.31	22.67	22.13	25.95	16.17	14.72	23.26	24.35	16.95	19.62	17.86
MnO	16.61	18.76	20.62	20.08	19.68	15.7	24.72	25	18.57	17.78	24.42	21.75	23
TiO2	0.02	0.02	0	0.02	0.01	0.02	0.09	0.06	0.02	0	0.17	0.05	0.11
Na2O	0.03	0	0.01	0.02	0.01	0.02	0.03	0.02	0.02	0	0	0.01	0.02
SiO2	36.51	36.65	36.69	36.81	37.69	37.58	37.17	37.68	37.52	37.28	37.04	37.37	36.71
Al2O3	20.62	20.61	20.85	20.81	20.52	20.19	20.55	20.53	20.43	20.45	20.47	20.19	20.5
MgO	0.38	0.32	0.11	0.13	0.12	0.23	0.12	0.1	0.13	0.15	0.07	0.08	0.13
CaO	0.58	0.68	0.56	0.55	0.62	0.5	0.71	0.75	0.54	0.54	0.64	0.61	0.7
SnO2	0.02	0.02	0	0	0.01	0	0.01	0	0.01	0	0	0.02	0.02
Total	100.4	100.8	101.1	101.0	100.7	100.1	99.57	99.85	100.5	100.5	99.76	99.69	99.05
Fe	1.762	1.624	1.522	1.546	1.502	1.774	1.109	1.004	1.586	1.663	1.163	1.346	1.233
Mn	1.154	1.300	1.424	1.386	1.353	1.087	1.717	1.796	1.283	1.230	1.698	1.512	1.609
Mg	0.046	0.040	0.014	0.016	0.013	0.027	0.013	0.011	0.015	0.017	0.007	0.009	0.015
Ca	0.050	0.060	0.048	0.048	0.053	0.043	0.061	0.065	0.047	0.047	0.055	0.053	0.061
Na	0.004	0	0.002	0.002	0.001	0.003	0.003	0.003	0.003	0	0	0.001	0.003
Al	1.972	1.984	1.958	2	1.963	1.985	1.979	1.955	1.958	1.989	1.958	1.961	1.958
Si	2.995	2.996	2.992	3	3.061	3.073	3.050	3.074	3.059	3.043	3.041	3.068	3.034
Ti	0.002	0.002	0	0.002	0	0	0.005	0.003	0.001	0	0.009	0.001	0.005
XAlm	0.584	0.537	0.505	0.516	0.513	0.604	0.382	0.349	0.540	0.562	0.397	0.460	0.422
XSpess	0.383	0.429	0.473	0.462	0.462	0.370	0.591	0.623	0.437	0.415	0.580	0.517	0.550
XGros	0.016	0.019	0.015	0.016	0.018	0.014	0.021	0.022	0.016	0.016	0.019	0.018	0.021
XPyr	0.015	0.013	0.004	0.005	0.004	0.009	0.004	0.004	0.005	0.006	0.002	0.003	0.005

Sample	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-31a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a	PK-43a
Anal. #	12	13	14	15	16	17	1	2	3	4	5	6	7				
Descript.	m	m	r	c	c	r	incl	incl	incl	incl	incl	incl	incl				
FeO	19.65	16.97	23.47	18.05	18.82	18.55	24.83	23.44	25.01	24.01	25.47	23.79	24.39				
MnO	21.72	24.11	18.26	23.28	22.66	21.16	16.9	18.33	16.53	17.93	16.42	18.14	17.65				
TiO2	0.18	0.12	0	0.06	0.07	0.03	0.04	0.04	0	0.01	0	0.05	0.01				
Na2O	0.02	0.02	0.02	0.01	0.01	0.02	0.01	0.04	0.04	0.01	0	0.03	0.02				
SiO2	36.99	37.26	37.18	37.43	37.37	36.98	37.79	37.52	37.74	37.33	37.53	37.86	37.79				
Al2O3	20.27	20.6	20.51	20.27	20.31	20.54	20.47	20.4	20.42	20.29	20.27	20.81	20.48				
MgO	0.19	0.1	0.14	0.13	0.14	0.04	0.18	0.11	0.22	0.16	0.21	0.13	0.17				
CaO	0.61	0.75	0.51	0.68	0.7	2.09	0.68	0.51	0.69	0.51	0.58	0.55	0.55				
SnO2	0.03	0.01	0	0	0.02	0.01	0.01	0.01	0	0	0.02	0	0.01				
Total	99.65	99.93	100.0	99.93	100.1	99.41	100.9	100.4	100.6	100.2	100.4	101.3	101.0				
Fe	1.351	1.161	1.607	1.237	1.287	1.273	1.684	1.600	1.700	1.643	1.738	1.606	1.652				
Mn	1.513	1.670	1.267	1.614	1.570	1.471	1.161	1.267	1.137	1.243	1.135	1.241	1.211				
Mg	0.021	0.011	0.015	0.015	0.017	0.003	0.021	0.013	0.025	0.017	0.025	0.015	0.019				
Ca	0.053	0.065	0.043	0.005	0.061	0.183	0.059	0.043	0.059	0.043	0.049	0.047	0.047				
Na	0.003	0.003	0.001	0.001	0.001	0.003	0.001	0.005	0.005	0.001	0	0.003	0.003				
Al	1.956	1.949	1.980	1.957	1.954	1.984	1.949	1.959	1.985	1.938	1.998	2.007	1.969				
Si	3.042	3.046	3.046	3.064	3.057	3.036	3.064	3.061	3.067	3.054	3.062	3.056	3.063				
Ti	0.011	0.007	0	0.003	0.003	0.001	0.001	0.001	0	0	0	0.001	0				
XAlm	0.459	0.399	0.547	0.430	0.438	0.434	0.575	0.547	0.581	0.557	0.589	0.551	0.563				
XSpess	0.514	0.574	0.431	0.561	0.534	0.501	0.396	0.433	0.389	0.421	0.384	0.426	0.413				
XGros	0.018	0.022	0.014	0.002	0.021	0.062	0.020	0.015	0.020	0.014	0.016	0.016	0.016				
XPyr	0.007	0.004	0.005	0.005	0.006	0.001	0.007	0.004	0.008	0.006	0.008	0.005	0.006				

Sample	PK-43a		PK-43a		PK-43a		PK-43a		PK-43a		PK-43a		PK-1a		PK-1a		PK-1a		PK-1a		PK-1a		PK-1a			
Anal. #	8	9	10	11	12	13	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18		
Descript.	r	r	c	c	r	c	c FI	c FI	sm	r	m	c FI	c FI	r	r	m	c FI	c FI	c FI	c FI	c FI	c FI	c FI	c FI		
FeO	25.8	24.49	23.64	22.21	25.63	25.71	15.72	16.35	25.24	26.51	25.46	25.54	24.72	16.3	16.73	18.57	19.57	15.87	16.52	25.71	24.94	15.97	14.69	15.95	16.21	16.8
MnO	0.01	0	0.04	0.14	0.02	0.05	0.07	0.04	0.02	0.01	0.01	0.14	0.08	0.02	0	0.03	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
TiO2	0.02	0	0.03	0.01	0.01	0.01	0.01	0.02	0.03	0.03	0.04	0.02	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	
Na2O	37.63	37.54	37.73	37.55	37.37	37.66	36.1	36.28	37.36	37.48	36.99	37.11	36.72	20.4	20.61	20.45	20.49	20.66	20.25	20.44	20.58	20.48	20.36	20.5	20.58	
SiO2	0.18	0.15	0.21	0.19	0.22	0.19	0.2	0.19	0.3	0.38	0.23	0.22	0.29	0.53	0.55	0.61	0.8	0.55	0.72	0.65	0.62	0.88	0.69	0.92	0.57	0.69
Al2O3	0.02	0	0.01	0.03	0.01	0	0.04	0.07	0	0.01	0	0.03	0.05	0.02	0	0.01	0.03	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
MgO	100.8	100.0	101.2	100.9	100.3	101.1	98.93	99.09	100.2	100.2	99.96	100.3	99.92	1.754	1.671	1.601	1.507	1.748	1.745	1.091	1.133	1.723	1.808	1.748	1.747	1.698
CaO	1.121	1.156	1.273	1.345	1.095	1.136	1.809	1.750	1.104	1.013	1.108	1.122	1.168	0.021	0.017	0.025	0.021	0.027	0.023	0.023	0.023	0.035	0.045	0.027	0.025	0.035
SnO2	0.021	0.017	0.025	0.021	0.027	0.023	0.023	0.023	0.035	0.045	0.027	0.025	0.035	0.045	0.047	0.053	0.069	0.047	0.061	0.057	0.055	0.075	0.059	0.079	0.049	0.059
Total	0.003	0	0.003	0.001	0.001	0.001	0.001	0.001	0.003	0.005	0.003	0.005	0.001	0.003	0	0.003	0.005	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001
Fe	1.968	1.966	1.976	1.990	1.967	1.959	1.915	1.954	1.950	1.948	1.978	1.977	1.976	3.060	3.065	3.054	3.048	3.047	3.059	3.000	3.005	3.048	3.057	3.036	3.035	3.016
Mn	0	0	0.001	0.007	0.001	0.001	0.003	0.001	0.001	0	0	0.007	0.005	0	0	0.001	0.007	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001
Ti	0.595	0.577	0.541	0.511	0.598	0.588	0.365	0.382	0.586	0.617	0.589	0.593	0.573	0.381	0.399	0.431	0.456	0.375	0.382	0.606	0.590	0.375	0.346	0.373	0.381	0.394
XAlm	0.015	0.016	0.018	0.023	0.016	0.020	0.019	0.018	0.025	0.020	0.026	0.016	0.020	0.007	0.006	0.008	0.007	0.009	0.008	0.008	0.008	0.012	0.015	0.009	0.008	0.012
XSpess	0.007	0.006	0.008	0.007	0.009	0.008	0.008	0.008	0.008	0.012	0.015	0.009	0.008	0.012	0.007	0.006	0.008	0.007	0.009	0.008	0.008	0.012	0.015	0.009	0.008	0.012
XGros																										
XPyr																										

Sample	PK-1a	PK-1a	PK-58	PK-58	PK-58	PK-58	PK-43b	PK-43b	PK-43b	PK-43b	PK-38d	PK-38d	PK-38d
Anal. #	8	9	1	2	3	4	1	2	3	4	1	2	3
Descript.	c	FI	m				c	r	sg	lr			
FeO	25.54	25.99	25.42	25.45	24.8	25.07	23.54	25.35	26.39	26.66	30.69	31.35	32.07
MnO	16.44	15.93	15.74	15.67	17.14	16.77	18.43	16.5	15.45	15.56	11.84	10.58	10.3
TiO2	0.08	0.08	0.03	0.05	0.03	0.11	0.09	0.04	0.01	0.03	0.02	0.03	0.05
Na2O	0.03	0.01	0.02	0.06	0.04	0.08	0.04	0	0.04	0.02	0	0.02	0.02
SiO2	37.17	37.27	37.49	37.58	37.68	37.61	36.93	37.08	37.29	37.32	37.64	37.43	37.79
Al2O3	20.5	20.38	20.43	20.35	20.45	20.41	20.55	20.53	20.61	20.51	20.61	20.9	20.63
MgO	0.25	0.26	0.39	0.36	0.37	0.39	0.28	0.25	0.31	0.31	0.33	0.33	0.42
CaO	0.66	0.71	0.55	0.9	0.77	0.76	0.7	0.66	0.74	0.51	0.63	0.57	0.59
SnO2	0.01	0.03	0.02	0	0.01	0.01	0.01	0.02	0	0	0.01	0	0
Total	100.6	100.6	100.0	100.4	101.2	101.2	100.5	100.4	100.8	100.9	101.7	101.2	101.8
Fe	1.742	1.773	1.735	1.732	1.680	1.698	1.609	1.733	1.796	1.814	2.069	2.122	2.161
Mn	1.134	1.100	1.089	1.080	1.174	1.150	1.275	1.142	1.064	1.072	0.809	0.724	0.703
Mg	0.029	0.031	0.047	0.043	0.043	0.045	0.033	0.029	0.037	0.037	0.039	0.039	0.049
Ca	0.057	0.061	0.047	0.077	0.065	0.065	0.061	0.057	0.063	0.043	0.053	0.049	0.049
Na	0.003	0.001	0.001	0.009	0.005	0.011	0.005	0	0.005	0.003	0	0.001	0.001
Al	1.968	1.963	1.993	1.958	1.981	1.961	1.993	1.971	1.978	1.990	1.992	1.995	1.959
Si	3.032	3.039	3.060	3.060	3.050	3.046	3.018	3.032	3.034	3.036	3.038	3.031	3.043
Ti	0.003	0.003	0.001	0.001	0.001	0.001	0.005	0.001	0	0.001	0	0.001	0.001
XAlm	0.587	0.597	0.594	0.590	0.566	0.573	0.539	0.584	0.606	0.611	0.696	0.722	0.729
XSpess	0.382	0.370	0.372	0.368	0.396	0.388	0.427	0.385	0.359	0.361	0.272	0.246	0.237
XGros	0.019	0.020	0.016	0.026	0.022	0.022	0.020	0.019	0.021	0.014	0.018	0.016	0.016
XPyr	0.010	0.010	0.016	0.014	0.014	0.015	0.011	0.010	0.012	0.012	0.013	0.013	0.016

Sample	PK-38d	PK-38d	PK-38d	PK-38d	PK-25	PK-25	PK-25
Anal. #	4	5	6	7	1	2	3
Descript.	c			c	r	c	
	sg 3				sg 1		
FeO	31.33	25.1	26.31	30.73	28.15	28.02	27.65
MnO	10.74	17.21	15.72	11.06	13.12	13.51	13.88
TiO2	0.01	0.07	0.04	0.07	0.07	0.03	0.02
Na2O	0.01	0.01	0	0.04	0.03	0.01	0
SiO2	37.24	37.44	37.4	37.54	36.93	37.42	37.27
Al2O3	20.67	20.55	20.89	20.64	20.38	20.72	20.65
MgO	0.33	0.25	0.23	0.36	0.32	0.29	0.24
CaO	0.65	0.79	0.79	0.77	0.57	0.56	0.46
SnO2	0	0.02	0.03	0.01	0.03	0	0
Total	100.9	101.4	101.4	101.2	99.6	100.5	100.1
Fe	2.131	1.700	1.779	2.081	1.939	1.906	1.889
Mn	0.739	1.181	1.076	0.758	0.914	0.932	0.961
Mg	0.039	0.029	0.027	0.043	0.037	0.033	0.029
Ca	0.055	0.069	0.067	0.067	0.049	0.049	0.039
Na	0.001	0.001	0	0.005	0.005	0.001	0
Al	1.982	1.962	1.991	1.971	1.977	1.988	1.989
Si	3.028	3.033	3.025	3.039	3.039	3.045	3.045
Ti	0	0.003	0.001	0.003	0.003	0.001	0.001
XAlm	0.718	0.570	0.602	0.705	0.659	0.652	0.646
XSpess	0.249	0.396	0.364	0.257	0.310	0.318	0.329
XGros	0.018	0.023	0.023	0.023	0.016	0.017	0.013
XPyr	0.013	0.010	0.009	0.014	0.012	0.011	0.010

APPENDIX IIe
**ELECTRON MICROPROBE ANALYSES OF OXIDES FROM NONG
SUA**

Abbreviations: cas cassiterite, wolf wolframite, col columbite, ix ixiolite, ru rutile, ilm ilmenite, Ksp K-feldspar, tour tourmaline, gar garnet incl inclusion in, diss disseminated.

Rutile, stoichiometry based on 2 oxy

Sample #	PK-3c	PK-3c	PK-36	PK-36	PK-36	PK-36	PK-45	PK-45	PK-45	PK-52
Anal. #	5	12	1	2	11	14	4	5	7	1
Descr	incl Ksp	incl Ksp	incl cas	incl cas	incl cas	incl cas	incl cas	incl cas	incl cas	diss
FeO	7.33	9.64	1.25	1.03	1.67	1.46	5.4	5.97	5.77	1.6
MnO	0.09	0.08	0.01	0	0.03	0.02	0.02	0.03	0.04	0.02
WO3	2.68	3.62	0.43	0.25	2.88	2.59	0.63	0.7	0.96	0
Ta2O5	20.34	23.11	1.22	0.86	0.16	0.02	3.98	4.38	5.57	0.55
SnO2	1.65	3.44	2.6	2.51	1.11	0.77	7.68	1.42	1.33	0.07
TiO2	58.58	45.36	91.39	92.89	91.67	93.04	71.68	73.46	72.26	96.36
Nb2O3	9.61	14.88	2.69	2.55	2.04	0.64	11.96	15.51	13.23	2.58
CaO	0.01	0.02	0.02	0.01	0.07	0.09	0	0	0.02	0.08
Tot.	100.28	100.14	99.6	100.12	99.88	98.67	101.1	101.52	99.25	101.26
Fe	0.100	0.141	0.014	0.012	0.019	0.017	0.067	0.073	0.072	0.018
Mn	0.001	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000
W	0.011	0.016	0.002	0.001	0.010	0.009	0.002	0.003	0.004	0.000
Ta	0.090	0.110	0.005	0.003	0.001	0.000	0.016	0.017	0.023	0.002
Sn	0.011	0.024	0.014	0.014	0.006	0.004	0.045	0.008	0.008	0.000
Ti	0.720	0.596	0.949	0.955	0.951	0.968	0.798	0.802	0.810	0.968
Nb	0.071	0.118	0.017	0.016	0.013	0.004	0.080	0.102	0.089	0.016
Ca	0.000	0.000	0.000	0.000	0.001	0.001	0.000	0.000	0.000	0.001
Cations	1.005	1.006	1.001	1.001	1.001	1.002	1.008	1.005	1.006	1.004

Rutile

Sample #	PK-52	PK-52	PK-52
Anal. #	5	9	11
Descr	incl cas	incl cas	incl cas
FeO	0.61	1.49	2.59
MnO	0	0	0.01
WO3	1.22	1.91	0.33
Ta2O5	0.97	0.11	1.47
SnO2	0.1	1.75	3.42
TiO2	94.46	94.67	86.32
Nb2O3	1.7	1.31	7.47
CaO	0.02	0	0.01
Tot.	99.08	101.25	101.62
Fe	0.007	0.017	0.030
Mn	0.000	0.000	0.000
W	0.004	0.007	0.001
Ta	0.004	0.000	0.006
Sn	0.001	0.009	0.019
Ti	0.972	0.962	0.899
Nb	0.011	0.008	0.047
Ca	0.000	0.000	0.000
Cations	0.998	1.003	1.001

Ilmenite, stoichiometry based on 3 oxy

Sample #	PK-36	PK-36	PK-36	PK-52
Anal. #	7	8	13	12
Descr	incl cas	incl cas	incl cas	incl cas
FeO	35.63	40.58	34.43	41.2
MnO	11.18	6.01	11.14	5.17
WO3	0.02	0	0.03	0.02
Ta2O5	0.12	0.03	0.09	0.03
SnO2	0.13	0.46	0.34	0.23
TiO2	53.86	52.03	51.82	54.2
Nb2O3	0.16	0.23	0.23	0.26
CaO	0	0.01	0.01	0.02
Tot.	101.09	99.36	98	101.15
Fe	0.741	0.863	0.740	0.855
Mn	0.235	0.129	0.242	0.109
W	0.000	0.000	0.000	0.000
Ta	0.001	0.000	0.001	0.000
Sn	0.001	0.005	0.003	0.002
Ti	1.007	0.995	1.001	1.011
Nb	0.002	0.003	0.003	0.003
Ca	0.000	0.000	0.000	0.001
Cations	1.987	1.996	1.990	1.981

Ixiolite, stoich. based on 8 oxy

Sample #	PK-31a	PK-52
Anal. #	4	6
Descr	diss	ix/col. incl cas
FeO	16.11	9.57
MnO	5.9	0.07
WO3	34.33	3.29
Ta2O5	6.18	23.89
SnO2	0.47	2.65
TiO2	1.5	45.55
Nb2O3	36.11	15.76
CaO	0.02	0
Tot.	100.67	101.77
Fe	1.161	0.548
Mn	0.431	0.004
W	0.767	0.058
Ta	0.145	0.445
Sn	0.016	0.072
Ti	0.097	2.396
Nb	1.407	0.488
Ca	0.002	0.000
Cations	4.024	4.014

Columbite, stoich. based on 6 oxy

Sample #	PK-17c	PK-3c	PK-3c	PK-45	PK-50d
Anal. #	1	19	20	6	7
Descr	col/ix	incl gar	incl gar	incl cas	diss
FeO	9.38	14.14	14.24	13.89	15.32
MnO	10.8	4.12	4.6	6.36	4.89
WO3	14.74	14.16	8.66	6.42	7.24
Ta2O5	5.36	22.39	15.66	3.97	5.85
SnO2	0.22	0.76	0.71	0.78	0.17
TiO2	1.61	4.23	3.5	4.81	3.89
Nb2O3	57.91	40.29	52.36	64.24	62.17
CaO	0.02	0.03	0.03	0.03	0.02
Tot.	100.31	100.12	99.75	101.22	99.56
Fe	0.470	0.754	0.727	0.663	0.748
Mn	0.548	0.223	0.238	0.307	0.242
W	0.229	0.234	0.137	0.095	0.109
Ta	0.087	0.388	0.260	0.062	0.093
Sn	0.005	0.019	0.017	0.018	0.004
Ti	0.073	0.203	0.161	0.206	0.171
Nb	1.568	1.162	1.446	1.657	1.640
Ca	0.001	0.002	0.002	0.002	0.001
Cations	2.980	2.983	2.987	3.007	3.006

Cassiterite, stoichiometry based on 2 oxy

Sample #	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36	PK-36
Anal. #	1	2	3	4	5	4	5	6	9
Descr	cas	cas	cas	cas	cas	cas	cas	cas	cas
FeO	0.07	0.11	0	0.02	0.03	0.08	0.11	0.04	0.01
MnO	0	0.01	0	0.02	0	0.01	0.04	0	0.03
WO3	0	0	0	0	0	0	0	0	0.03
Ta2O5	0.15	0.1	0.24	0.22	0.24	0.07	0	0.16	0.1
SnO2	99.64	99.22	99.68	99.67	99.37	99.39	97.73	99.26	99.26
TiO2	0.48	0.33	0.27	0.32	0.63	0.47	0.76	0.35	0.6
Nb2O3	0	0.21	0.08	0.12	0.1	0.01	0.04	0	0
CaO	0.26	0.31	0.27	0.33	0.33	0.29	0.27	0.25	0.23
Tot.	100.64	100.32	100.64	100.72	100.7	100.35	98.99	100.13	100.13

Fe	0.001	0.002	0.000	0.000	0.001	0.002	0.002	0.001	0.000
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.001
W	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ta	0.001	0.001	0.002	0.001	0.002	0.000	0.000	0.001	0.001
Sn	0.986	0.985	0.988	0.986	0.980	0.986	0.980	0.988	0.984
Ti	0.009	0.006	0.005	0.006	0.012	0.009	0.014	0.007	0.011
Nb	0.000	0.002	0.001	0.001	0.001	0.000	0.000	0.000	0.000
Ca	0.007	0.008	0.007	0.009	0.009	0.008	0.007	0.007	0.006
Cations	0.997	0.996	0.996	0.995	0.995	0.997	0.998	0.997	0.997

Sample #	PK-45	PK-45	PK-50d	PK-51	PK-51	PK-53	PK-53
Anal. #	3	8	1	1	2	1	2
Descr	cas	cas	cas	cas	cas	cas	cas
FeO	0.03	0.03	0	0	0.02	0.05	0.06
MnO	0	0	0	0.01	0	0	0
WO3	0	0	0.03	0	0	0	0
Ta2O5	0.14	0.02	0.27	0.11	0	0.39	0.06
SnO2	99.35	99.58	97.81	99.83	100.3	99.15	99.77
TiO2	0.48	0.44	0.27	0.31	0.05	0.05	0.26
Nb2O3	0.05	0.1	0.07	0.01	0.02	0.08	0.24
CaO	0.28	0.16	0.28	0.27	0.33	0.27	0.29
Tot.	100.46	100.34	98.72	100.62	100.73	100.06	100.67

Fe	0.001	0.001	0.000	0.000	0.000	0.001	0.001
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.000
W	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ta	0.001	0.000	0.002	0.001	0.000	0.003	0.000
Sn	0.985	0.988	0.987	0.989	0.994	0.990	0.987
Ti	0.009	0.008	0.005	0.006	0.001	0.001	0.005
Nb	0.001	0.001	0.001	0.000	0.000	0.001	0.003
Ca	0.007	0.004	0.008	0.007	0.009	0.007	0.008
Cations	0.996	0.998	0.995	0.996	0.996	0.996	0.996

Wolframite					Scheelite	
Sample #	PK-31a	PK-31a	PK-31a	PK-54b	PK-45	PK-54b
Anal. #	1	2	3	1	2	2
Descr	after tour	after tour	after gar		after wolf	after wolf
FeO	16.34	16.23	8.43	15.39	0.9	0
MnO	7.85	8.13	14.53	8.56	0.26	0
WO3	75.01	75.34	76.02	76	80.04	80.58
Ta2O5	0.73	0	0.25	0.11	0	0.11
SnO2	0.1	0.16	0.09	0.05	0	0
TiO2	0.05	0.03	0.11	0	0.03	0
Nb2O3	0.81	1.38	0.52	0.59	0.61	0
CaO	0.04	0.02	0.04	0.02	19	19.91
Tot.	100.93	101.35	100.01	100.74	100.85	100.64

Fe	0.681	0.672	0.355	0.643	0.036	0.000
Mn	0.331	0.341	0.619	0.362	0.010	0.000
W	0.969	0.967	0.991	0.985	0.984	0.994
Ta	0.010	0.000	0.003	0.001	0.000	0.001
Sn	0.002	0.003	0.002	0.001	0.000	0.000
Ti	0.002	0.001	0.004	0.000	0.001	0.000
Nb	0.018	0.031	0.012	0.013	0.013	0.000
Ca	0.002	0.001	0.002	0.001	0.966	1.015
Cations	2.016	2.016	1.989	2.007	2.011	2.010

APPENDIX III
FLUID INCLUSION DATA FROM NONG SUA

CASSITERITE L-V
 SAL ice calc. from Oakes et al. (1990) assuming X_{NaCl}=1, SAL clathr calc. from Bozdo et al., > 10 °C given 0.0 salinity
 V ice molar volume using SAL ice, V clathr molar volume using SAL clathr
 ** with solid, * with CO₂

sample #	origin	TrCO ₂	TrICE	TrCLATH	TrHydr	Th L	SAL ice	SAL clathr	V ice	V clathr
35d**	p?					293				
35d*	p?		-1.8	10.8		298	3.05	0	24.9	26.04
35a*	p?			15.2		307		0		26.85
35a*	p?			8.9		309		2.22		
35d*	p?	-57.1				378				
35a*	p?		-2.2			305	3.69		25.18	
35d	p?					283				
36*	p?	-57.1		8.8		355		2.42		30.81
36*	p?	-57.2		9				2.03		
36*	ps			3.5		364		11.29		26.33
36*	ps		-17.5	2.5		361	20.58	12.63	24.3	25.67
36*,**	ps			3.1		362		11.84		26
36	ps					329				
36*	ps	-57.0				410				
36	ps					415				
36	p?		-20.5			395	22.69		25.54	
36*,**	p?	-57.1						7.31		
36*	p?	-57.2		6.1				7.64		
36*	p?	-57.0		5.9						
35b**	ps?		-4.3			348	6.87		27.06	
35b	ps?		-4.8			340	7.59		26.16	
35b	ps?					347				
35b**	p?		-4.3			345	6.87		26.85	
35b	p?		-4.8			344	7.59		26.45	
35b	p?		-4.8			345	7.59		26.52	
35b	p?					344				
35b	p?		-3.3			348	5.4		27.88	
35b	p?		-18.3			396	21.16		25.44	
35b	p?		-17.3			374	20.43		24.73	
35b	p?		-7.8			371	11.5		26.69	
35b	p?		-7.8			371	11.5		26.69	
35b	p?		-16.8			395	20.05		25.53	
21*	p?	-60.8		12.5				0.00		
21*	p?	-59.8		12.1				0.00		
21*	p?			12.5		307		0.00		26.85
49*	?	-57.5			3.0	379				
49	?				3.0					
49	?				4.6					
49*	?		-17.8			359	20.8		24.21	
49*	?	-57.7				340				
49*	?	-58.3		6.8		338		6.12		26.63
49	?					330				
49*	?	-58.1								

sample #	origin	TmCO2	TmICE	TmCLATH	TmHyar	Th L	SAL ice	SAL clath	V ice	V clath
49	?		-14.8				18.48			
49	?		-7.5				11.13			
49	?					382				
49*	?	-58.8		8.9				2.22		
49	?		-15.8			386	19.28		25.29	
49*	?		-1.2			369	2.06		33.3	
49	?		-14			387	17.81		25.59	
62	?	-56.8		7.1		355		5.59		28.43
62	?	-56.8		7.3		348		5.23		27.96
62	?	-57.0		7.1		355		5.59		28.43
62	PS or S					366				
62*	PS or S	-56.7		6.2		358		7.14		27.8
62	PS or S					374				
62	PS or S					370				
21*	PS or S			-10.3		406		20.45		
21*	PS or S			-1.7		397		17.11		
21*	PS or S		-28.4	-3.8		369	26.54	18.65		

CASSITERITE L-L-V homogenizes to H2O or critical
 * solid and CO2 homogenization visible
 cr Th tot. is critical XCH4 calc. after Buruss (1981)

sample #	origin	TmCO2	TmCLATH	ThCO2L	ThCO2V	ThCO2 Cr	XCH4	Th TOT L	SAL clathVol	% CO2	XCO2
35d*	p?	-56.8	7.1		28.9		0.01	340	5.59	56	0.13
35d	p?	-57.7	6.9		28.8		0.03	341	5.94	35	0.06
35d*	p?	-56.9		27.5			0.01				
49	?	-59.3	8.2	23.6			0.10	293	3.57		
49	?	-60.5	7.8			18.8	0.16	297	4.32		
62	?	-57.0	7.5		24.0		0.01	352	4.87	23	0.03
62 cr	P?	-56.9	6.5		29.3		0.01	352	6.63	11	0.02
62 cr	P?	-57.0	6.7		29.5		0.01	352	6.29	15	0.02
62 cr	P?	-57.0	6.7		29.2		0.01	346	6.29	27	0.05
62 cr	P?	-56.9	6.7		29.2		0.01	347	6.29	23	0.04
62	P?	-56.9	6.7		29.0		0.01	352	6.29		
62 cr	P?	-56.8	6.3				0.01	360	6.97		
62 cr	P?	-56.8	6.3		28.5		0.01	360	6.97		
62 cr	P?	-56.9	6.3		29.5		0.01	360	6.97		
62	P?	-56.8	6.9		27.0		0.01		5.94		
62*	PS or S	-56.6	-4.7		13.0		0.00	401	19.16		
62*	PS or S	-56.6	-4.9		17.5		0.00	407	19.27		
62*	PS or S	-56.7	-6.2		21.5		0.00	405	19.83	27	0.03
21	PS or S	-59.0	11.7		18.0		0.14	311	0.00		
21	PS or S	-59.1	12.1		17.0		0.14	309	0.00		
21*	PS or S	-60.0	12.6	12.8			0.10	309	0.00		
21	PS or S	-59.1	11.9	16.1			0.07	307	0.00		
21	PS or S	-59.2	12.3	13.6			0.08	309	0.00		

CASSITERITE		V, V-L, V-L-L		homogenizes to CO2					
sample #	orig.	T-CO2	T-CLATH	T-CO2L	T-CO2V	XCH4	Tm TOT V	SAL clath	Vol% CO2
35d	p?	-56.9	6.7	27.2		0.01	357	6.29	54
35b	p?	-57.8		23.4		0.04			0.24
35b	p?		10.3					0.00	
21	p?	-60.3							
21	p?	-59.3	11.3	14.8		0.1		0.00	
21	p?	-59.4	11.5	15.7		0.1		0.00	
21	p?	-60.0	11.8	9.6		0.13		0.00	
49	?	-58.0							
49	?	-57.6		27.8		0.01			
49	?	-58.2		24.6		0.04			
49	?	-58.3		25.2		0.04	427		37
49	?	-61.2	11.2						0.15
49	?	-59.2							
49	?	-58.8	8.2	24.0		0.09	295	3.57	
49	?	-58.6	6.8	24.6		0.08	297	6.12	
49	?	-58.9	8.6	22.0		0.1		2.81	
49	?	-59.7	7.8	25.0		0.1		4.32	
49	?	-59.9							
62	S or PS	-56.6		26.6		0			
62	S or PS	-56.6							
62	S or PS		2.0		-1.0		342	13.26	
62	PS or S	-56.6							
62	PS or S	-56.6	6.5	29.2		0		6.63	

TOURMALINE		L-V		* with CO2, ** with S					
sample #	origin	TmCO2	TmICE	TmCLATH	Th	SAL ice	SAL clath		
35d*	p?	-56.7		10.3	313		0.00		
35d*	p?			10.7	293		0.00		
35d*	p?	-57.5		10.3	323		0.00		
35d*	p?	-57.4		10.5			0.00		
35d*	p?	-57.4		10.1	318		0.00		
35d*	p?	-57.4		10.1			0.00		
35d*	p?	-57.4		10.5	310		0.00		
35d*	p?	-57.5		10.7	309	4.48	0.00		
35d*	p?		-2.7	9.5	321	4.17	1.03		
35d	p?		-2.5		219	4.01			
35d*	p?		-1.5	9.1	245	2.56	1.83		
35d	p?		-1.9		271	3.21			
35d*	p?		-1.8		261	3.05			
35d**	p?		-2.5		232	4.17			
35d**	p?	-56.8	-15.5	5.1	305	19.04	8.93		
35d	p?		-2.0		319	3.37			
35d*	p?		-2.5		247	4.17			
36	p?				328				

SAMPLE #	origin	TmCO2	TmICE	TmCLATH	Th	SAL ice	SAL clath
36*	p?			10.2	328		0.00
36*	p?	-58.1	-10.5	6.3	241	14.52	6.0
36*	p?	-58.0			331		
36*	p?			11.1	328		0.00
49*	S or PS		-4.2		293	6.73	
49	S or PS				316		
49	S or PS				314		
49	S or PS		-3.4		301	5.55	
49*	?		-4.5	7.8	297	7.16	4.32
49*	S or PS		-4.8		307	7.59	
49*	S or PS		-4.8		208	7.59	
49	S or PS		-4.5		320	7.16	
49*	?		-4.3		272	6.87	
49*	?		-4.3		274	6.87	
49*	?		-4.2		326	6.73	
49*,**	P?		-4.9			7.73	
49*,**	P?		-3.8			6.14	
49*,**	P?	-60.0	-6.2	8.3		9.49	3.38
49*,**	P?		-5.8			8.96	
49*,**	P?	-62.2	-7.1	8.1		10.64	3.76
49*	P?		-27.1			26.29	
49*	P?	-62.2		8.4			3.19
49	S		-2.2		293	3.69	
49	S		-2.6		235	4.32	
49	S		-1.9		299	3.21	
49	S		-2.5		229	4.17	
49	S		-2.2		233	3.69	
49	?		-2.4		324	4.01	
49*	?		-3.8	9.0	240	6.15	
49**	?		-4.2		240	6.73	
49	?	-56.8	-4.2			6.73	
2c*	P?		-2.4	9.8	309	4.01	
2c	P?		-2.3		288	3.85	

TOURMALINE L-L-V nomogenizes to H2O

sample #	origin	TmCO2	TmICE	TmCLATH	ThCO2L	ThCO2V	XCH4	Th TOT L	Td	SAL ice	SAL clath	Vol% CO2	XCO2
49**	P?	-61.8			26.9		0.24						
49	?	-57.0	-12.0	6.7		29.6	0.01			16.00	6.29		
49	?	-57.1		6.7		29.6	0.01	357			6.29	27	0.05
49	?	-57.0		6.5	28.9		0.01	273			6.63	19	0.06
49*	?	-57.0		6.9	28.5		0.01	312			5.94	48	0.19
49	?	-56.8		6.5	18.6		0.01		200		6.63		
49	?	-57.4	-15.4	3.3		25	0.01		220	18.96	11.57		
49	?	-57.4		8.3		27.4	0.01	296			3.38	33	0.05
49	?	-58.2		8.9		26	0.08	281			2.22		

TOURMALINE V, V-L, V-L-L homogenizes to CO2

sample #	origin	T-CO2	T-CLATH	T-CO2L	XCH4	Tn	TOT V	SAL clatr	Vol. % CO2	XCO2
35d	p?	-56.9	8.2	29.5	0.01	322	3.57	59	0.27	
35d	p?	-56.8	8.7	29.7	0.01		2.62			

TOURMALINE MULTISOLIDS

sample #	origin	# solids	T-CO2 L	T-CO2	T-ICE	Tn	SAL ice
35d	p?	3	4.9	-56.9	-10.0		13.99
35d	p?	6			-3.2	127	5.25

QUARTZ L-V

* with CO2
** with S

sample #	origin	TmCO2	T-ICE	TmHYDR	TmCLATH	Tn	Tn S	Td	SAL ice	SAL clatr
35d**	s?	-56.9	-10.5	7.9				305	14.52	
35d*	s?	-56.8			2.1	353				13.14
35d*	s?	-56.7	-13.5		4.1	315			17.37	10.44
35d*,**	s?	-56.6			0.3					15.21
35d*	s?	-56.6	-12.5		3.9	313			16.47	10.72
35d*,**	s?	-56.7	-10.5		3.1	315			14.52	11.84
35d*	s?	-56.7			4.9	368				9.24
35d*	s?	-56.9			4.8	372				9.39
35d	s?		-4.2			224			6.73	
35d	s?		-5.8						8.96	
35d	s?					235				
35d*	s?	-56.8			4.3	376				10.14
21*	s?		-23.5			252		251	24.61	
21**	s?			20.0			213	342		
2c*	S	-57.0			7.9	344				4.14
2c*	S	-57.0			6.2					7.14
2c	S		-2.8						4.63	
2c*	S	-57.0			7.9	327				4.14
2c*,**	S	-56.7			3.1					11.84
2c*	S	-56.9			-2.4	415				17.67
2c*	S	-56.7			-2.8	418				17.97
2c*,**	S	-56.7	-20.3			378			22.56	
21*	S	-63.1			12.9	316				0.00
21*	S	-57.4			9.9	314				0.21
21*	S	-57.0			5.5	268				8.29
21*,**	S				9.8	310				0.41
21*,**	S				11.6	248				0.00

QUARTZ L-L-V homogenizes to H2O
CO2 homogenization visible
* with S

sample #	origin	TmCO2	TmICE	TmCLATH	ThCO2L	ThCO2V	ThCO2 Cr XCH4	Th TOT L	Td	SAL ice	SAL clathVol%	CO2	XC02
35d	?	-56.9		4.3		29.6	0.01				10.14		
35d	?	-56.9					0.01						
35d	?	-56.8			29.8		0.01						
35d*	?	-56.8	-10.0	-1.3			0.01		322	13.99	16.76		
35d*	?	-56.9		4.3	29.5		0.01				10.14		
35b	s?	-57.6		8.3	27.0		0.04		302		3.38		
35b	s?	-57.7		8.0	27.0		0.04				3.95		
35b	s?	-57.1		6.9	26.3		0.01		252		5.94	36	0.15
35b	s?	-57.5		8.3	26.3		0.01	318			3.38		
35b	s?	-57.2		8.0	26.8		0.01		231		3.95		
35b	s?	-57.6		8.5	24.8		0.01		282		3.00		
35b*	s?	-57.5		8.5	25.5		0.01		302		3.00		
35b	s?	-57.6		9.1	25.3		0.01	308			1.83	36	0.14
35b	s?	-57.3		8.9	25.9		0.01	308			2.22	30	0.11
21	?	-57.2		9.3	26.8		0.01	314			1.43	49	0.21
21	s?	-57.2		9.4	27.2		0.01	302			1.23	38	0.14
49*	?	-56.8			28.7		0.01						
49*	?	-56.9			28.5		0.01						
2c*	S	-57.0		-3.6		30.0	0.01				18.52		
2c*	S	-56.9		-0.7		29.2	0.01				16.21		
2c*	S	-56.8		0.2		28.8	0.01				15.31		
21	S	-57.4		9.9	26.2		0.01		240		0.21		
21	S	-57.4		10.1	27.2		0.01				0.00		

QUARTZ V,V-L,V-L-L
Homogenizes to CO2 phase
* with S

sample #	origin	TmCO2	TmCLATH	ThCO2L	ThCO2 Cr XCH4	Th TOT V	Td	SAL clathVol%	CO2	XC02
35d	s	-56.9			30.1	0				
35d	?	-56.8	8.2	28.2	0.01		292	3.57		
35d	?	-56.8	4.5	29.4	0.01	404		9.84	48	0.2
35b	s	-57.5								
21	?	-57.2	9.3	26.8	0.01	314		1.43	53	0.23
49	?	-56.8		28.4	0.01	335				
49	?	-56.8		28.7	0.01	345				
49	?	-56.8				395				
49*	?	-56.9		28.6	0.01					
2c	S	-56.7	3.2	29.3	0.01			11.70		
2c	S	-56.8	3.9	28.3	0.01			10.72		

QUARTZ

HALITE-BEARING

of solids includes halite
w/CO2 with CO2

Wt% calc from Potter et al. (1977)

sample #	origin	# solids	Tm ICE	Tm HYDR	Tm L	Tm NaCl	Tm Si	SAL
35b	s	2	-34.4		149	256		34.98
35b	s	2	-31.8	15.0	167	252		34.73
35b	s	2	-32.0		156	246	36b	34.40
35b	s	2	-28.5	13.0	143	224		33.17
35a	?	2	-34.8		200	244		34.31
49	?	2	-26.7		186	228		33.37
2c	s	3	-25.5	23.0				
2c	s	2				282		36.68
2c w/CO2	s	3	-24.3	9.0		256		35.01
2c	s	3	-24.3			246		34.40

GARNET

L-V

* with CO2

** with S

sample #	origin	Tm ICE	Tm	SAL ice
3c	P		302	
3c	P		307	
3c	P		302	
3c	P		304	
3c	P	-1.6	303	2.72
3c	P	-1.7	310	2.88
3c	P	-1.5	315	2.56
3c	P	-1.3	315	2.23
3c	P		317	
3c	P	-1.4	317	2.39
3c	P		317	
3c	P		317	
3c	P		295	
3c	P		295	
3c	P	-1.2	306	2.06
3c	P		295	
3c	P	-1.2	304	2.06

APPENDIX IV
ELECTRON MICROPROBE DATA FROM PILOK

FELDSPARS FROM PILOK

Sample #	1-6c1	1-6c1	1-6c1	1-6c1	1-6c1	5-25	5-25	5-25	5-25	5-25	5-25	5-25
Anal. #	1	2	3	4	5	1	2	3	4	5	6	7
FeO	0.04	0	0	0.02	0.01	0	0	0	0.01	0.01	0	0.04
Na2O	0.46	0.2	0.18	1.21	0.72	11.19	0.47	11.37	11.47	0.53	11.1	0.39
SiO2	64.99	63.18	65.15	64.79	64.23	68.19	64.25	68.19	67.98	65.18	67.62	64.5
Al2O3	18.15	20.04	18.28	18.32	18.27	19.75	18.41	19.77	19.68	18.25	19.73	18.1
MgO	0	0	0	0.01	0	0.01	0.01	0.01	0	0.01	0	0.02
K2O	16.86	16.94	17.49	15.86	16.48	0.12	16.8	0.06	0.07	16.68	0.09	16.95
CaO	0	0	0	0	0	0.44	0.01	0.44	0.42	0	0.42	0
BaO	0.01	0	0.06	0.05	0.11	0	0	0	0.04	0	0.07	0.13
Tot	100.5	100.3	101.1	100.2	99.83	99.69	99.95	99.84	99.67	100.6	99.01	100.1
Fe	0.001	0	0	0.000	0.000	0	0	0	0.000	0.000	0	0.001
Na	0.041	0.017	0.016	0.108	0.064	0.949	0.042	0.964	0.975	0.047	0.949	0.035
Si	3.000	2.925	2.995	2.990	2.987	2.985	2.984	2.982	2.981	3.000	2.981	2.994
Al	0.987	1.093	0.990	0.996	1.001	1.019	1.007	1.019	1.017	0.990	1.025	0.990
Mg	0	0	0	0.000	0	0.000	0.000	0.000	0	0.000	0	0.001
K	0.992	1.000	1.025	0.933	0.977	0.006	0.995	0.003	0.003	0.979	0.005	1.003
Ca	0	0	0	0	0	0.020	0.000	0.020	0.019	0	0.019	0
Ba	0.000	0	0.001	0.000	0.002	0	0	0	0.000	0	0.001	0.002
X ab	0.039	0.017	0.015	0.103	0.062	0.972	0.040	0.975	0.976	0.046	0.974	0.033
X an	0	0	0	0	0	0.021	0.000	0.020	0.019	0	0.020	0
X or	0.960	0.982	0.984	0.896	0.937	0.006	0.958	0.003	0.003	0.953	0.005	0.966
Sample #	5-16	5-16	5-16	5-16	5-16	2-7g	2-7g	2-7g	1-4c	1-4c	1-4c	1-4c
Anal. #	1	2	3	4	5	1	2	3	1	2	3	4
FeO	0	0	0.02	0	0	0.03	0	0.04	0.01	0.02	0	0.03
Na2O	11.58	11.5	11.45	0.48	11.49	11.5	0.25	11.36	0.72	0.93	8.66	10.79
SiO2	67.7	67.87	67.57	63.93	67.86	67.83	64.87	68.36	64.62	64.24	66.41	66.29
Al2O3	19.71	19.76	19.86	18.11	19.46	19.63	18.33	19.32	18.27	18.26	19.18	20.82
MgO	0	0.01	0.01	0	0.02	0	0	0	0	0.01	0	0
K2O	0.09	0.15	0.12	16.96	0.07	0.1	17.08	0.13	16.4	15.92	3.88	0.31
CaO	0.4	0.35	0.51	0	0.29	0.12	0	0.15	0	0	0.03	0.03
BaO	0	0	0	0.02	0	0	0.08	0	0	0.13	0.06	0.03
Tot	99.49	99.64	99.54	99.5	99.18	99.21	100.6	99.35	100.0	99.49	98.22	98.29
Fe	0	0	0.000	0	0	0.001	0	0.001	0.000	0.000	0	0.001
Na	0.987	0.978	0.975	0.043	0.931	0.981	0.022	0.967	0.064	0.083	0.756	0.929
Si	2.976	2.978	2.970	2.987	2.988	2.985	2.994	3.002	2.993	2.989	2.990	2.944
Al	1.021	1.021	1.028	0.997	1.010	1.018	0.997	0.999	0.997	1.001	1.017	1.089
Mg	0	0.000	0.000	0	0.001	0	0	0	0	0.000	0	0
K	0.005	0.008	0.006	1.011	0.003	0.005	1.005	0.007	0.969	0.945	0.222	0.017
Ca	0.018	0.016	0.024	0	0.013	0.005	0	0.007	0	0	0.001	0.001
Ba	0	0	0	0.000	0	0	0.001	0	0	0.002	0.001	0.000
X ab	0.976	0.975	0.969	0.041	0.982	0.988	0.021	0.985	0.062	0.081	0.771	0.919
X an	0.018	0.016	0.023	0	0.013	0.005	0	0.007	0	0	0.001	0.001
X or	0.004	0.008	0.006	0.958	0.003	0.005	0.978	0.007	0.937	0.918	0.227	0.018
Sample #	1-4c	1-4c	1-4c									
Anal. #	5	6	7									
FeO	0.02	0	0.04									
Na2O	11.39	0.34	0.27									
SiO2	67.77	63.14	63.64									
Al2O3	19.45	17.98	18.27									
MgO	0.01	0.02	0.01									
K2O	0.09	16.98	17.26									
CaO	0.09	0	0									
BaO	0.01	0	0.01									
Tot	98.84	98.46	99.51									
Fe	0.000	0	0.001									
Na	0.975	0.031	0.024									
Si	2.992	2.984	2.979									
Al	1.012	1.001	1.007									
Mg	0.000	0.001	0.000									
K	0.005	1.023	1.030									
Ca	0.004	0	0									
Ba	0.000	0	0.000									
X ab	0.990	0.029	0.023									
X an	0.004	0	0									
X or	0.005	0.970	0.976									

MUSOVITE FROM PILOK

samples DP-1b, DP-9b, DP-11 are from barren pegmatite, PP-1-4c and PP-2-4a from wolframite-bearing pegmatite, PP-2-5a and PP-5-31e from greisen

sample	DP-1b	DP-1b	DP-1b	DP-1b	DP-1b	DP-1b	PP-2-5a	PP-2-5a	PP-2-5a	PP-2-5a
anal.	1.00	2.00	3.00	4.00	5.00	6.00	1.00	2.00	3.00	4.00
FeO	2.47	2.68	2.60	3.20	2.59	2.67	5.43	5.58	5.36	3.91
MnO	0.29	0.26	0.39	0.54	0.31	0.34	0.39	0.46	0.22	0.42
Na2O	0.32	0.24	0.63	0.27	0.28	0.29	0.16	0.14	0.08	0.27
Al2O3	33.17	33.33	33.54	31.59	33.61	33.34	30.46	30.55	28.82	32.42
SiO2	46.92	46.95	45.85	47.17	46.64	46.57	46.92	47.20	47.68	45.59
F	1.11	0.95	1.13	1.35	1.06	1.26	0.70	0.95	0.74	1.39
MgO	0.26	0.12	0.15	0.35	0.02	0.12	0.13	0.15	1.13	0.24
K2O	11.47	11.44	10.88	11.20	11.54	11.29	11.41	11.14	11.52	10.93
CaO	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.02
TiO2	0.01	0.01	0.18	0.20	0.02	0.04	0.01	0.08	0.02	0.11
SnO2	0.06	0.01	0.00	0.09	0.12	0.04	0.06	0.01	0.10	0.12
-O=F	0.47	0.40	0.48	0.57	0.45	0.53	0.29	0.40	0.31	0.59
Tot	95.62	95.59	94.87	95.39	95.74	95.43	95.37	95.84	95.36	94.82
Fe	0.28	0.31	0.30	0.37	0.30	0.31	0.63	0.64	0.62	0.46
Mn	0.03	0.03	0.05	0.06	0.04	0.04	0.05	0.05	0.03	0.05
Mg	0.05	0.02	0.03	0.07	0.00	0.02	0.03	0.03	0.23	0.05
Al IV	1.59	1.60	1.69	1.49	1.64	1.61	1.51	1.49	1.40	1.64
Al VI	3.75	3.75	3.76	3.64	3.77	3.78	3.45	3.47	3.29	3.68
Si	6.41	6.40	6.31	6.51	6.36	6.39	6.49	6.51	6.60	6.36
Ti	0.00	0.00	0.02	0.02	0.00	0.00	0.00	0.01	0.00	0.01
Sn	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.01
Na	0.08	0.06	0.17	0.07	0.07	0.08	0.04	0.04	0.02	0.07
K	2.00	1.99	1.91	1.97	2.01	1.98	2.01	1.96	2.03	1.94
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	0.48	0.41	0.49	0.59	0.46	0.54	0.30	0.41	0.32	0.61
OH	3.52	3.59	3.51	3.41	3.54	3.46	3.70	3.59	3.68	3.39
sample	DP-9b	DP-9b	DP-9b	DP-9b	DP-9b	DP-11	DP-11	DP-11	DP-11	DP-11
anal.	1.00	2.00	3.00	4.00	5.00	1.00	2.00	3.00	4.00	5.00
FeO	2.64	2.85	3.35	3.04	2.48	2.09	1.99	1.84	3.06	2.17
MnO	0.37	0.33	0.43	0.34	0.36	0.14	0.14	0.03	0.21	0.08
Na2O	0.32	0.61	0.41	0.34	0.20	0.72	0.65	0.62	0.75	0.82
Al2O3	33.74	34.21	33.02	33.32	33.52	34.39	34.98	35.08	33.09	35.02
SiO2	46.82	46.53	46.40	45.79	45.98	46.58	46.42	45.07	46.77	45.80
F	1.26	1.26	1.32	1.33	1.04	0.57	0.35	0.42	0.68	0.46
MgO	0.14	0.08	0.13	0.22	0.15	0.42	0.22	0.31	0.37	0.23
K2O	11.44	10.93	11.17	11.38	11.37	10.53	10.91	11.02	10.40	10.59
CaO	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.02
TiO2	0.18	0.14	0.16	0.21	0.19	0.30	0.21	0.22	0.32	0.06
SnO2	0.07	0.05	0.03	0.05	0.03	0.05	0.02	0.00	0.00	0.05
-O=F	0.53	0.53	0.55	0.56	0.44	0.24	0.15	0.18	0.29	0.20
Tot	96.45	96.46	95.85	95.48	94.90	95.54	95.74	94.42	95.37	95.10
Fe	0.30	0.32	0.38	0.35	0.29	0.23	0.22	0.21	0.35	0.24
Mn	0.04	0.04	0.05	0.04	0.04	0.02	0.02	0.00	0.02	0.01
Mg	0.03	0.02	0.03	0.05	0.03	0.08	0.04	0.06	0.07	0.05
Al IV	1.64	1.69	1.63	1.68	1.68	1.73	1.78	1.86	1.65	1.81
Al VI	3.76	3.78	3.71	3.73	3.76	3.72	3.75	3.77	3.64	3.77
Si	6.36	6.31	6.37	6.32	6.32	6.27	6.22	6.14	6.35	6.19
Ti	0.02	0.01	0.02	0.02	0.02	0.03	0.02	0.02	0.03	0.01
Sn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.09	0.16	0.11	0.09	0.05	0.19	0.17	0.16	0.20	0.22
K	1.98	1.89	1.96	2.00	2.00	1.81	1.86	1.91	1.80	1.83
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	0.54	0.54	0.57	0.58	0.45	0.24	0.15	0.18	0.29	0.20
OH	3.46	3.46	3.43	3.42	3.55	3.76	3.85	3.82	3.71	3.80

sample	PP-1-4c	PP-1-4c	PP-1-4c	PP-1-4c	PP-1-4c	PP-2-4a	PP-2-4a	PP-2-4a	PP-2-4a	PP-2-4a
anal.	1.00	2.00	3.00	4.00	5.00	1.00	2.00	3.00	4.00	5.00
FeO	7.07	3.75	4.09	3.28	4.19	8.98	7.22	5.30	6.84	6.50
MnO	0.73	0.25	0.47	0.25	0.39	0.84	0.53	0.35	0.50	0.45
Na2O	0.30	0.26	0.24	0.29	0.30	0.19	0.20	0.37	0.22	0.29
Al2O3	28.79	32.99	31.79	33.50	32.18	26.96	28.85	29.63	28.57	28.21
SiO2	46.33	46.03	46.49	46.78	46.28	45.27	44.39	46.85	43.48	46.19
F	2.64	1.52	1.68	1.51	1.60	2.77	2.21	2.18	2.30	2.31
MgO	0.77	0.20	0.40	0.14	0.27	0.92	0.96	0.90	0.83	0.88
K2O	11.06	11.32	11.43	11.40	11.23	11.05	11.05	11.15	11.04	11.33
CaO	0.01	0.00	0.01	0.00	0.00	0.03	0.01	0.01	0.00	0.00
TiO2	0.38	0.13	0.21	0.04	0.02	0.37	0.43	0.33	0.40	0.35
SnO2	0.09	0.09	0.09	0.06	0.05	0.08	0.09	0.03	0.03	0.07
-O=F	1.11	0.64	0.71	0.63	0.67	1.17	0.93	0.92	0.97	0.97
Tot	97.06	95.90	96.21	96.62	95.83	96.30	95.00	96.18	93.23	95.63
Fe	0.84	0.43	0.47	0.37	0.49	1.09	0.87	0.62	0.84	0.78
Mn	0.09	0.03	0.06	0.03	0.05	0.10	0.06	0.04	0.06	0.05
Mg	0.16	0.04	0.08	0.03	0.06	0.20	0.21	0.19	0.18	0.19
Al IV	1.43	1.65	1.56	1.62	1.58	1.42	1.58	1.42	1.58	1.39
Al VI	3.39	3.72	3.63	3.77	3.67	3.20	3.33	3.48	3.39	3.36
Si	6.57	6.35	6.44	6.38	6.42	6.58	6.42	6.58	6.42	6.61
Ti	0.04	0.01	0.02	0.00	0.00	0.04	0.05	0.04	0.04	0.04
Sn	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.08	0.07	0.06	0.08	0.08	0.05	0.06	0.10	0.06	0.08
K	2.00	1.99	2.02	1.98	1.99	2.05	2.04	2.00	2.08	2.07
Ca	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
F	1.18	0.66	0.74	0.65	0.70	1.27	1.01	0.97	1.07	1.04
OH	2.82	3.34	3.26	3.35	3.30	2.73	2.99	3.03	2.93	2.96

sample	PP-2-4a	PP-5-31e	PP-5-31e	PP-5-31e	PP-5-31e
anal.	6.00	1.00	2.00	3.00	4.00
FeO	6.99	2.91	3.48	4.74	3.65
MnO	0.47	0.26	0.38	0.62	0.49
Na2O	0.20	0.52	0.25	0.26	0.27
Al2O3	28.00	34.00	33.23	29.87	31.34
SiO2	46.00	46.13	45.73	46.20	46.39
F	2.46	0.82	0.73	1.32	1.08
MgO	0.95	0.14	0.01	0.54	0.45
K2O	11.08	10.90	11.35	11.17	11.38
CaO	0.01	0.01	0.00	0.01	0.00
TiO2	0.51	0.10	0.00	0.22	0.28
SnO2	0.08	0.02	0.05	0.08	0.03
-O=F	1.04	0.35	0.31	0.56	0.45
Tot	95.71	95.46	94.91	94.47	94.89
Fe	0.84	0.33	0.40	0.56	0.42
Mn	0.06	0.03	0.04	0.07	0.06
Mg	0.20	0.03	0.00	0.11	0.09
Al IV	1.40	1.72	1.70	1.49	1.56
Al VI	3.33	3.74	3.69	3.47	3.56
Si	6.60	6.28	6.30	6.51	6.44
Ti	0.06	0.01	0.00	0.02	0.03
Sn	0.00	0.00	0.00	0.00	0.00
Na	0.05	0.14	0.07	0.07	0.07
K	2.03	1.89	1.99	2.01	2.01
Ca	0.00	0.00	0.00	0.00	0.00
F	1.12	0.35	0.32	0.59	0.47
OH	2.88	3.65	3.68	3.41	3.53

APPENDIX V
FLUID INCLUSION DATA FROM PILOK

AQUEOUS INCLUSIONS

For all inclusion hosts: SALice calculated from Bozzo et al. 1973

SALice calculated from Oakes et al. 1990, using XNaCl=0

XCH4 calculated from Meyer et al. 1982

QUARTZ

carbonic-free L and L-V inclusions

Sample	Chip #	Syn	Or	Type	Te	Tm ice	Tm hyd	SALice	Tm (L)	Ta	Tf
PP-2-4b	1	-	??	L-V					266		
PP-2-4b	3	-		L-V	-10	-3.6		5.8	243		
PP-2-4b	3	-		L-V	-7	-2.8		4.6	244		
PP-2-4b	3	-	S	L-V					132		
PP-2-4b	3	-	S	L-V	-13	-2.2		3.7	208		
PP-2-4b	3	-	S	L	-19	-0.3		0.5			
PP-2-4b	5	-	S	L-V					257		
PP-2-4b	5	-	S	L-V					255		
PP-2-4b	6	+		L-V		-2.7		4.5	205		
PP-2-4b	6	+	S	L-V		-3.4		5.5			
PP-2-4b	7	+		L-V					186		
PP-2-5a	4	+		L-V	-7				327		
PP-3-2a	4	+	S	L-V	-21	-2.8		4.6	195		
PP-3-2a	4	+	S	L-V		-6.3		9.6			
PP-3-2a	4	+	S	L-V					235		
PP-3-2a	3	+	S	L-V	-12	-1.2		2.1	201		
PP-3-2a	3	+	S	L-V		-1.2		2.1	199		
PP-3-2a	3	+	S	L-V		-1.2		2.1	195		
PP-3-2a	1	+	S	L-V					152		
PP-3-2a	1	+	S	L-V	-20	-3.5		5.7	204		
PP-3-2a	1	+	S	L-V	-19	-3.3		5.4	215		
PP-2-5b	1	+		L-V		-3.0		4.9	269		
PP-2-5b	1	+		L-V		-4.5		7.2	267		
PP-2-5b	2	+		L-V	-29	-4.3		6.9	233		
PP-2-5b	2	+		L-V	-26	-4.3		6.9	274		
PP-2-5b	2	+		L-V		-0.2		0.4	201		
PP-2-5b	2	+		L-V	-6	-0.8		1.4	213		
PP-2-5b	3	+		L-V		-2.5		4.2	240		
PP-2-5b	3	+		L-V	-18	-1.5		2.6	215		
PP-2-5b	3	+		L-V		-1.4		2.4	221		
PP-2-5b	4	+		L-V	-17	-2.8		4.6	232		
PP-2-5b	4	+		L-V		-0.2		0.4	193		
PP-2-5b	4	+		L-V	-22	-1.4		2.4	208		
PP-2-7b	6	+	?	L-V					137		
PP-2-14	2	+	?	L-V		-1.5		2.6	256		
PP-2-14	2	+	?	L-V		-1.0		1.7	212		
PP-2-14	2	+	?	L-V	-8	-2.0		3.4	154		
PP-2-14	2	+	?	L-V		-2.1		3.5	239		
PP-2-4b	2	+	S	L-V-2	-40	-12.8	18.0	16.7		250	
PP-2-4b	2	+	S	L-V-S	-35	-12.0	15.0	16.0		320	
PP-2-4b	2	+	S	L-V-H	-57	-25.0	19.0	32.2	161	340	208

Sample	Chip #	Symb	Or	Type	Te	Tm ice	Tm hyd	SALice	Th (L)	Td	Tf H
PP-2-4b	5	+		L-V-S		-2.5		4.2			
PP-2-4b	7	+		L-V-S		-2.6		4.3	200	300	
PP-3-2a	4	+	S	L-V-S	-47	-6.8		10.3	262		
PP-3-2a	4	+	S	L-V-S	-38	-5.7		8.8	247		
PP-2-7b	6	+	?	L-V-S					146		
PP-2-14	2	+	?	L-V-S		-2.0		3.4	242		

CASSITERITE

Sample	Chip #		Or	Type	Te	Tm ice	Tm hyd	SALice	Th(L)
PP-2-7b	5	o	PS/S	L-V					271
PP-2-7b	5	o	PS/S	L-V					272
PP-2-7b	5	o	PS/S	L-V		-2.8		4.6	271
PP-2-7b	5	o	PS/S	L-V					272
PP-2-7b	5	o	PS/S	L-V		-2.8		4.6	273
PP-2-7b	5	o	PS/S	L-V					272

FLUORITE

Sample	Chip #		Or	Type	Te	Tm ice	Tm hyd	SALice	Th(L)	Td
PP-2-12c	1	x		L-V					253	
PP-2-12c	1	x		L-V					251	
PP-2-12c	1	x		L-V					226	
PP-2-12c	1	x		L-V					221	
PP-2-12c	1	x		L-V					251	
PP-2-12c	1	x		L-V					254	
PP-2-12c	3	x	S	L-V		-1.0		1.7	225	
PP-2-12c	3	x	S	L-V	-25	-1.1		1.9	225	
PP-2-12c	3	x	S	L-V	-25	-1.2		2.1	229	
PP-2-12c	3	x	S	L-V		-1.1		1.9	229	
PP-2-12c	3	x	S	L-V	-29	-1.1		1.9	225	
PP-2-12c	4	x	S	L-V		-0.9		1.6	215	
PP-2-12c	4	x	S	L-V		-0.9		1.6	210	
PP-2-12c	4	x	S	L-V	-14	-1.0		1.7	209	
PP-2-12c	4	x	S	L-V		-1.1		1.9	223	
PP-2-12c	4	x	S	L-V		-1.0		1.7	214	
PP-2-12c	1	x	P?	L-V-S					307	
PP-2-12c	1	x	P?	L-V-4S					300	
PP-2-12c	2	x	P?	L-V-3S					291	327

BERYL

Sample	Chip #		Or	Type	Te	Tm ice	Tm hyd	SALice	Th(L)	Ta	Tf S
PP-2-7b	1	*	S	L-V-S	-38				313		>400
PP-2-7b	1	*	S	L-V-S	-36	-8.6		12.4	304		>400
PP-2-7b	1	*	S	L-V-S	-32	-9.3		13.2		340	

CARBONIC INCLUSIONS

L-V*, not enough CO2 to see T-CO2
 L-L-V T* to H2O
 V-L-L T* to CO2, or critical homogenization

QUARTZ

Sample	Crp #	Type	T-CO2	T-CO2L	T-CO2V	T-CO2crT-c.th	T* iceSALce	SALce	T*	Td	XC-4	areaCO2	V CO2	X C
PP-2-4a	4	P?	L-L-V	-56.8			8.4	3.2		274				
PP-2-4a	4	P?	L-L-V	-56.9		26.8	8.4	3.2		275		0.03	0.26	150 0.04
PP-2-4a	4	P?	L-L-V	-56.8		29.0	8.3	3.4		276		0.03	0.29	136 0.05
PP-2-4a	5	?	L-L-V	-56.6	28.8						250	0.00		
PP-2-4a	6	?	L-L-V	-57.1	26.5						209	0.03		
PP-2-4a	6	?	L-L-V	-57.0	29.1		8.9	2.2			300			
PP-2-4a	6	?	L-L-V	-57.0	29.1		8.4	2.4			280	0.02		
PP-2-4a	6	?	L-L-V	-57.1	28.5							0.02		
PP-2-4a	7	?	L-L-V	-57.3	29.0							0.04		
PP-2-4a	7	?	L-L-V	-57.3	28.7		8.8	2.4				0.03		
PP-2-5a	1	?	L-L-V	-58.6	22.6		10.6	0.0			310	0.12		
PP-2-7acr	2	?	L-L-V	-59.4	23.0		9.7	0.6		355		0.12	0.54	95 0.18
PP-2-7a	2	?	L-V	-59.3			1.8	6.1			320			
PP-2-7a	6	?	L-L-V	-57.9		22.5	10.7	0.0				0.09		
PP-2-7a	7	?	L-L-V	-56.7		6.0	7.3	5.2			325	0.03		
PP-2-7a	7	?	L-L-V	-56.6		29.8					280	0.00		
PP-2-7a	7	?	L-L-V	-56.8		29.0					300	0.01		
PP-2-7a	7	?	L-L-V	-56.7	29.4						350	0.01		
PP-2-14	2	?	L-L-V	-58.9		17.5	11.3	0.0		348		0.16	0.52	148 0.12
PP-2-14	2	?	L-L-V	-57.5		26.6	10.1	0.0			240	0.07		
PP-2-14	3	P?	L-V	-56.6			9.3	-3.4	5.5	280				
PP-2-4b	1	P?	V-L-L	-57.2	27.1		8.0	3.9			260	0.03		
PP-2-4b	1	P?	V-L-L	-57.0	28.2		8.5	3.0		330		0.03	0.58	75 0.25
PP-2-4b	1	P?	V-L-L	-57.0	29.0		8.0	3.9				0.03		
PP-2-4b	1	P?	V-L-L	-57.1	27.6							0.03		
PP-2-4b	5	?	V-L	-56.9			8.7	2.6						
PP-2-4b	5	?	V-L-L	-56.8	28.8							0.02		
PP-2-4b	5	?	V-L-L	-56.7										
PP-2-4b	7	?	V-L-L	-57.3	28.6		9.5	1.0		310		0.03	0.60	76 0.26
PP-2-5a	4	?	V-L-L	-59.7	13.3		11.6	0.0				0.17		
PP-2-5a	4	?	V	-58.8	18.3							0.13		
PP-2-5a	4	?	V-L-L	-58.8		16.3	11.2	0.0		325		0.16	0.83	86 0.50
PP-3-2a	1	?	V	-58.6	23.8							0.08		
PP-3-2a	1	?	V	-58.6										
PP-3-2a	1	?	V-L-L	-59.6	23.9					314		0.12	0.86	100 0.53
PP-3-2a	1	?	V-L-L	-58.5	22.2		10.2	0.0		355		0.08	0.65	75 0.31
PP-3-2a	3	S?	V	-58.2			10.0	0.0						
PP-3-2a	3	S?	V	-58.1										
PP-3-2a	3	S	V-L-L	-58.8		23.5	10.2	0.0		340		0.11	0.85	134 0.43
PP-3-2a	3	S	V-L-L	-59.0		21.0	10.8	0.0		325		0.12	0.74	145 0.26
PP-3-2a	3	S	V-L-L	-59.5			11.3	0.0						

Sample	Chip #	Or	Type	TmCO2	ThCO2L	ThCO2V	ThCO2crTmclth	Tm iceSALclth	SALice	Th	Ta	XCH4	areaCO2	V CO2	X C
PP-3-2a	4	P?	V-L-L	-58.5		24.7	10.0	0.0			314	0.12			
PP-3-2a	4	P?	V-L-L	-58.4		25.0	10.0	0.0		340		0.12	0.56	120	0.16
PP-3-2a	4	P?	V-L-L	-58.4		24.5	10.1	0.0		340		0.12	0.55	123	0.15
PP-3-2a	4	P?	V-L-L	-58.1			10.1	0.0		340					
PP-3-2a	4	P?	V	-58.2											
PP-2-5b	1	?	V	-58.0											
PP-2-5b	1	?	V-L-L	-58.4						350					
PP-2-5b	1	?	V	-58.6											
PP-2-5b	3	?	V-L-L	-57.1	26.9		9.4	1.2			275	0.03			
PP-2-5b	3	?	V-L-L	-58.2											
PP-2-5b	3	?	V-L-L	-57.6		25.5	9.8	0.4				0.07			
PP-2-5b	3	?	V-L-L	-57.3	28.3		8.8	2.4				0.03			
PP-2-5b	4	PS?	V-L-L	-59.2			11.7	0.0							
PP-2-5b	4	PS?	V-L-L	-59.0			12.8	0.0							
PP-2-5b	4	PS?	V	-58.1											
PP-2-5b	4	PS?	V-L-L	-59.1		19.5	10.7	0.0				0.15			
PP-2-7b	2	?	V-L-L	-58.1		13.5	8.5			285		0.15	0.91	188	0.49
PP-2-7b	6	?	V-L-L	-57.2		27.2	10.3	0.0				0.03			
PP-2-7b	6	?	V-L-L	-58.0		20.5	10.9	0.0				0.05			
PP-2-7b	6	?	V-L-L	-58.2		20.6	10.9	0.0		344	380	0.12	0.58	148	0.14
PP-2-7b	6	?	V-L-L	-58.2		20.6	10.9	0.0			367	0.12			
PP-2-7b	6	?	V-L-L	-57.9		22.0	10.7	0.0		385		0.09	0.55	154	0.13
PP-2-7b	7	?	V-L-L	-56.7	29.5						347	0.02			
PP-2-14	2	?	V-L-L	-57.6		26.0	9.9	0.2			273	0.07			
PP-2-14	2	?	V-L-L	-57.7		25.0	10.1	0.0			320	0.07			
PP-2-14	2	?	V	-58.0											
PP-2-14	3	P?	V-L-L	-57.5		25.5	10.1	0.0		335		0.06	0.63	145	0.18
PP-2-14	4	?	V-L-L	-57.2	28.2		9.7	0.6				0.03			
PP-2-14	4	?	V-L-L	-57.0		28.5	9.5	1.0				0.03			
PP-2-14	4	?	V-L-L	-57.4		26.7	9.5	1.0				0.06			
PP-2-14	4	?	V-L-L	-57.2		26.8	9.5	1.0				0.05			
PP-2-14	4	?	V-L-L	-57.4		26.8	9.7	0.6				0.06			
PP-2-14	4	?	V-L-L	-57.0		28.6	9.5	1.0				0.03			
PP-2-4b	3	?	L-V*	-57.2			8.0	3.9		336					
PP-2-4b	3	?	L-V*	-56.7			8.4	3.2							
PP-2-4b	4	P?	L-V*	-56.8			8.7	2.6		275					
PP-2-4b	4	P?	L-V*	-57.0						251					
PP-2-4b	7	?	L-V*	-57.2			7.5	-7.0	4.9	10.5	280				
PP-2-4b	7	?	L-V*	-57.2						296					
PP-2-4b	7	?	L-V*	-57.3						300					
PP-2-5a	4	?	L-V*							304					
PP-2-5a	4	?	L-V-S*				11.6	0.0		294					
PP-3-2a	3	S?	L-V-S	-62.0			12.0	0.0		319					
PP-2-5b	2	?	L-V*	-57.5			8.9	-4.2	2.2	6.7	273				
PP-2-5b	4	P?	L-V*	-60.5			12.8	0.0			280				
PP-2-7b	2	?	L-V-S	-60.1			10.5	0.0			280				
PP-2-7b	7	?	L-V-S	-56.6			7.8	-5.9	4.3	9.1					
PP-2-14	3	P?	L-V*	-56.9							250				

[illegible]

Sample	Crip #	Cr	Type	T-CO2	TbCO2L	T-CO2V	T-CO2crT	clcrT	ice	SALcath	SALice	Th	Td	XC#4	areaCO2	V CO2	X C
PP-2-5a	1		L-L-V	-59.8			16.6	11.0		0.0		325		0.21	0.47	123	0.11
PP-2-5cr?	1		L-L-V	-59.8	17.9			11.0		0.0		317		0.14	0.48	83	0.17
PP-2-5a	1	S	L-L-V	-60.5	16.5			11.2		0.0				0.16			
PP-2-5a	2		L-L-V	-58.5		17.0		11.2		0.0		310		0.15	0.15	157	0.02
PP-2-5a	2		L-L-V	-59.4		16.3		11.2		0.0		310		0.19	0.31	140	0.05
PP-2-5a	2		L-L-V	-59.3	17.9							310		0.12	0.37	76	0.12
PP-2-5a	3		L-L-V	-59.3	18.0									0.12			
PP-2-5a	3		L-L-V	-58.9	17.9			11.2		0.0		305		0.10	0.37	70	0.13
PP-2-5a	3		L-L-V	-59.4	18.5			11.2		0.0		305		0.12	0.38	78	0.12
PP-2-5a	3		L-L-V	-59.3	17.4			11.7		0.0		303		0.12	0.26	75	0.08
PP-2-5a	3	S/PS	LLVS	-59.9		13.0		11.3		0.0		304		0.21	0.28	152	0.04
PP-2-5a	5	S/PS	LLV	-61.5		2.2		11.2		0.0		312		0.32	0.48	165	0.09
PP-2-5a	5	S/PS	LLV	-61.6		4.0		12.7	-5.0	0.0	7.9	306		0.32	0.45	150	0.09
PP-2-5a	5	S/PSLLV3S		-59.2		17.2		11.3		0.0		300		0.16	0.60	150	0.15
PP-2-5a	5	S/PSLLV3S		-59.0		17.2						305		0.16	0.44	150	0.09
PP-3-2cr?	2		L-L-V	-58.4		20.0		8.4		3.2		330		0.13	0.43	147	0.08
DP-1a2	1	?	L-L-V	-57.9		13.0								0.13			
DP-1a2	1	?	L-L-V	-58.8	18.1			11.1		0.0				0.10			
DP-1a2	1	?	L-L-V	-59.1	15.0			11.2		0.0				0.11			
DP-1a2	1	?	L-L-V	-58.8	18.8									0.10			
DP-1a2	1	?	L-L-V	-59.0	20.4			10.3		0.0				0.11			
DP-1a2	2	?	L-L-V	-57.5	25.4									0.03			
PP-2-5a	5	S/PS	V	-63.0													
PP-3-2cr?	2	V-L-L		-57.9		21.0	9.8		0.4			330	0.08	0.56		72	0.24
PP-2-7b	5	?	V-L-L	-59.5	16.9			11.3		0.0		405		0.14	0.52	80	0.20
DP-1a2	1	?	V-L-L	-58.2		15.0								0.14			
DP-1a2	1	?	V-L-L	-58.4	21.1			10.3		0.0				0.10			
DP-1a2	1	?	V-L-L	-58.2				10.4		0.0							

Sample	Chip #	Or	Type	TmCO2	ThCO2L	ThCO2V	ThCO2crTm	clthTm	ice	SALclth	SALice	Th	Td	XCH4	areaCO2	V CO2	X C
DP-1a2	1	?	V	-58.8	17.5										0.12		
DP-1a2	1	?	V-L-L	-58.8	17.1										0.12		
DP-1a2	1	?	V-L-L	-59.2	17.0										0.12		
DP-1a2	1	?	V	-58.3	22.2			9.7		0.6					0.08		
DP-1a2	2	?	V-L-L	-57.6		22.0		10.7		0.0					0.09		
DP-1a2	2	?	V-L-L	-57.4				10.5		0.0							
DP-1a2	2	?	V	-57.6													
DP-1a2	2	?	V-L-L	-57.7	24.5			11.4		0.0					0.06		
PP-2-5a	3		L-V*									303					
PP-2-5a	5	S/PS	L-V*					12.2		0.0		304					
PP-3-2a	2		L-V*	-59.3				10.4		0.0		310					
PP-3-2a	2	*	L-V-S	-59.5													

BERYL

Sample	Chip #	Or	Type	TmCO2	ThCO2	ThCO2V	ThCO2crTm	clthTm	ice	SALclth	SALice	Th	Td	XCH4	areaCO2	V CO2	X C
PP-2-7b	3	P	L-L-V	-58.5		16.0		10.3		0.0					0.15		
PP-2-7b	3	P	L-L-V	-57.9		22.0		9.3		1.4			400		0.09		
PP-2-7b	4	?	L-L-V	-58.8		18.0		10.9		0.0		332			0.14	0.42	156 0.06
PP-2-7b	4	?	L-L-V	-58.2				11.1		0.0		345					
PP-2-7cr	4	?	L-L-V	-59.4		6.0		11.4		0.0		355			0.26	0.46	177 0.08
PP-2-7b	1	P?	V-L-L	-58.0		22.5		9.1		1.8		330			0.10	0.68	145 0.21
PP-2-7b	1	S/PS	VLL	-57.8		15.0						335			0.13	0.83	188 0.32
PP-2-7b	1	P?	V-L-L	-58.1			22.4	9.1		1.8		340			0.09	0.83	79 0.53
PP-2-7b	1	P?	V-L-L	-58.9								325					
PP-2-7b	2	PS	V-L-L	-60.0		10.5		11.5		0.0		335			0.25	0.68	146 0.21
PP-2-7b	2	PS	V-L-L	-59.1		14.0		11.5		0.0		400			0.19	0.38	156 0.07
PP-2-7b	2	PS	V-L-L	-59.8				11.3		0.0							
PP-2-7b	2	PS	V-L-L	-57.4		16.0						325			0.09	0.60	206 0.12
PP-2-7b	2	PS	V-L-L	-58.3													
PP-2-7b	2	PS	V-L-L	-57.4													
PP-2-7b	3	P	V-L-L	-58.1		21.9		9.9		0.2		295			0.11	0.82	144 0.36
PP-2-7b	3	P	V-L-L	-58.1		23.4		10.1		0.0		290			0.10	0.83	139 0.39
PP-2-7b	4	?	V-L-L	-58.9		16.5		11.3		0.0		350			0.17	0.59	149 0.15
PP-2-7b	4	?	V-L-L	-57.3		26.6		9.9		0.2		375			0.06	0.64	138 0.19
PP-2-7b	4	?	V-L-L	-57.7		20.0		10.3		0.0		365			0.09	0.60	169 0.14
PP-2-7b	4	S/PSV		-58.2		20.0									0.12		
PP-2-7b	3	P?	L-V*						-3.3		5.4	256					
PP-2-7b	3	P?	L-V-2s*					9.7		0.6		227					

APPENDIX VI
ELECTRON MICROPROBE DATA FROM
TAKUA PIT THONG

BIOTITE

Sample	2b	2b	2b	2b	2b	2b	2b	2b	2b	2b	6c	6c	6c
colour	green	green	brown	brown	green	brown	green	brown	green	brown	green	green	green
FeO	27.00	26.79	28.07	29.37	24.67	22.31	27.67	28.07	17.87	14.41	26.40	25.26	24.37
MnO	0.21	0.21	0.19	0.19	0.17	0.15	0.26	0.23	0.18	0.20	0.15	0.17	0.11
Na2O	0.03	0.06	0.02	0.02	0.02	0.03	0.04	0.06	0.03	0.05	0.04	0.05	0.00
Al2O3	13.80	13.40	14.82	15.23	12.47	12.04	16.37	15.69	13.32	13.45	13.83	13.75	17.42
SiO2	38.21	38.80	36.05	35.20	40.45	41.37	37.07	36.12	42.78	42.26	37.30	38.54	38.02
F	1.02	1.49	1.08	1.12	2.02	2.23	0.92	1.14	3.17	2.91	2.39	2.39	2.37
HgO	8.19	8.03	6.67	6.53	10.23	10.47	7.02	6.86	12.74	17.02	8.86	9.80	8.28
K2O	9.64	9.27	9.88	9.25	7.00	9.35	9.65	9.56	9.24	8.94	9.79	9.75	9.52
CaO	0.00	0.03	0.03	0.02	0.00	0.02	0.01	0.03	0.03	0.02	0.00	0.01	0.00
TiO2	0.35	0.29	1.12	0.94	0.26	0.89	0.14	0.30	0.38	0.08	0.26	0.24	0.30
SnO2	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
O=F	0.43	0.63	0.45	0.47	0.85	0.94	0.39	0.48	1.34	1.22	1.00	1.01	1.00
Tot	98.03	97.74	97.49	97.40	96.44	97.93	98.77	97.58	98.41	98.09	98.01	98.95	99.37
Fe	3.53	3.52	3.74	3.93	3.23	2.89	3.59	3.72	2.27	1.80	3.54	3.32	3.15
Mn	0.03	0.03	0.03	0.03	0.02	0.02	0.03	0.03	0.02	0.03	0.02	0.02	0.01
Hg	1.91	1.88	1.58	1.56	2.39	2.41	1.62	1.62	2.89	3.79	2.11	2.29	1.91
Al VI	2.51	2.48	2.52	2.51	2.30	2.19	2.74	2.66	2.39	2.37	2.59	2.54	3.04
Al IV	0.03	0.00	0.26	0.36	0.00	0.00	0.25	0.27	0.00	0.00	0.03	0.00	0.13
Si	5.97	6.10	5.74	5.64	6.33	6.40	5.75	5.71	6.50	6.32	5.97	6.05	5.87
Ti	0.04	0.03	0.13	0.11	0.03	0.10	0.02	0.04	0.04	0.01	0.03	0.03	0.03
Sn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.01	0.02	0.01	0.01	0.00	0.01	0.01	0.02	0.01	0.01	0.01	0.02	0.00
K	1.92	1.86	2.01	1.89	1.40	1.84	1.91	1.93	1.79	1.70	2.00	1.95	1.88
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
F	0.50	0.74	0.54	0.57	1.00	1.09	0.45	0.57	1.53	1.38	1.21	1.19	1.16
Sample	6c	6c	9c	9c	9c	9c	9c	9c	9c	14d	14d	14d	16a
colour	green	green	green	green	green	green	green	green	green	brown	brown	brown	brown
FeO	24.92	24.81	27.34	26.73	26.53	26.14	29.18	30.43	29.69	7.28	1.80	1.87	20.81
MnO	0.16	0.20	0.23	0.21	0.26	0.22	0.26	0.22	0.19	0.07	0.00	0.00	0.15
Na2O	0.10	0.05	0.00	0.03	0.03	0.02	0.00	0.00	0.02	0.24	0.51	0.54	0.08
Al2O3	14.01	13.65	11.53	11.44	11.54	11.50	12.51	13.84	14.51	13.06	14.27	12.34	12.72
SiO2	37.83	38.40	38.15	38.62	38.44	38.26	37.81	35.72	35.50	41.45	41.71	43.08	39.61
F	2.88	3.24	2.63	2.69	2.53	2.31	1.71	1.41	1.12	5.07	5.44	5.82	3.10
HgO	10.09	10.49	9.88	9.64	9.92	9.75	8.00	6.58	6.24	23.54	26.91	27.82	13.55
K2O	9.37	9.70	9.29	9.31	9.83	9.67	9.66	9.45	9.48	10.58	10.45	10.52	9.85
CaO	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.03	0.00	0.02
TiO2	0.22	0.18	0.53	0.61	0.67	0.63	0.69	0.32	0.64	0.44	0.65	0.35	0.10
SnO2	nd	nd	nd	nd	0.02	0.00	0.01	0.00	0.07	0.01	0.00	0.00	0.02
O=F	1.21	1.41	1.11	1.13	1.06	0.97	0.72	0.59	0.47	2.13	2.29	2.45	1.30
Tot	98.39	99.42	98.47	98.13	98.70	97.53	99.11	97.37	96.94	99.62	99.47	99.89	98.71
Fe	3.32	3.29	3.67	3.59	3.54	3.51	3.87	4.13	4.01	0.91	0.22	0.23	2.71
Mn	0.02	0.03	0.03	0.03	0.03	0.03	0.04	0.03	0.03	0.01	0.00	0.00	0.02
Hg	2.39	2.48	2.37	2.31	2.56	2.34	1.89	1.59	1.50	5.89	6.10	6.10	3.15
Al VI	2.63	2.55	2.18	2.16	2.17	2.18	2.33	2.44	2.49	2.31	2.47	2.14	2.34
Al IV	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.21	0.27	0.00	0.00	0.00	0.00
Si	6.02	6.09	6.12	6.20	6.14	6.15	5.99	5.79	5.73	6.21	6.12	6.34	6.18
Ti	0.03	0.02	0.06	0.07	0.08	0.08	0.08	0.04	0.08	0.05	0.07	0.04	0.01
Sn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.03	0.02	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.07	0.15	0.15	0.03
K	1.90	1.96	1.90	1.91	2.00	1.98	1.95	1.95	1.95	2.02	1.96	1.97	1.96
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	1.45	1.68	1.33	1.37	1.27	1.17	0.86	0.72	0.57	2.40	2.52	2.71	1.53
Sample	16a	16a	16a	18a	18a	18a	18a	18a	31c	31c	31c	31c	31c
colour	brown	brown	brown	green	green	green	green	green	brown	brown	brown	green	green
FeO	17.05	18.89	17.17	33.91	32.97	33.82	32.99	34.60	24.78	20.21	23.79	27.48	21.67
MnO	0.12	0.16	0.13	0.42	0.32	0.32	0.36	0.32	0.18	0.19	0.21	0.25	0.19
Na2O	0.10	0.13	0.13	0.03	0.01	0.03	0.02	0.04	0.07	0.10	0.05	0.06	0.08
Al2O3	11.78	13.15	13.38	12.74	14.61	14.09	13.94	14.42	15.45	15.43	16.12	16.30	15.07
SiO2	41.72	39.99	39.56	37.40	36.60	36.00	34.89	36.22	37.19	39.41	37.04	35.75	37.73
F	4.02	3.43	3.54	0.81	0.83	0.64	0.75	0.87	1.42	2.30	1.36	1.05	2.02
HgO	16.72	14.93	16.00	4.22	4.68	4.18	4.11	4.25	8.01	10.51	8.39	7.43	8.78
K2O	10.03	9.81	10.13	9.13	9.07	9.36	9.19	9.27	9.63	9.80	9.82	8.58	9.84
CaO	0.02	0.04	0.04	0.00	0.00	0.00	0.00	0.02	0.01	0.01	0.00	0.00	0.00
TiO2	0.04	0.10	0.08	0.18	0.18	0.16	0.15	0.18	1.56	0.21	0.68	0.20	0.10
SnO2	0.01	0.01	0.03	0.00	0.03	0.04	0.02	0.06	0.03	0.02	0.06	0.00	0.05
O=F	1.69	1.45	1.49	0.34	0.35	0.27	0.32	0.37	0.60	0.97	0.57	0.44	0.85
Tot	99.91	99.21	98.68	98.51	98.97	98.36	96.11	99.89	97.71	97.22	96.93	96.66	94.69
Fe	2.18	2.44	2.22	4.55	4.38	4.54	4.55	4.60	3.24	2.62	3.12	3.64	2.91
Mn	0.02	0.02	0.02	0.06	0.04	0.04	0.05	0.04	0.02	0.02	0.03	0.03	0.03
Hg	3.81	3.43	3.69	1.01	1.11	1.00	1.01	1.01	1.86	2.43	1.96	1.76	2.10
Al VI	2.12	2.39	2.44	2.41	2.54	2.45	2.46	2.46	2.65	2.82	2.78	2.71	2.86
Al IV	0.00	0.00	0.00	0.00	0.19	0.22	0.24	0.24	0.19	0.00	0.20	0.33	0.00
Si	6.38	6.17	6.12	6.00	5.81	5.78	5.76	5.76	5.81	6.11	5.80	5.67	6.07
Ti	0.00	0.01	0.01	0.02	0.02	0.02	0.02	0.02	0.18	0.02	0.08	0.02	0.01
Sn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.03	0.04	0.04	0.01	0.00	0.01	0.01	0.01	0.02	0.03	0.02	0.02	0.03
K	1.96	1.93	2.00	1.87	1.84	1.92	1.93	1.88	1.92	1.94	1.96	1.74	2.02
Ca	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	1.95	1.68	1.73	0.41	0.42	0.32	0.39	0.44	0.70	1.13	0.67	0.53	1.03

Sample	31c	31c	31c	31c	31c	31c	31c	31c	31c	31c	31c	40b	40b
colour	green	green	green	green	green	green	green	green	green	green	green	brown	brown
FeO	23.59	19.42	19.59	24.63	19.58	3.00	21.96	3.96	24.29	4.18	24.60	19.52	19.15
MnO	0.16	0.13	0.16	0.19	0.15	0.01	0.18	0.00	0.20	0.06	0.18	0.14	0.13
Na2O	0.04	0.04	0.07	0.01	0.04	0.20	0.02	0.20	0.02	0.20	0.04	0.04	0.04
Al2O3	14.12	12.27	12.24	14.05	12.24	13.54	13.00	13.53	13.40	13.35	13.42	15.48	15.88
SiO2	39.00	42.53	42.10	36.67	42.15	42.62	38.45	42.55	38.13	42.18	37.78	19.06	39.84
F	1.85	2.96	3.14	1.65	2.91	5.87	1.98	5.91	1.89	5.67	1.92	2.02	2.09
HgO	9.63	13.60	12.64	9.82	12.71	25.45	10.28	25.76	9.94	24.49	9.93	11.14	11.12
K2O	9.83	9.72	9.85	8.90	9.77	11.05	9.61	11.10	9.45	10.94	9.41	10.13	10.04
CaO	0.00	0.00	0.02	0.09	0.02	0.01	0.05	0.02	0.04	0.02	0.01	0.00	0.00
TiO2	0.15	0.20	0.23	0.27	0.17	0.29	0.26	0.32	0.30	0.31	0.24	1.26	1.10
SnO2	0.00	0.02	0.00	0.04	0.02	0.00	0.02	0.03	0.02	0.00	0.00	0.00	0.02
O-F	0.78	1.25	1.32	0.70	1.22	2.47	0.83	2.49	0.79	2.39	0.81	0.85	0.88
Tot	97.59	99.64	98.70	95.62	98.54	99.58	96.99	100.91	96.89	99.01	96.73	97.94	98.54
Fe	3.08	2.46	2.52	3.30	2.51	0.37	3.17	0.49	3.22	0.52	3.28	2.50	2.43
Mn	0.02	0.02	0.02	0.03	0.02	0.00	0.02	0.00	0.03	0.01	0.02	0.02	0.02
Hg	2.24	3.07	2.90	2.35	2.91	3.64	2.43	5.66	2.35	5.47	2.36	2.54	2.51
Al VI	2.60	2.19	2.22	2.54	2.21	2.37	2.43	2.35	2.50	2.36	2.52	2.77	2.83
Al IV	0.00	0.00	0.00	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00
Si	6.09	6.44	6.48	5.88	6.47	6.33	6.09	6.27	6.04	6.32	6.01	5.98	6.01
Ti	0.02	0.02	0.03	0.03	0.02	0.03	0.03	0.04	0.04	0.04	0.01	0.14	0.13
Sn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.01	0.01	0.02	0.00	0.01	0.06	0.01	0.06	0.01	0.06	0.01	0.01	0.01
K	1.96	1.88	1.94	1.82	1.91	2.09	1.94	2.09	1.91	2.09	1.91	1.98	1.94
Ca	0.00	0.00	0.00	0.02	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
F	0.92	1.42	1.53	0.84	1.41	2.76	0.99	2.76	0.95	2.69	0.97	0.98	1.00

Sample	40b	40b	45	45
colour	brown	brown	brown	brown
FeO	20.47	16.28	20.62	21.81
MnO	0.18	0.16	0.06	0.09
Na2O	0.03	0.06	0.05	0.04
Al2O3	15.00	13.56	15.32	16.01
SiO2	37.67	40.62	38.19	37.99
F	2.04	3.22	2.13	1.69
HgO	10.80	14.61	11.75	10.81
K2O	10.11	10.50	10.04	9.66
CaO	0.00	0.01	0.02	0.02
TiO2	1.47	1.28	0.00	0.00
SnO2	0.02	0.00	0.02	0.04
O-F	0.86	1.35	0.90	0.71
Tot	96.94	98.94	97.30	97.44
Fe	2.68	2.08	2.68	2.82
Mn	0.02	0.02	0.01	0.01
Hg	2.52	3.32	2.73	2.49
Al	2.76	2.44	2.81	2.92
Al VI	2.65	2.44	2.75	2.79
Al IV	0.11	0.00	0.06	0.13
Si	5.89	6.19	5.94	5.87
Ti	0.17	0.15	0.00	0.00
Sn	0.00	0.00	0.00	0.00
Na	0.01	0.02	0.01	0.01
K	2.02	2.04	1.99	1.91
Ca	0.00	0.00	0.00	0.00
F	1.01	1.55	1.05	0.82

FELDSPARS

Sample #	91a3	91a3	91a3	91a3	91a3	91a3	91a3	91a3	91a3	24b	24b	24b	24b	24b
Anal #	1	2	4	5	6	7	8	9	1	2	3	4	5	
Descrip	pl incl	pl c	untw	pl incl	untw	pl c	pl cas	pl rpl	untw	pl c	ksp rpl	pl c	pl untw	pl untw
	qtz	con	cas	cas	cas	cas	cas	cas	cas	cas	pl	cas	pl	pl tw
FeO	0 00	0 00	0 00	0 00	0 07	0 01	0 02	0 00	0 29	0 28	0 06	0 10	0 25	
Na2O	11 56	11 62	0 25	11 54	11 56	11 71	11 52	11 65	0 23	11 11	11 13	10 98	10 83	
SiO2	68 17	68 06	64 19	68 16	67 84	68 00	67 25	67 73	64 28	66 59	66 88	66 45	66 00	
Al2O3	19 80	19 57	18 12	19 51	19 71	17 74	19 79	19 64	18 00	20 18	20 44	20 81	20 47	
MgO	0 03	0 00	0 00	0 03	0 01	0 00	0 00	0 01	0 00	0 00	0 02	0 00	0 00	
K2O	0 07	0 08	17 37	0 05	0 12	0 09	0 09	0 13	17 24	0 55	0 17	0 10	0 15	
CaO	0 34	0 41	0 00	0 36	0 45	0 45	0 45	0 36	0 01	0 64	1 31	1 44	1 30	
BaO	0 01	0 06	0 01	0 00	0 00	0 08	0 08	0 01	0 09	0 02	0 02	0 00	0 06	
Tot	99 97	99 75	99 93	99 66	99 77	100 08	100 09	99 53	100 15	99 38	100 03	99 88	99 05	
Fe	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 01	0 01	0 00	0 00	0 01	
Na	0 98	0 99	0 02	0 98	0 98	1 01	0 99	0 99	0 07	0 95	0 95	0 94	0 93	
Si	2 98	2 98	2 99	2 99	2 98	3 03	2 97	2 98	2 99	2 94	2 94	2 92	2 93	
Al	1 02	1 01	0 99	1 01	1 02	0 93	1 03	1 02	0 99	1 05	1 06	1 08	1 07	
Mg	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	
K	0 00	0 00	1 03	0 00	0 01	0 01	0 01	0 01	1 02	0 03	0 01	0 01	0 01	
Ca	0 02	0 02	0 00	0 02	0 02	0 02	0 02	0 02	0 00	0 03	0 06	0 07	0 06	
Ba	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	
X ab	0 98	0 98	0 02	0 98	0 97	0 97	0 97	0 98	0 02	0 94	0 93	0 93	0 93	
X an	0 02	0 02	0 00	0 02	0 02	0 02	0 02	0 02	0 00	0 03	0 06	0 07	0 06	
X or	0 00	0 00	0 98	0 00	0 01	0 00	0 01	0 01	0 98	0 03	0 01	0 01	0 01	

Sample #	24b	24b	65b	65b	65b	65b	73g	73g	73g	73g	73g	1a	1a
Anal #	6	7	1	2	3	4	1	2	3	4	5	1	2
Descrip	ksp rpl	pl	ksp untw	ksp tw	pl tw	pl con	pl tw	pl c	pl sg	2 pl	pl c	ksp rpl	pl tw
	cas	cas	cas	cas	cas	cas	cas	cas	cas	cas	cas	cas	cas
FeO	0 02	0 04	0 00	0 01	0 00	0 01	0 00	0 07	0 01	0 00	0 00	0 02	0 03
Na2O	0 57	11 43	0 32	11 39	11 48	0 31	11 75	11 42	11 75	11 73	0 32	10 13	10 09
SiO2	64 39	67 41	64 07	66 63	65 93	62 72	66 05	65 54	67 96	68 03	64 72	65 09	65 31
Al2O3	18 34	19 67	18 18	19 96	20 09	18 12	19 80	19 87	19 80	19 58	18 06	21 63	21 44
MgO	0 00	0 00	0 01	0 00	0 01	0 01	0 00	0 00	0 00	0 01	0 00	0 01	0 00
K2O	16 81	0 10	17 07	0 11	0 14	17 05	0 06	0 09	0 13	0 14	16 90	0 18	0 17
CaO	0 02	0 47	0 01	0 66	0 81	0 02	0 35	0 56	0 51	0 34	0 01	2 68	2 57
BaO	0 11	0 00	0 03	0 00	0 00	0 07	0 06	0 00	0 00	0 00	0 00	0 00	0 01
Tot	100 27	99 12	99 69	98 76	98 46	98 31	98 07	97 55	100 17	99 84	100 01	99 73	99 61
Fe	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00
Na	0 05	0 98	0 03	0 98	0 99	0 03	1 02	1 00	1 00	1 00	0 03	0 87	0 86
Si	2 99	2 97	2 99	2 96	2 94	2 97	2 95	2 95	2 97	2 98	3 00	2 87	2 88
Al	1 00	1 02	1 00	1 04	1 06	1 01	1 04	1 05	1 02	1 01	0 99	1 13	1 12
Mg	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00
K	0 99	0 01	1 02	0 01	0 01	1 03	0 00	0 01	0 01	0 01	1 00	0 01	0 01
Ca	0 00	0 02	0 00	0 03	0 04	0 00	0 02	0 03	0 02	0 02	0 00	0 13	0 12
Ba	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00	0 00
X ab	0 05	0 97	0 03	0 96	0 96	0 03	0 98	0 97	0 97	0 98	0 03	0 86	0 87
X an	0 00	0 02	0 00	0 03	0 04	0 00	0 02	0 03	0 02	0 02	0 00	0 13	0 12
X or	0 95	0 01	0 97	0 01	0 01	0 97	0 00	0 01	0 01	0 01	0 97	0 01	0 01

Sample #	1a	1a	1a	110d	40b	40b
Anal #	3	4	5	5	4	5
Descrip	pl ser	pl ser	pl ser	pl con	pl con	pl con
	bio	bio	bio	cpx	bio	bio
FeO	0 00	0 04	0 04	0 06	0 12	0 06
Na2O	11 68	11 33	11 72	6 36	7 41	11 38
SiO2	68 68	68 04	68 51	58 42	60 04	69 70
Al2O3	19 81	19 80	19 47	26 61	25 77	19 74
MgO	0 00	0 00	0 00	0 00	0 00	0 00
K2O	0 06	0 07	0 05	0 00	0 17	0 16
CaO	0 39	0 69	0 18	8 82	7 58	0 41
BaO	0 06	0 00	0 01	nd	nd	nd
Tot	100 67	99 96	99 99	100 60	101 25	100 78
Fe	0 00	0 00	0 00	0 00	0 00	0 00
Na	0 98	0 96	0 99	0 55	0 63	0 96
Si	2 98	2 98	2 99	2 60	2 65	2 99
Al	1 01	1 02	1 00	1 40	1 34	1 01
Mg	0 00	0 00	0 00	0 00	0 00	0 00
K	0 00	0 00	0 00	0 00	0 01	0 01
Ca	0 02	0 03	0 01	0 42	0 36	0 02
Ba	0 00	0 00	0 00	0 00	0 00	0 00
X ab	0 98	0 96	0 99	0 57	0 63	0 97
X an	0 02	0 03	0 01	0 43	0 36	0 02
X or	0 00	0 00	0 00	0 00	0 01	0 01

CALC-SILICATE AND MISCELLANEOUS MINERALS

amphibole

Sample	2b	14d	14d	14d	14d	118d
FeO	5.66	12.93	0.90	11.38	13.20	21.98
MnO	0.27	0.42	0.05	0.32	0.45	0.32
Na2O	0.18	0.25	0.52	0.15	0.26	0.98
Al2O3	1.38	0.35	0.75	0.34	0.43	11.24
SiO2	56.99	55.70	58.20	56.07	54.98	43.11
F	1.11	1.45	1.05	1.55	1.32	0.36
MgO	20.72	16.48	24.50	17.42	15.76	6.00
K2O	0.11	0.16	0.12	0.10	0.19	1.50
CaO	13.35	12.12	12.99	12.51	12.14	12.05
TiO2	0.01	0.05	0.04	0.02	0.09	0.11
SnO2	nd	0.00	0.00	0.01	0.00	0.37
Tot	99.77	99.92	99.12	99.88	98.81	98.02

sample 118d with cassiterite

idocrase

Sample	74c	74c
FeO	2.92	2.74
MnO	0.06	0.04
Na2O	0.06	0.08
Al2O3	16.99	17.61
SiO2	37.26	37.91
F	1.97	2.15
MgO	1.73	2.01
K2O	0.00	0.00
CaO	35.52	36.17
TiO2	1.73	1.13
SnO2	0.02	0.03
Tot	98.25	99.87

titanite

Sample	2b	31c	31c	99e	99e
FeO	0.38	0.27	0.18	0.84	1.15
MnO	0.00	0.06	0.01	0.10	0.12
Na2O	0.14	0.05	0.01	0.03	0.01
Al2O3	6.58	1.91	1.08	6.60	6.38
SiO2	33.61	30.69	31.24	28.59	28.83
F	2.31	0.35	0.21	nd	nd
MgO	0.02	0.00	0.00	0.04	0.25
K2O	0.10	0.03	0.01	nd	nd
CaO	27.67	28.30	28.43	27.23	27.25
TiO2	28.36	38.25	40.88	23.24	25.04
SnO2	nd	0.30	0.02	5.42	4.13
Tot	99.17	100.21	102.07	92.09	93.16

garnet

Sample	99e	99e	99e	99e	99e	93a	93a	93a
FeO	6.92	6.79	7.29	6.71	7.69	7.31	7.58	6.66
MnO	7.05	6.59	6.18	6.15	6.86	7.16	7.17	6.17
Na2O	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.01
Al2O3	19.16	19.07	18.29	18.58	19.46	19.20	19.39	19.16
SiO2	36.94	36.62	36.58	36.36	36.62	36.91	36.90	36.97
F	nd	nd	nd	nd	nd	nd	nd	nd
MgO	0.05	0.06	0.04	0.03	0.04	0.10	0.07	0.07
K2O	nd	nd	nd	nd	nd	nd	nd	nd
CaO	26.62	27.09	27.84	28.04	25.51	25.80	25.73	27.49
TiO2	0.21	0.14	0.14	0.15	0.19	0.13	0.10	0.14
SnO2	0.06	0.05	0.09	0.09	0.03	0.09	0.03	0.07
Tot	97.01	96.40	96.46	96.12	96.39	96.70	96.97	96.73
XGr	0.6680	0.6786	0.6693	0.6835	0.6580	0.6581	0.6552	0.6911
XSp	0.1398	0.1304	0.1175	0.1184	0.1398	0.1444	0.1442	0.1226
XAlm	0.0595	0.0514	0.0364	0.0316	0.0888	0.0750	0.0794	0.0546
XPyr	0.0018	0.0019	0.0013	0.0011	0.0015	0.0033	0.0024	0.0024
XAndr	0.1307	0.1375	0.1752	0.1651	0.1116	0.1189	0.1184	0.1290

XGr grossular, XSp spessartin, XAlm almandine
XPyr pyrope, XAndr andradite

pyroxene

Sample	2b	2b	74c	74c	118d	118d	93a	93a	93a	93a	93a	93a	93a
FeO	1.27	3.51	4.75	5.46	7.79	6.37	4.93	7.90	8.63	5.35	13.28	12.53	14.45
MnO	0.06	0.14	0.12	0.11	0.34	0.19	0.96	1.43	1.61	0.94	2.36	2.08	1.91
Na2O	0.07	0.08	0.10	0.09	0.13	0.16	0.06	0.10	0.06	0.06	0.09	0.09	0.09
Al2O3	0.21	0.46	0.28	0.20	0.39	0.37	0.31	0.44	0.28	0.23	0.23	0.40	1.03
SiO2	55.86	55.07	55.31	55.88	54.08	55.02	54.42	53.38	53.29	54.05	52.04	51.98	51.50
F	0.00	0.00	0.00	0.00	0.00	0.00	nd	nd	nd	nd	nd	nd	nd
MgO	17.79	16.87	15.70	12.06	13.34	14.27	14.96	12.54	12.09	14.79	9.02	9.10	8.56
K2O	0.01	0.00	0.00	0.05	0.01	0.01	nd	nd	nd	nd	nd	nd	nd
CaO	25.81	25.31	25.44	25.77	24.80	25.25	24.96	24.83	24.55	25.17	24.27	23.85	23.69
TiO2	0.03	0.06	0.00	0.00	0.00	0.00	0.03	0.00	0.01	0.13	0.02	0.00	0.02
SnO2	nd	nd	0.00	0.00	0.01	0.00	nd	nd	nd	nd	nd	nd	nd
Tot	101.11	101.50	101.70	98.63	100.89	101.63	100.62	100.61	100.51	100.72	101.30	100.02	101.24

Sample	93a	93a	99e	99e	99e	99e	99e	99e	99e	99e	99e	94a	94a
FeO	15.22	14.81	14.20	11.27	11.53	12.45	13.50	14.48	14.87	16.30	14.33	7.73	6.13
MnO	2.04	1.92	2.25	1.67	1.71	1.56	1.62	2.96	3.03	2.26	2.41	1.42	1.07
Na2O	0.08	0.08	0.10	0.07	0.11	0.11	0.08	0.11	0.08	0.11	0.07	0.10	0.08
Al2O3	0.26	0.25	0.27	0.26	0.29	0.22	0.28	0.30	0.27	0.22	0.22	0.34	0.32
SiO2	52.11	51.91	52.14	52.38	52.65	52.78	51.93	51.86	51.55	51.14	51.77	53.55	54.30
F	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
MgO	7.93	8.27	8.37	10.52	10.32	10.12	9.29	7.67	7.17	6.85	8.30	12.48	14.15
K2O	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
CaO	24.06	24.14	23.99	24.54	24.35	23.97	24.01	23.75	23.56	23.76	23.92	34.85	24.89
TiO2	0.00	0.02	0.00	0.02	0.03	0.00	0.02	0.00	0.00	0.01	0.03	0.03	0.01
SnO2	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Tot	101.69	101.38	101.31	100.73	100.98	101.21	100.73	101.13	100.52	100.65	101.05	100.49	100.96

Sample	94a	94a	94a	94a	94a	97c	97c	97c	97c	99e
FeO	8.37	8.10	8.00	10.50	11.23	14.78	14.76	14.84	9.33	16.34
MnO	1.52	0.80	0.82	1.80	1.71	2.11	2.62	2.27	1.49	1.37
Na2O	0.09	0.13	0.07	0.20	0.15	0.08	0.05	0.08	0.07	0.12
Al2O3	0.29	0.15	0.14	0.39	0.35	0.33	0.26	0.29	0.40	0.30
SiO2	53.65	53.77	54.02	53.05	53.03	52.05	51.96	52.09	53.20	48.75
F	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
MgO	12.20	13.49	13.59	10.74	10.19	8.18	8.02	8.15	11.72	8.25
K2O	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
CaO	24.77	24.33	24.53	24.18	24.27	23.96	23.73	23.67	24.76	23.61
TiO2	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.02	0.03	0.00
SnO2	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.00
Tot	100.90	100.77	101.18	100.86	100.93	101.49	101.40	101.41	101.00	98.72

magnetite

Sample	9c	9c	18a
FeO	89.92	89.86	92.97
MnO	0.10	0.08	0.04
Na2O	0.03	0.00	0.01
Al2O3	0.07	0.05	0.08
SiO2	2.03	1.26	0.47
F	0.00	0.01	0.00
MgO	0.23	0.12	0.00
K2O	0.00	0.01	0.00
CaO	0.07	0.00	0.01
TiO2	0.11	0.07	0.02
SnO2	nd	0.01	0.01
Tot	92.55	91.46	93.62

cassiterite

Sample	99e
FeO	0.72
MnO	0.04
Na2O	0.00
Al2O3	0.48
SiO2	4.55
F	nd
MgO	0.01
K2O	nd
CaO	0.00
TiO2	0.23
SnO2	93.12
Tot	99.14

APPENDIX VII
FLUID INCLUSION DATA FROM
TAKUA PIT THONG

AQUEOUS INCLUSIONS

Salinities in L-V incl. calc by eq. of Oakes et al (1990)
for Tmice > -20 assumed XNaCl 1.0, for < -20 XNaCl 0.1

Salinities for L-V-H incl. calc by eq. of Bischoff (1991)

L-V +/-S Cassiterite

Sample	Tmice	TmCO2	Tmclath	Tmhydr1	Tmhydr2	Salinity	Th
6c							466
6c	-23.5					22.4	454
6c	-23.5					22.4	458
6c	-22.5					22.4	467
6c							474
6c							470
6c							471
6c	-25					23.1	470
6c	-27.6				15.1	24.2	201
6c	-27.1				14.3	24.0	392
6c	-27.3				14.6	24.1	195
6c	-27.2			0.5	15.1	24.0	200
6c	-27.3				15.4	24.1	
35					261		
35					262		418
35					267		422
35			-15				475
35			-18.4				459
35			-17.8				
35			-14.6				
35	-11.4		7.9			15.4	404
35	-28.6		-0.8			24.6	
35			-0.1				409
35	-5.8		2.6			9.0	429
35			-1.3				422
35			-0.2				414
35	-3.2					5.3	377
35	-4.2					6.7	

Quartz L-V * with CO2

Sample	Tmice	TmCO2	Tmclath	Tmhydr	Salinity	Th
6c				14.9		
6c				14.7		288
6c						
20a	0				0.0	203
20a	-0.2				0.4	214
20a	-1.8				3.1	221
20a	-1.6				2.7	221
20a	-2.9				4.8	245
20a	-3.3				5.4	
20a	-2.9				4.8	238
20a	-3.1				5.1	238
20a	-3				4.9	229
20a*	-4.8		8.9		7.6	252
20a*	-4.8	-57	9.6		7.6	259
20a	-3				4.9	229
20a*	-6.5	-57.6	9.7		9.9	282
20a*	-4.8		8.9		7.6	252

Sample	Tmice	TmCO2	Tmclath	Tmhydr	Salinity	Th
20a	-3.1				5.1	235
20a*	-3.8	-57.7	10.6		6.2	275
20a	-2.1				3.5	215
20a	-6.3				9.6	220

Fluorite L-V

Sample	Tmice	TmCO2	ThCO21	Tmclath	Salinity	Th
14d	-1.3				2.2	
14d	-1				1.7	
14d				1.7		179
14d	-1.7				2.9	
14d	-1.2				2.1	
14d	-1				1.7	168
14d						125
14d						131
14d	-0.5				0.9	125
14d	-32.1			13.6	25.9	
14d	-32.2			13.6	25.9	
14d	-31.8			13	25.8	
14d	-31.8			13.6	25.8	

L-V-H +/- S

Cassiterite

Sample	Tmice	Tmclath	Tmhydr1	Tmhydr2	Th halite	Salinity	Th
6c	-19				173	30.7	469

Quartz L-V-H +/- S

Sample	Tmice	TmCO2	Tmclath	ThCO21	Tmhydr2	Th halite	Salinity	Th
6c	-30.2				22.9			
6c	-33.7				23.1			
6c	-35.1				12.1	255	34.7	
6c						222	32.9	
6c						258	34.9	
6c	-36.5				15			
6c		-65	-9.5	-41		231	33.4	
6c			-15.9			262	35.1	
6c			-14.9			229	33.3	
6c			-17.7			245	34.1	
6c						228	33.2	
6c						257	34.8	
6c						285	36.7	
6c						193	31.5	
20a	-38					268	35.5	175
20a						271	35.7	178
20a	-42					266	35.4	180
20a						275	36.0	187
20a	-25.9					218	32.7	284
20a	-26.2					206	32.1	290
20a	-31.5					261	35.1	193
20a	-31.5					273	35.9	187
20a	-38					241	33.9	175
20a	-38					241	33.9	178
20a	-39					267	35.5	181

**L-V-H +/- S
Fluorite**

Sample	Tmice	TmCO2	Tmclath	Tmhydr1	Tmhydr2	Th halite	Th
14d				-12.7			
14d				-7.1			
14d				-20.8			
14d				-24.9			
14d				-24.9			
14d				-23.2			
14d	-23						
14d	-22.8						
14d	-20.4						
14d	-32			13.6			
14d	-32			13.6			
14d	-31.8			13.9			

AQUEOUS-CARBONIC INCLUSIONS

Cassiterite, V & V-L

Sample	TmCO2	Tmice	Tmclath	ThCO2l	ThCO2v	ThCO2cr	Th tot V
6c	-63						
6c	-65.4						
6c	-66.8						
6c	-65.8						
6c	-69.8						

Quartz, L-L-V

Sample	TmCO2	Tmice	Tmclath	ThCO2l	ThCO2v	ThCO2cr	Th tot L
6c	-64.5			-20.6			
20a	-56.8		10.5		27		296
20a	-56.8		10.5		26.8		307
20a	-57.8		10.9		24.4		286
20a	-57.5		9.9				317
20a	-57.6		10.3		23.4		325
20a	-57.8		10.9				302
20a	-57.9		10.6		25.5		276
20a	-57.5		10.3		19		294
20a	-57.3		-2.1	22.6			
20a	-57.4			22.6			

Quartz, V & V-L

Sample	TmCO2	Tmice	Tmclath	ThCO2l	ThCO2v	ThCO2cr	Th tot V
6c	-64.5		-8.8	-20.6			
6c	-62.9		-3.8	-2.8			
6c	-63.7		15.5	-11.4			
6c	-64.9		15.7				
6c	-64.4		15.5				
6c	-64		4.3	-15.4			
6c	-60.4			11.8			
6c	-62.8			0.6			
6c	-64.5		-5.6	-17.5			
6c	-64.3		-7.9	-14.2			
6c	-60.2			17.5			
20a	-57		9.7				315

Fluorite, L-L-V

Sample	TmCO2	Tmice	Tmclath	ThCO2l	ThCO2v	ThCO2cr	Th L	Salinity
14d	-57.5	-14.9		19.1				18.56
14d	-57.3	-14.9		16.7				18.56
14d	-57.2	-14.7		1.9				18.39

Fluorite, V & V-L

Sample	TmCO2	Tmice	Tmclath	ThCO2l	ThCO2v	ThCO2cr	Th tot V
14d				-69.7			
14d				-80.8			
14d				-81.4			
14d				-82.2			
14d			18.7	-82			
14d					-74.2		
14d				-64.8			
14d				-64.6			
14d				-64.6			
14d	-73.6		18.6	-66.1			
14d	-73.6		19.1	-65.8			
14d	-73.3		19.3	-65.6			
14d	-73.3			-65.2			
14d	-71.2		18.1		-60.2		
14d	-71.2				-60.5		
14d	-71.3		17.9		-61		
14d	-75.2			-68			
14d	-75.2			-68			
14d	-75.4			-68.5			
14d	-75.4			-68.4			
14d	-74.4			-68.4			
14d	-68.6			-54.9			
14d	-68.9			-55.2			
14d	-68.4					-54	

APPENDIX VIII
OTHER DEPOSITS IN WESTERN THAILAND

In addition to Nong Sua, Pilok, and Takua Pit Thong the field relationships and fluid inclusion petrography of cassiterite mineralization at the Som Porn, Chat Chai, Hub Kapong, and Tap Sakae areas in the western Thai granite belt were also examined. These deposits were not studied in detail, nor was microthermometric data collected. However, very little published information is available and therefore the descriptions are included.

SOM PORN

The Som Porn deposits are located near the Burmese border, between the Pilok and Takua Pit Thong deposits (Chapter 7, Fig. 1). Stockwork, replacement, vein, and pegmatite styles of mineralization are all observed in the Som Porn area which covers approximately 0.5 x 2 km. Cassiterite is the sole ore mineral and is mined both at surface and underground in adits.

All the known mineralization in the Som Porn area is spatially associated with the contacts between granite and limestone/marble. The limestone is light grey, massive, with local pelitic phlogopite-bearing horizons. It is in contact with a shale unit, although the age relationship and nature of the contact is not clear. The shale is massive and varies in composition from shale, locally with high organic carbon contents, to poorly sorted quartzite. The granite varies texturally from a coarse-grained biotite granite with K-feldspar phenocrysts to a medium-grained equigranular biotite granite. A tourmaline-muscovite "leucogranite" was also observed, but only

in the immediate vicinity of the tin deposits. Pegmatite-aplite in the Som Porn area has two modes of occurrence. South of the Goobon deposit is a large massive pegmatite-aplite body. Both the pegmatite and aplite are composed almost entirely of quartz, feldspar, and muscovite, and layering or internal structure is not apparent. Near the Som Porn deposits pegmatite and less commonly pegmatite-aplite occur as dykes <1 m thick. These dykes are also composed largely of quartz, feldspar, and muscovite, but in contrast to the massive pegmatite-aplite body, host cassiterite mineralization and commonly contain garnet and lepidolite. Six cassiterite deposits at Som Porn were examined and sampled and the characteristic of each are described below.

Goobon is the largest deposit in the Som Porn area and was the only deposit being mined during the field investigation. The deposit occurs at the contact between a medium-grained biotite granite and limestone, at the apex of the granite intrusion. Cassiterite is largely hosted by quartz veins which cut both the granite and limestone. Quartz-cassiterite veins cutting granite typically have muscovite alteration haloes whereas no alteration haloes are visible around veins cutting limestone. Some cassiterite has also been observed disseminated in granite, in association with feldspathic alteration. However, the granite at this location is strongly weathered to clay minerals. The disseminated style of mineralization is economically very important; during the course of fieldwork one such zone yielded approximately 1 tonne of tin concentrate from 5 m³ of rock (roughly 7 % SnO₂).

In the immediate mine area the limestone is weathered/altered red (presumably Fe-oxide) and contains abundant evidence of karst dissolution and brecciation whereas outside the mine area the limestone is grey and contains no karst or breccia features. The karst/breccia has a high porosity, contains angular fragments of red limestone, and is infilled by coarse grained clear calcite crystals commonly several cm across. Cassiterite has been mined from karst/breccia in the past however, none was observed in this study, and it is not known whether the cassiterite occurred as breccia fragments or precipitated during the karst/breccia event. Minor fluorite was observed at this deposit, locally along late fractures cutting quartz veins, but is not associated with cassiterite. A minor amount of clinopyroxene-garnet skarn is also present but was not observed to host cassiterite mineralization.

The Goolang deposit hosts sulphide replacement cassiterite mineralization. The deposit was mined from an adit which has subsequently collapsed and the only mineralized samples were taken from the ore dump. Cassiterite is disseminated in either pyrite- or pyrrhotite-rich rocks similar in appearance to ore from the Takua Pit Thong deposit. A minor amount of barren clinopyroxene-garnet skarn was also observed at Goolang.

At Bon Dy My Ruak, cassiterite is hosted by pegmatite dykes. These dykes are located from the area around Bon Dy My Ruak, westward to the limestone-granite contact south of the Mahkahmong deposit. The dykes generally crosscut

granite where they lack alteration haloes. Less typically they cut limestone where skarn is developed at some but not all of the contacts. The pegmatites are composed dominantly of quartz, feldspar, and muscovite, with accessory lepidolite, cassiterite, tourmaline, and garnet. Cassiterite is typically disseminated in a quartz-muscovite matrix which is interstitial to coarser grained feldspar. Lepidolite is late along fractures cutting pegmatite. Cassiterite from pegmatite is black in contrast to the other styles of mineralization where it is red to honey coloured and only pegmatite cassiterite contains inclusions of columbite-tantalite.

The V-shape deposit is mined out and covered a very small volume, roughly 10 m³, but it produced approximately 40 tonnes of cassiterite concentrate. No mineralization was observed in situ but a grab sample of a cassiterite-quartz (90% cassiterite) cutting "leucocratic granite" was collected from the ore dump. Pegmatite dykes ± lepidolite are present in the immediate area but are not mineralized.

The Mahka mong deposit was formerly mined at surface and from a small adit. No mineralization was observed in situ but two styles of mineralization are interpreted on the basis of grab samples collected from the ore dump. The first style is quartz vein hosted. These veins cut the more pelitic phlogopite-bearing horizons of the limestone and the adit follows one of such impure limestone horizons. The second style of mineralization is hosted by calc-silicate altered limestone. Cassiterite is either disseminated within the calc-silicate alteration or in quartz-cassiterite veins cutting pervasive calc-silicate altered rocks. Abundant late stage coarse-grained

calcite veins and minor red limestone with karst/breccia features similar to Goobon also occur at this deposit. Tourmaline "leucocratic granite" is present at the base of the deposit but does not contain visible cassiterite.

The Mahkahmong Lang deposit produced cassiterite from a small adit which has subsequently collapsed and the only mineralized samples were collected from the ore dump. The deposit is located at the contact between aplite-pegmatite and limestone. Cassiterite is hosted by quartz veins and is accompanied by abundant sulphide minerals, pyrite >> arsenopyrite > chalcopyrite. Cassiterite, the sulphide minerals, and minor muscovite are present along fractures that cut the quartz vein. According to the mine owner there is a general correlation between cassiterite and arsenopyrite at this deposit. Minor feldspar and tourmaline are also present in the quartz veins and the tourmaline is altered to chlorite. The veins cut aplite-pegmatite but were not observed to cut limestone.

FLUID INCLUSION PETROGRAPHY

Four fluid inclusion types from the stockwork style mineralization are present in cassiterite and quartz. Type 1 are L-V inclusions. Type 2 are vapour-rich V and V-L inclusions. Type 3 are aqueous-carbonic vapour-rich V-L-L inclusions. Type 4 are halite-bearing L-V-H inclusions. The inclusions have rounded or elliptical shapes and are generally less than 10 μm across. Two fluid inclusion types are observed in cassiterite, types 1 and 2. Type 2 inclusions are dominant and type 1

are less common. Inclusions are distributed throughout cassiterite grains and no origin can be positively determined. All four inclusion types are present in quartz. The first 3 types are common whereas type 4 inclusions are rare. Most inclusions in quartz occur along secondary planes.

Fluid inclusion types 1 and 2 are observed in the pegmatite style mineralization. Both types are present in quartz where they occur in secondary planes. Type 1 is dominant and type 2 less common. Fluid inclusions in cassiterite are very poorly preserved. Both type 1 and 2 inclusions are present however, they are generally $<5\ \mu\text{m}$ across whereas those in quartz are typically $<10\ \mu\text{m}$ across. Cassiterite contains abundant red solid inclusions, similar to Kow Daeng, which are presumably Nb-Ta minerals.

CHAT CHAI

Chat Chai is a stockwork cassiterite deposit located 150 km west of Bangkok in Ratburi province (Chapter 7; Fig. 1). Cassiterite mineralization is associated with biotite and leucocratic granite which intruded into psammitic to pelitic metasedimentary rocks. The bulk of the mineralization occurs in quartz veins which cut the biotite and leucocratic granite close to the metasedimentary contact. The dominant alteration type is greisen, composed of quartz, muscovite, fluorite, and rarely topaz. Sulphides, in particular pyrite, are also common in the greisen. Greisen alteration commonly forms alteration haloes around quartz and quartz-

cassiterite veins and less commonly is pervasive. Minor early tourmaline alteration is present locally. Cassiterite in quartz veins is typically intergrown with muscovite. Adjacent to the deposit, cassiterite is also present in massive fluorite and in quartz-pyrite-arsenopyrite-galena veins.

FLUID INCLUSION PETROGRAPHY

Four fluid inclusion types have been recognized in cassiterite, quartz, and fluorite. Type 1 are L-V inclusions. Type 2 are aqueous-carbonic L-L-V inclusions. Type 3 are liquid only L inclusions. Type 4 are vapour-rich V, V-L, and V-L-L. The fluid inclusions generally have elliptical shape and are very small, generally less than 10 μm across. All four fluid inclusion types are present in quartz. Inclusion type 1 is the most common. Most of the inclusions in quartz occur along secondary planes. Two fluid inclusion types occur in fluorite, types 1 and 3. Type 1 is the dominant inclusion type. Most inclusions in fluorite are secondary. Inclusion types 1 and 3 both commonly occur along the same secondary inclusion planes suggesting that some type 1 inclusions are very low temperature. Two fluid inclusion types have been observed in cassiterite, types 1 and 4. Both inclusion types are common and they are distributed throughout cassiterite grains without any apparent planar orientation.

HUB KAPONG

The Hub Kapong tin deposits are located approximately 140 km southwest of Bangkok (Chapter 7, Fig. 1). There are several former tin producing deposits in this, but none are currently in production. The only primary deposit, near Ban Tha Lao, was mined for cassiterite from quartz veins. The Ban Tha Lao deposit is hosted by tourmaline-muscovite granite within 20 m of the contact with pelitic metasediments. At least three veins were mined, all of which have the same general orientation. Only one vein could be measured and is 20 to 50 cm thick and 50 m long with a strike and dip of 072/80°S. The veins are composed of quartz and tourmaline with accessory cassiterite, arsenopyrite, pyrite, muscovite, and trace scheelite and molybdenite. Diffuse tourmaline alteration haloes occur around the quartz veins which locally are overprinted by muscovite (greisen) alteration. No feldspathic alteration was observed although none of the samples have been stained for K-feldspar. The cassiterite and tourmaline are intergrown and apparently coeval whereas muscovite is later, replacing tourmaline and occurs along fractures in cassiterite. Very few intergrowths of scheelite or sulphides with cassiterite or tourmaline were observed, however, tungsten and base metal mineralization appears to be contemporaneous with that of tin.

FLUID INCLUSION PETROGRAPHY

Four types of fluid inclusions have been identified in quartz, cassiterite, tourmaline, and scheelite. Type 1 inclusions are L-V inclusions. Type 2 are L-L-V.

Type 3 are vapour-rich V, V-L, and V-L-L inclusions, and type 4 are liquid only L. Most fluid inclusions have spherical or ellipsoidal shapes although some irregularly shaped inclusions occur in cassiterite and tourmaline, and are generally <10 μm across in quartz and <5 μm across in the other minerals.

The fluid inclusions in cassiterite are poorly developed and/or preserved. Type 1 are the most common and type 3 less common. Both types generally are isolated in small zones which contain several inclusions. Less commonly, type 1 inclusions occur along short planes <100 μm long and thus may either be secondary or pseudosecondary.

All four types of fluid inclusions are observed in quartz. Each of the four types has been observed along secondary planes. Therefore most if not all the fluid inclusions are secondary. Fluid inclusions are abundant, have spherical or ellipsoidal shapes. Type 1 inclusions are the most common, type 3 are also fairly common, type 4 are less common, and type 2 are rare. In inclusion types 2 and 3 coexisting carbonic liquid and vapour are present at 25°C indicating that at least some of these inclusions are CO₂-rich (high Th CO₂).

In tourmaline type 1 inclusions are common, type 3 inclusions are less common, and type 2 inclusions are rare. The distribution of fluid inclusions is similar to that in cassiterite although short planes of secondary or pseudosecondary type 1 inclusions is more common.

Very few grains of scheelite were observed in thin sections and the fluid

inclusions in these grains are poorly developed and/or preserved. Type 1 are the most common and type 3 are less common.

TAP SAKAE

The Tap Sakae mining district is located approximately 300 km south of Bangkok (Chapter 7, Fig. 1). The tin mines in this area are presently closed, but historically significant tin and, as a by-product tantalum, was produced. The Thai Prasit deposit contains primary mineralization and was therefore sampled. Cassiterite in this deposit is hosted by "pegmatite" dykes which crosscut psammitic metasediment occurring at a distance <5 km from the contact between the metasediments and a batholith of the western Thailand granite belt. Approximately 2 % Ta (plus Nb?) was recovered from the cassiterite ore at this deposit. Most of the cassiterite is disseminated in "pegmatites", in particular quartz-muscovite "pegmatite" dykes that are <1 m thick and crosscut metasedimentary rocks. The metasedimentary rocks are biotite grade, have psammitic compositions, and are similar to those at Nong Sua. Cassiterite is disseminated in dykes that are composed of quartz and muscovite and less commonly quartz-muscovite-feldspar. The composition of these dykes suggest that they may not be entirely magmatic in origin however, no alteration haloes in the surrounding metasediments are present. A second type of pegmatite, composed of quartz, feldspar, muscovite, garnet, and tourmaline also occur as dykes which crosscut metasedimentary rocks but these are

barren of cassiterite.

FLUID INCLUSION PETROGRAPHY

Two types of fluid inclusions are present in quartz and cassiterite. Type 1 are L-V and type 2 are vapour-rich V and V-L. Fluid inclusions in either quartz or cassiterite are poorly developed and/or preserved, typically have spherical or ellipsoidal shapes, and are $<5\text{ }\mu\text{m}$ across.

Only a few fluid inclusions were observed in cassiterite and both types are present as isolated inclusions. Both inclusion types are also observed in quartz but they generally occur along secondary planes. Coexisting carbonic vapour and liquid are present at 25°C indicating that at least some type 2 inclusions are CO_2 -rich.