PROGRESSIVE ALBITISATION IN THE "MIGMATITE CREEK" REGION, WEEKEROO INLIER, CURNAMONA

Thesis submitted for the degree of
Master of Science
Geology &Geophysics
School of Earth & Environmental Sciences
The University of Adelaide

by

CHENG LIN YANG

July 2008

4   INTERPRETATION

In order to generate an accurate account of progressive albitisation during the geological evolution, it is necessary to understand and determine the enrichment/removal of elements that the rocks have been subjected to. This requires the accurate quantification of the mobility of elements associated with the changes in mineral assemblage phases during albitisation. The quantifications rely on mass balance estimates in rock types and minerals based on whole rock analyses, EMA and LAICPMS. The changes of mineral assemblages were recorded on the albitisation map (Figure 3.1) associated with the observation of field and microscope petrography. Further, the chemical equilibria of minerals were used to infer variation of temperature e.g. by using geothermometers of feldspars (Kiryukhin et al. 2002; McCaig et al. 1990; Brown & Parsons 1981), minor and accessory minerals (Strauss 2003; Marshall et al. 2006; Niiranen et al. 2005; John & Stormer 1975), e.g. feldspar–amphibole, albite–magnetite, albite–biotite and biotite-garnet (Strauss 2003; Enami et al. 1994., Niiranen et al. 2005). Mobility of REE elements was found to be related to the intensity of albitisation.

Mapping in the Migmatite Creek study area produced
- a lithological map distinguishing initial lithologies,
- a thematic map of albitisation intensity distinguishing low, medium and high intensity albitisation in the different lithologies based on visual inspection. The boundaries were set as follows:
- Low = 0-45% albite (LA),
- Medium = 45 to 65% albite (MA),
- High = >65% albite in visible albitisation (HA).

The mapping approach of albitisation was easy applied in the field. The mapping exposed or revealed the relationship of albitisation to structures and lithologies. Albitisation is structurally controlled by antiforms and breccias together. HA, MA and LA are related to network of breccias, single direction group of breccia pipes and unbrecciaed rock. The criteria provided the insight of albitisation in different genetic lithologies. For example, Low intensity albitisation of amphibolite and complete albitisation of psammopelite are present in MA. 20-30% pegmatite is albitised clearly but most of it is not. Pegmatites have extreme variation of chemical compositions. Thus, the intensity of albitisation is also influenced partly by the different genetic rock.

4.1   Changes in Rock compositions

Lithologies in the Migmatite Creek area are characterised by high Na and low K and Ca contents, as evident from whole rock analysis (see Figure 3.17). Determination and comparison of bulk rock compositions allow inferring the compositional changes during an alteration event and providing insight into the character and compositional variation of the fluids involved during alteration. Applying mass balance calculations allows quantification of the changes the fluids and rock underwent during an alteration event.

4.1.1 Variation of Na₂O, K₂O and CaO

4.1.1 Variation of Na_2O, K_2O and CaO

The most prominent alteration in the studied area is a pervasive sodic alteration or albitisation affecting all lithologies to variable degrees. As this albitisation primarily affects feldspars in the respective lithologies it is best expressed as the change in the ratios of Na_2O to K_2O and CaO. In most rocks, the Na_2O/K_2O and Na_2O/CaO ratios decrease with progressing albitisation (see Chapter 3.5 and Figure 4.2). This results in a relative Na_2O enrichment as evident from whole rock analyses.

The fluid causing this albitisation has been shown to be Na-rich (Lu & Lottermoser 1997), and as a consequence of the formation of albite the break down of other feldspars can be readily assumed to change to a more K + Ca rich composition in the process. This spent fluid may in turn cause secondary alteration, observed in this study as a localised biotite alteration, and also the formation of Fe-Mg or calc silicate matrix breccias.

The ratios of Na_2O/K_2O are (as the mean of 8 samples) the lowest 1.56 in unaltered/altered pegmatite. This compares to the high Na_2O/K_2O ratio of metasediment and relates to an increase in K in the pegmatites. The ratio of Na_2O/K_2O in unaltered/altered pegmatites is about 1.0 in sample Op031 and about 2.4 in sample Bp032 respectively. These were different from the ratios of 3-53 in metasediments (see Chapter 4.2.1).

The concentrations of Na_2O and Al_2O_3 were taken to demonstrate the effect of albitisation in all rock types. Figure 4.1 shows that the concentrations of Al_2O_3 strongly correlate with the concentrations of Na_2O. The albitised rock has the highest concentration of Na_2O and Al_2O_3 in the distinguishing domains of rock types (Figure 4.1). This is largely effect of albite firstly replacing feldspars and subsequently most other minerals, e.g. micas, amphiboles and to some extent quartz.

Al is one of immobile elements possibly during low intensity albitisation and biotite alteration (Chapter 2.5) but Al in fluids was fixed into rock during medium–high intensity albitisation. Concentrations of Na_2O+K_2O+CaO have increased with Al_2O_3 concentration with a stable zone of plotted points for each rock type (Figure 4.1). The narrow compositional variation and near linear distribution reflects that the rock types have evolved from a complex to a restricted, Na_2O dominated composition during progressive albitisation, i.e. original minerals assemblage replaced by albite.
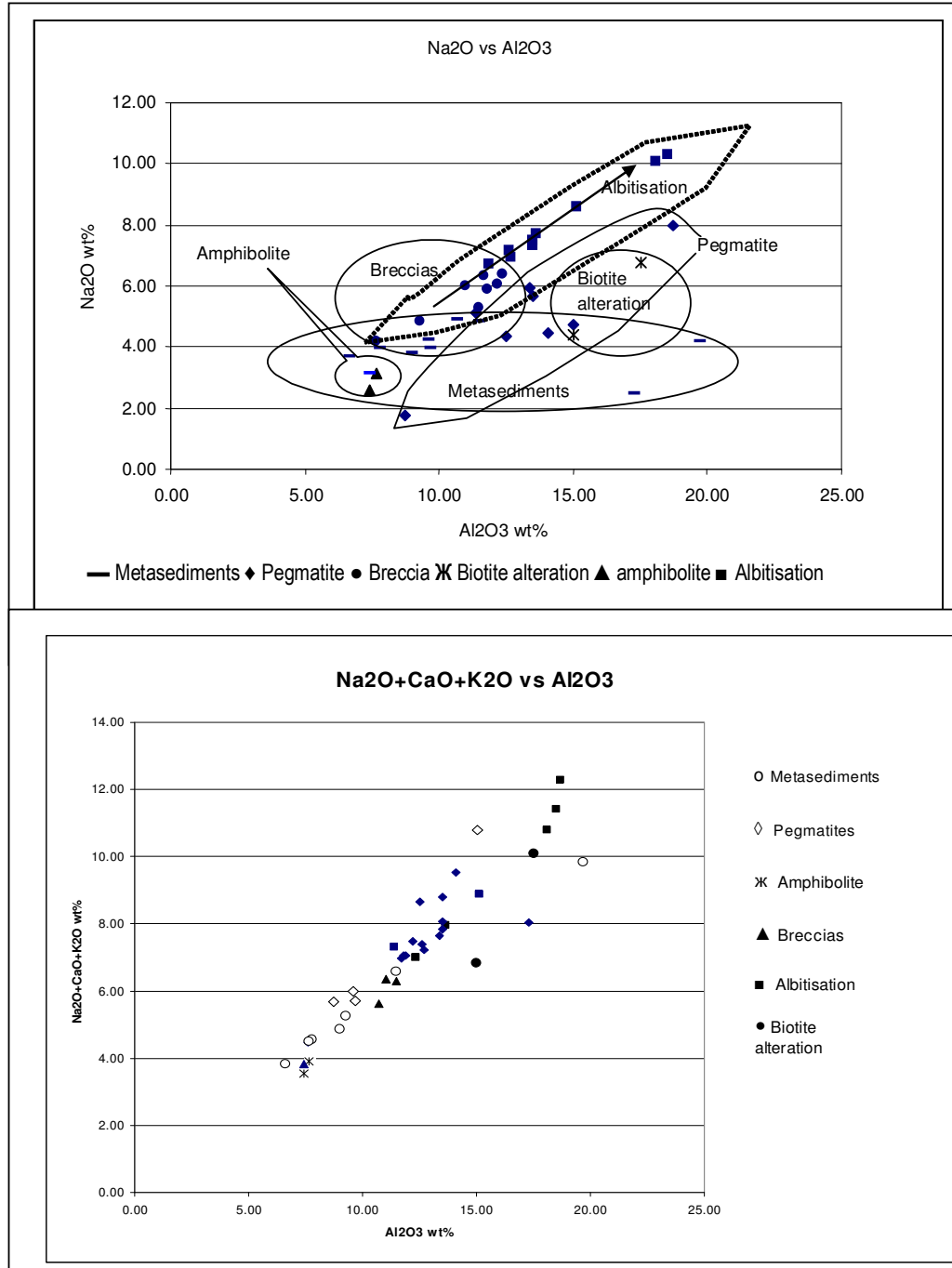
Figure 4.1 Na$_2$O versus Al$_2$O$_3$ and ratios of (Na$_2$O + K$_2$O + CaO)/Al$_2$O$_3$ in 41 samples.
Notes: 1.Concentrations of Na$_2$O and Al$_2$O$_3$ in all types of rock on the top. 2. The ratios of (Na$_2$O + K$_2$O + CaO)/Al$_2$O$_3$ plot in rock types at the bottom.
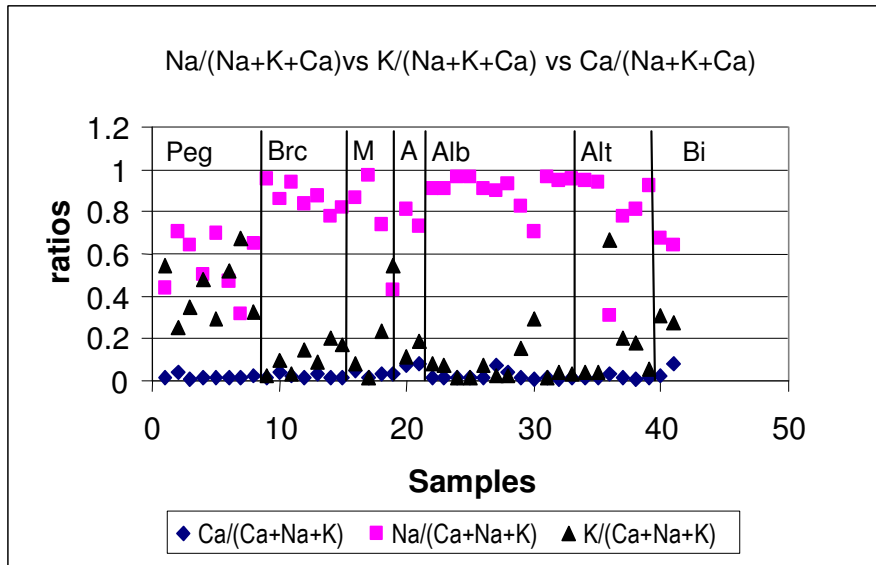
Figure 4.2 The ratios of Na_2O, K_2O and CaO in all rock types.
Notes: Three ratios, $Na_2O/(Na_2O+K_2O+CaO)$, $K_2O/(Na_2O+K_2O+CaO)$ and $CaO/(Na_2O+K_2O+CaO)$, are present here in 7 units, pegmatite (Peg), breccias (Brc), metasediments (M), amphibolite (A), albitisation (Alb), alteration (Alt) and biotite alteration (Bi). This is a modifying of figure 3.17.

4.1.2 Evolution of compositions

Quantification of the compositional changes during albitisation can be done by comparison of the compositions of equivalent unaltered and altered rocks using isocon based on approach (Grant 1986, Baumgartner & Olsen 1995). This approach was applied to:

1) unalbitised and albitised metasediments,
2) breccias and unalbitised metasediments,
3) biotite alteration and unalbitised metasediments,
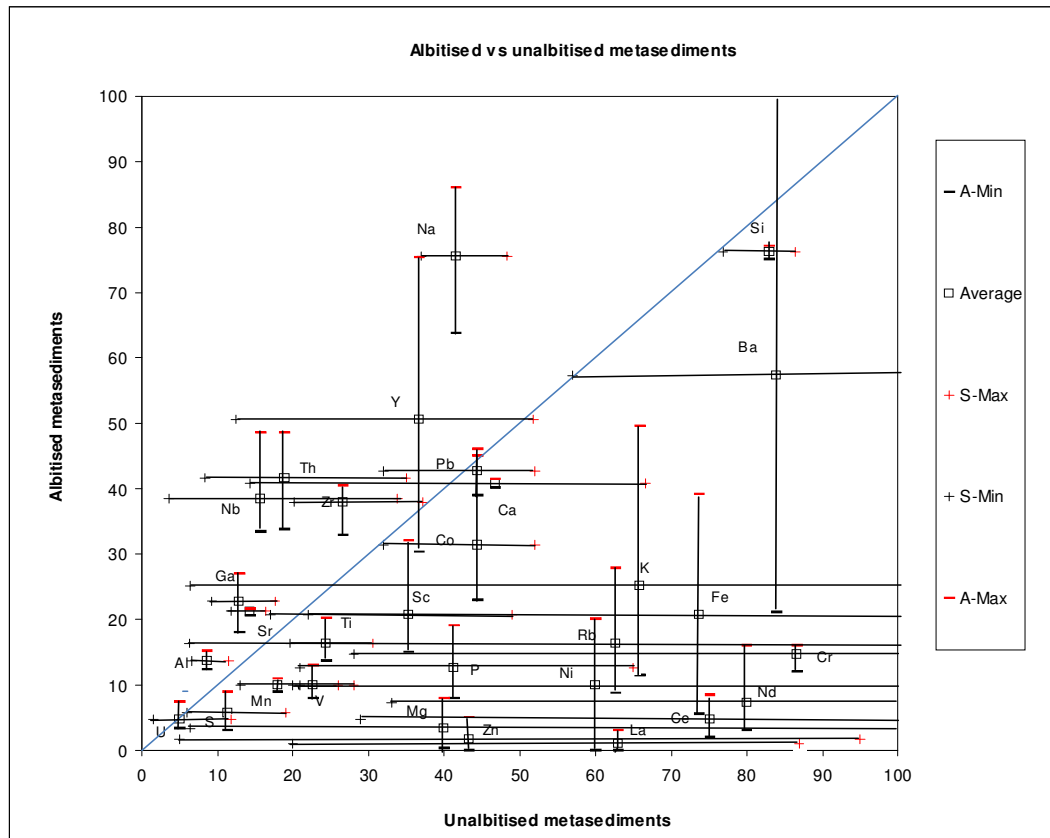4) unalbitised and albitised pegmatites.

Isocon mass balance estimates were used to determine and partly quantify enrichment/depletion of elements during alteration. Evolution of rock composition demonstrated through pegmatites (migmatisation), albitisation and biotite alteration in metasediments. The change during migmatisation has been considered as possible part of composition of albitisation fluids based on a study of the inclusions in pegmatite in the area (Lu & Lottermoser 1997).

4.1.3 Quantification of compositional changes between unaltered and altered rock – the isocon approach

4.1.3.1 Unalbitised and albitised metasediments

Six samples were selected to represent the unalbitised metasediments (samples L003, Uap007 and Ps008) and the albitised metasediments (samples Ab004, Mb006 and Ab010), see Figure 4.3. Following this, the bulk chemical and volumetric changes were

quantified during albitisation and conclusions drawn about the compositional changes of the involved fluid phases based on mean values of each set.

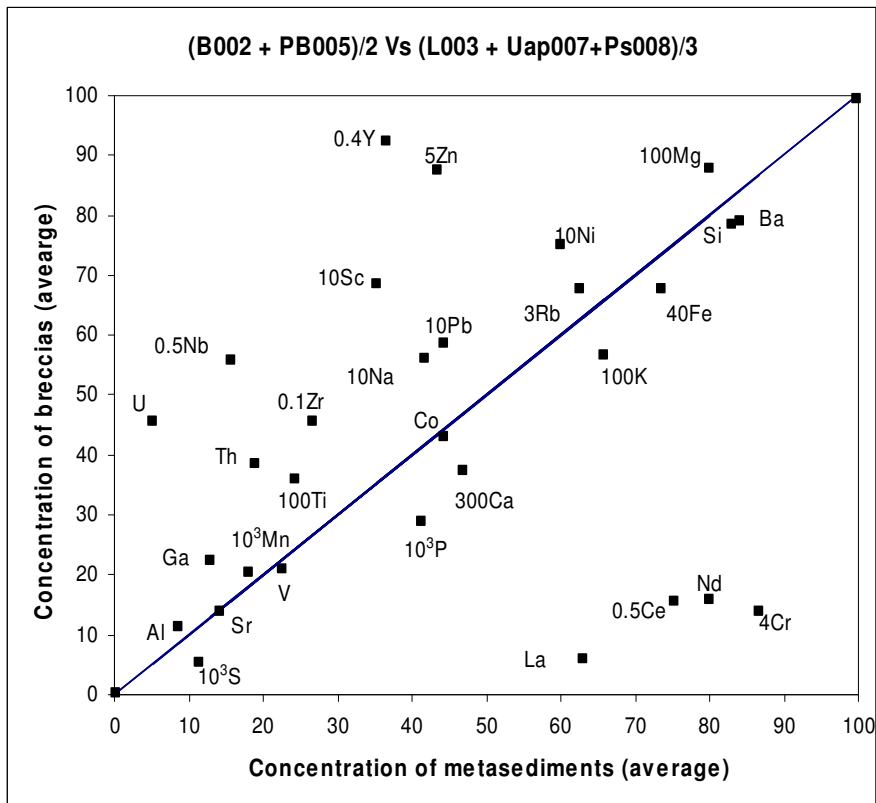**Figure 4.3** Element concentration changes between unalbitised and albitised metasediments. 1. Element concentration (major and trace element are respective in oxidised (Fe_2O_3) and metal (Ni) forms) is arbitrarily scaled to aid in plotting. 2. Al, Ga, Zr and Y have the same behavior on average (see text). 3. K_2O, Fe_2O_3, MgO, Ni, Ba, Rb, Cr, Nd, and Ce has the range of arbitrarily scaled maximum 100.5 (Ce) -154 (K_2O) in X-axies. Ba has the arbitrarily scaled maximum 125 in Y-axies (see appendix B).

4.1.3.2 Breccias and metasediments

Two samples (B002 and PB005) were taken to represent the breccias and three samples (L003, Uap007 and Ps008) to represent unalbitised metasediments, to determine changes of compositions between breccias and unalbitised metasediments. Figure 4.4 show that the samples demonstrate an increase of Na_2O – U – Nb – Th – Sc – Zn – Pb in breccias but an extreme decrease of SO_4 – P_2O_5 - K_2O – CaO – La – Cr – Nd – Ce elements compared to metasediments. The changes in composition involved three processes: albitisation in clasts, brittle deformation (T-P changes) and mafic alteration in matrix. The behavior of the trace elements, U, Th, Zn and Pb are different from that of the REE, La, Nd and Ce, as well as Cr in re-mobilisation.

4.1.3.3 Unalbitised and albitised pegmatites

Two samples of each type, (Pm017 and Op031 for unalbitised pegmatite and ABP022 and AP041 for albitised pegmatite) were taken to represent compositional changes during albitisation of pegmatites. (Figure 4.5) The isocon plot shows a relative increase of Na_2O - TiO_2 - Zr - Zn - Nb - Ga ± Cu and a decrease of Fe_2O_3 - MgO - U - La - Ce - Nd - Pr - Sm - Pb - W as an effect of the albitisation.

Figure 4.4 Composition of breccias compared to metasediments on average. Element concentration is 0.1-10³ scaled to aid in plotting, e.g. 10Na (10×Na$_2$O wt%) and 0.1 Zr (0.1×Zr ppm).
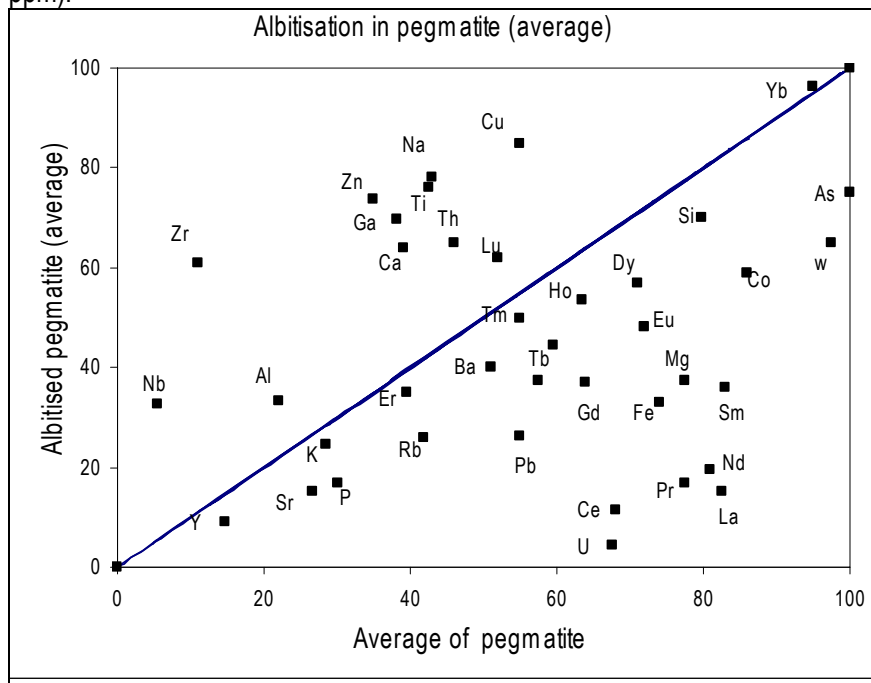
Figure 4.5 Composition changes between unalbitised (Pm017+Op031) and albitised pegmatites (ABP022 + Ap041). Element concentration is arbitrarily scaled to aid in plotting.

4.1.4 Enrichment/removal of elements

4.1.4.1 Albitisation

Six samples were taken to evaluate the chemical changes caused by albitisation; the results are summarised in Table 8. Three pairs of albitised and unalbitised samples are, respectively, AB004 vs L003 (Figures 4.6); Mb006 vs Uap007 (Figure 4.7) and Ab010 vs Ps008 (Figure 4.8). These were taken to represent the albitised metasediments and their unalbitised precursors in zones of three intensities, to demonstrate the effects of progressive albitisation. Based on observation and selected statistic results (Grant 1986; Baumgartner & Olsen 1995), isocons were fitted for immobile elements:
　　　　- Al, Ti, Zr, Nb and Ga for low intensity albitisation,
　　　　- Si, Ti and Sc for medium and high intensity albitisation.
The parameters and results are summarized in Table 8. Figure 4.6 shows that the samples demonstrate during low intensity albitisation relative losses of major elements K_2O, CaO and TiO_2; trace elements Zn, Ni, Nd, Ce, Rb, Cr, U and V, and gains of Na_2O, Y, Co and Sr. Figure 4.7 demonstrates relative losses during medium intensity albitisation of major elements Fe_2O_3, MgO and P_2O_5 as well as trace elements Zn, Co, Ni, La, Ce, Nd, Cr, Ba and V and gains of major elements Na_2O, Al_2O_3, CaO and K_2O as well as trace elements Nb, Th, U, Ga, Y, Zr and Sr. Figure 4.8 demonstrates relative losses during high intensity albitisation of major elements Fe_2O_3, MgO, K_2O, P_2O_5 and SO_3; trace elements Zn, Ni, La, Ce, Nd, Cr, Co, Ba, Rb and V, gains of major elements Na_2O, Al_2O_3; trace elements Nb, Th, U, Ga and Zr.

　　　　Two samples were taken to represent the unalbitised (Ps034) and albitised (Ps035) pegmatite. The alkaline elements, Na_2O, Al_2O_3, P2O5, CaO, K_2O, Sr and Ba, were enriched in the albitised pegmatite, while Fe_2O_3, MnO, MgO and TiO_2 as well as the trace elements Ni, Cr, Zr, Nd, Cs, Pr, Sm, Sc, Th, U, Gd, V and Tm were depleted (Figure 4.9). Moderately mobile elements were Eu, Mo, Cu, Co, Dy and Nb.
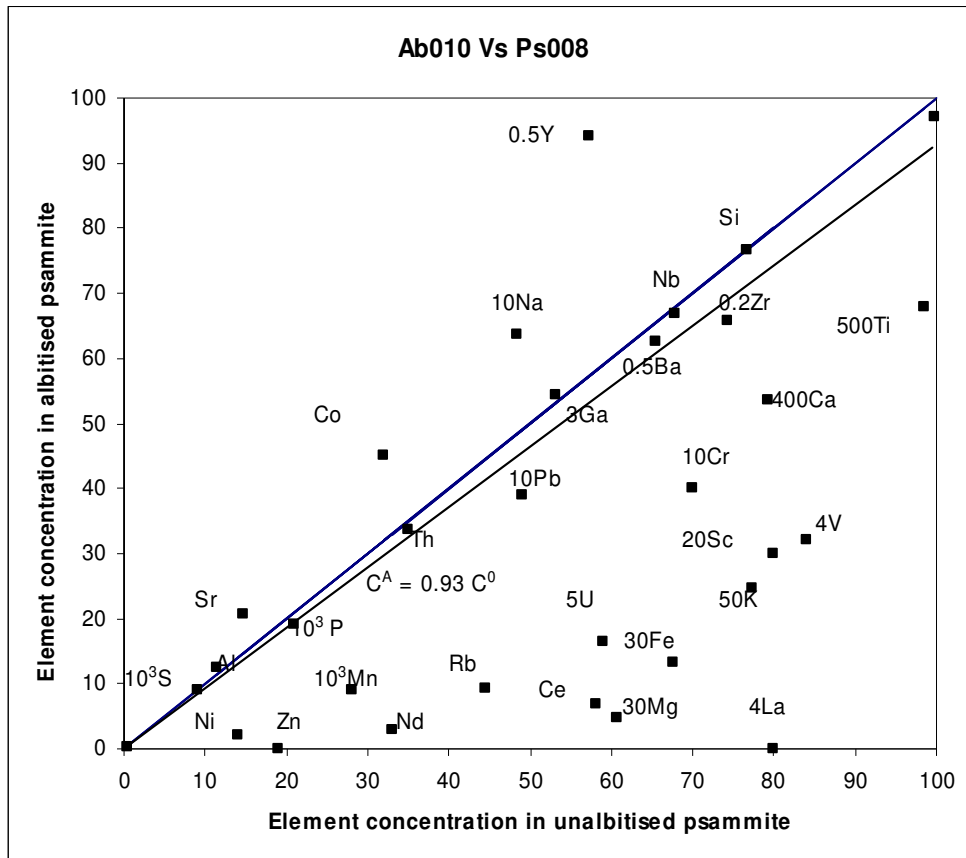
4.1.4.2 Biotite alteration

Samples Ps008 and Bi009 were taken to represent the biotite alteration in psammite. An isocon was fitted using Al, Ti, Nb, Y and Zr as the immobile element group (Grant 1986; Baumgartner & Olsen 1995). The parameters and results are summarized in Table 8. Figure 4.10 shows that the samples demonstrate relative losses of major elements SiO_2, P_2O_5 and CaO; trace elements Nd, La, Ce, Cr, Co and U and gains of major elements Fe_2O_3, MgO, K_2O and MnO; trace elements Zn, Sc, Ga and Rb during the biotite alteration.

Table 8 Summary of parameters and results for the alterations in the Migmatite Creek.

Samples	Alteration	Immobile elements	Iscons	Figures	Enrichment	Removal
Ab010 vs ps008	Albitisation LA	Al, Ti, Zr, Nb, Ga	Slope 0.9323, 7.26% of volume decrease	Figure 4.6	Na_2O, Y, Co and Sr	K_2O, Fe_2O_3, MgO, MnO and CaO; Zn, Nd, Ce, Rb, U and V
Mb006 Vs Uap007	Albitisation MA	Si, Ti, Sc	Slope 0.8113, 23.25% of volume decrease	Figure 4.7	Na_2O, Al_2O_3, CaO and K_2O; Nb, Th, U, Ga, Y, Zr	Fe_2O_3, MgO and P_2O_5; Zn, La, Ce, Nd Cr, Ba and V
Ab004 vs L003	Albitisation HA	Si, Ti, Sc	Slope 0.7298; 37.24% of volume decrease	Figure 4.8	Na_2O, Al_2O_3; Nb, Th, U, Ga and Zr	Fe_2O_3, MgO, K_2O, P_2O_5 and SO_3; Zn, La, Ce, Nd Cr, Ba, Rb and V
Bi009 vs ps008	Biotite alteration	Al, Ti, Nb, Y and Zr	Slope 1.388, 27.95% of volume increase	Figure 4.10	Fe_2O_3, MgO, K_2O and MnO; Zn, Sc, Ga and Rb	SiO_2, P_2O_5 and CaO; Nd, La, Ce, Co and U

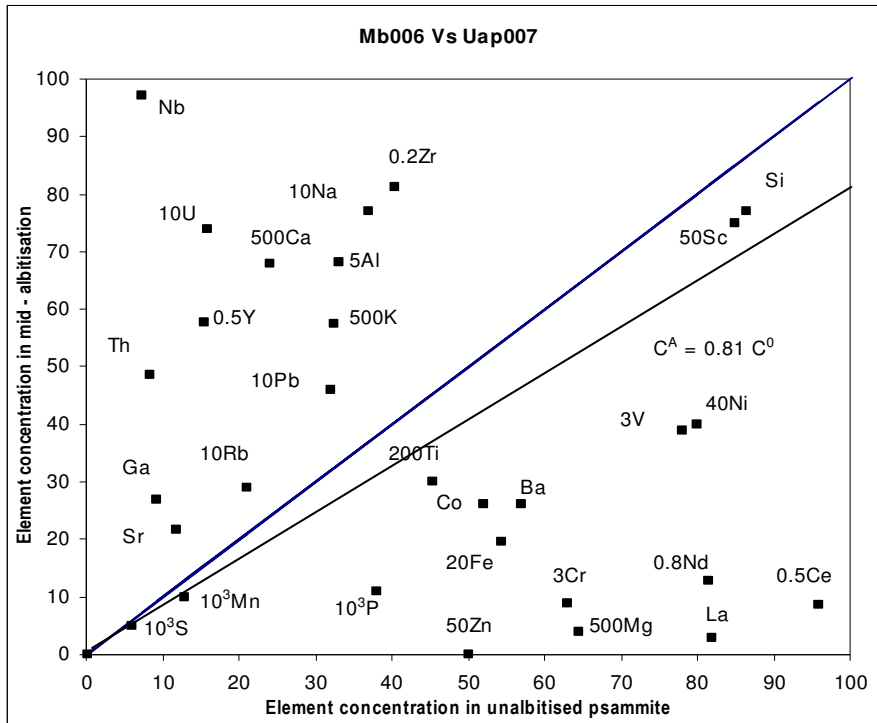**Figure 4.6** Low intensity albitisation of psammite.

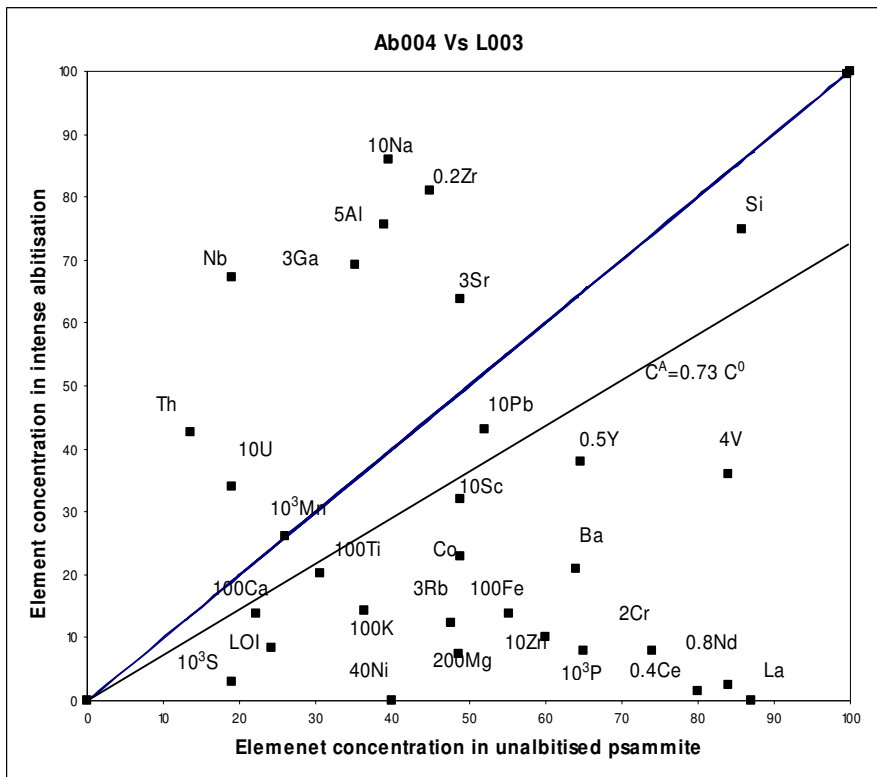**Figure 4.7** Medium intensity albitisation of psammite.

Figure 4.8 High intensity albitisation of psammite.
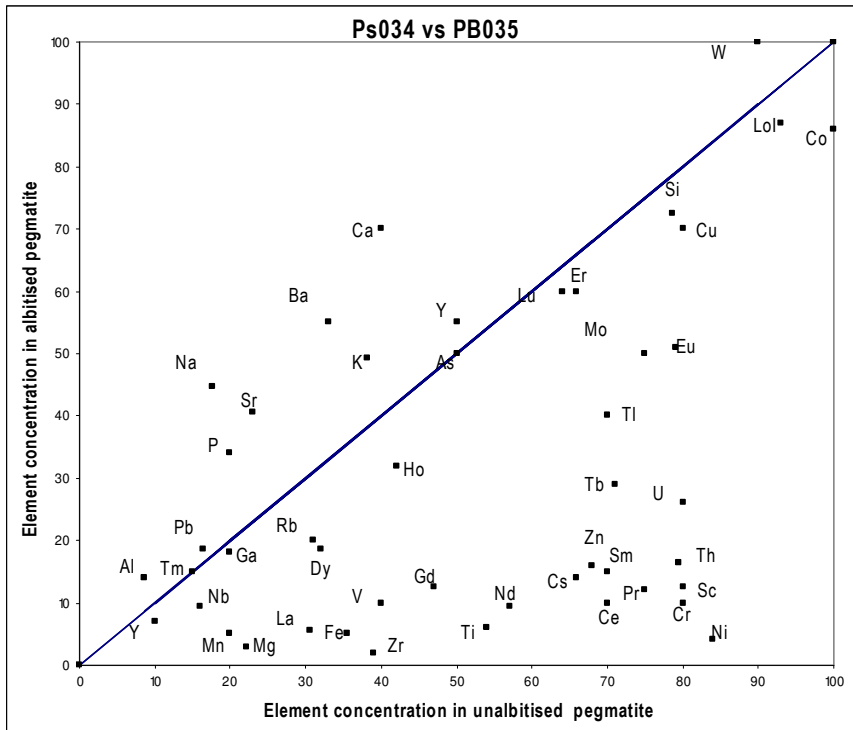
Figure 4.9 Albitisation of pegmatite.
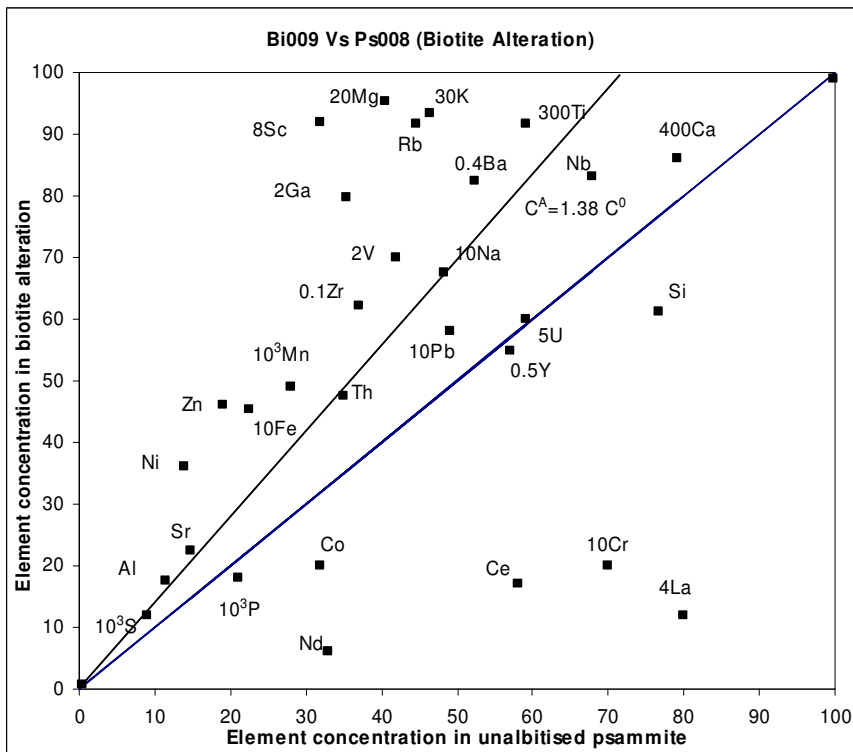
Figure 4.10 Biotite alteration in psammite.

4.1.5 Bulk trace elements variation

Ni, Ba and Rb have similar patterns of depletion or enrichment as the K_2O concentrations in 31 samples of rock types (Figure 4.11). Fourteen REE elements and Y were depleted during albitisation in the variable range of rock sequence (Figure 4.12).
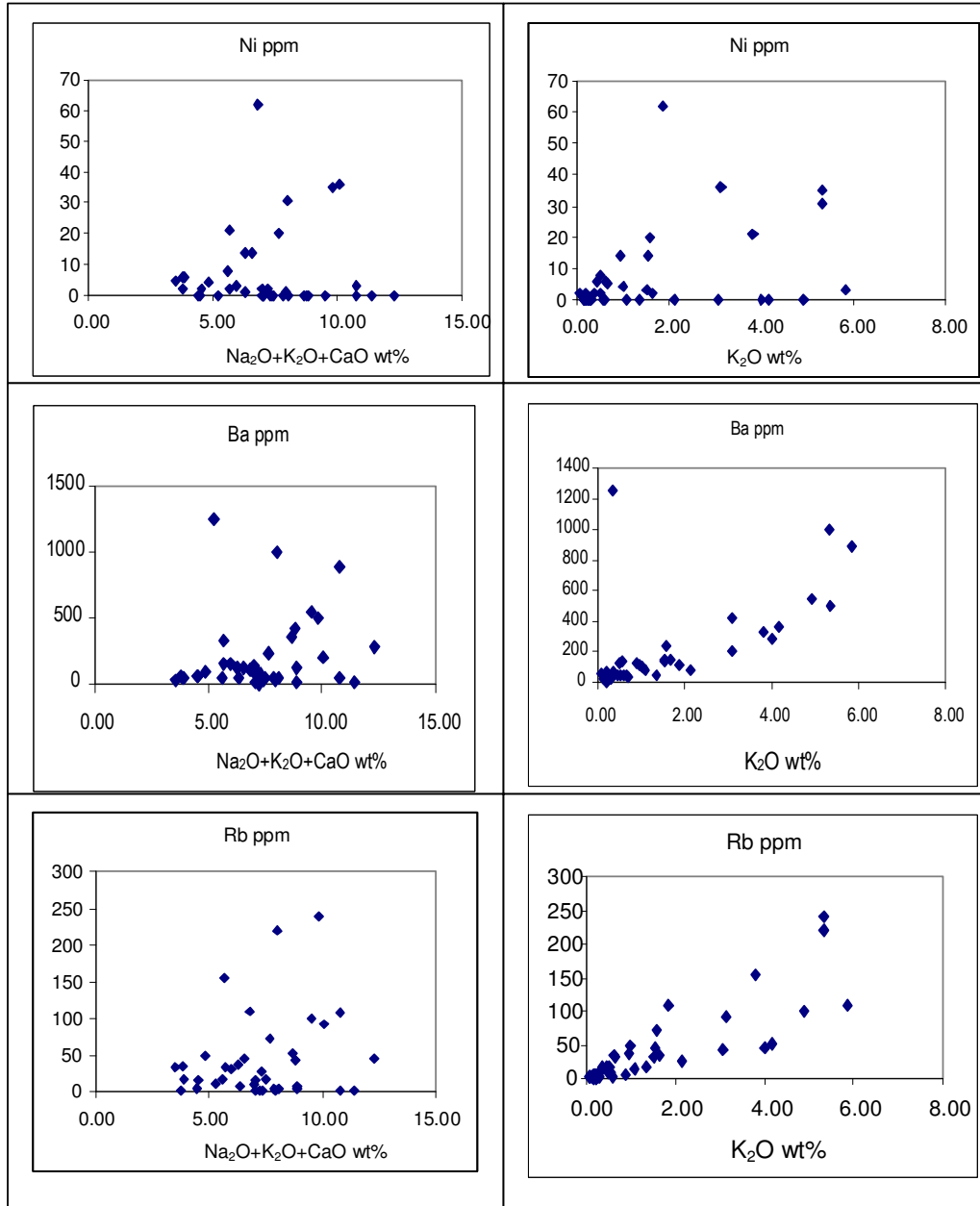
Figure 4.11 Concentration of Ni, Ba and Rb vs Na_2O+K_2O+CaO as well as K_2O in all samples. Na_2O+K_2O+CaO and K_2O are at wt% in the columns. Y-axis = Ni, Ba and Rb at ppm.
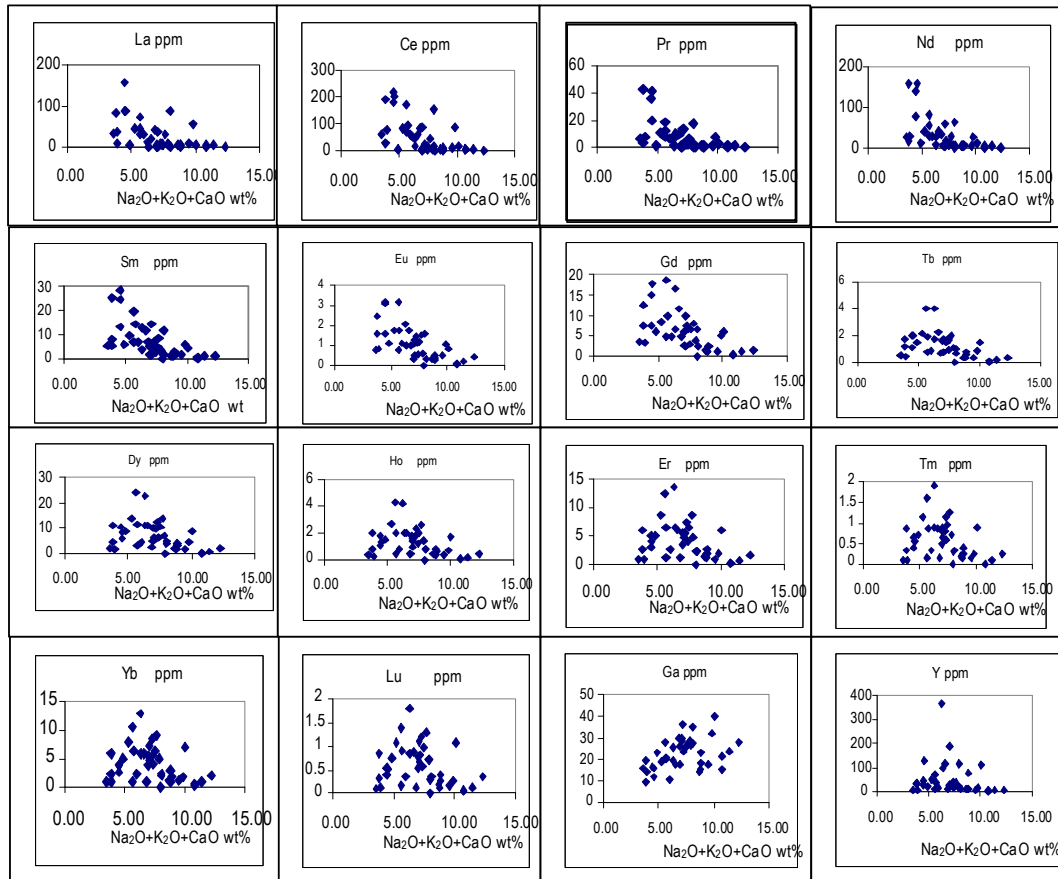
Figure 4.12 REE (La to Lu), Ga and Y vs Na_2O+K_2O+CaO in 31 samples.
Notes: 1. Decrease of light REE concentrations is more apparent than heavy REE during increase of $Na_2O + K_2O + CaO$ wt% in 31 samples of rock types. 2. Ga is opposite to the characteristics of light REE and Y is similar to heavy REE in the 31 samples.

4.1.6 Fluid evolution during albitisation and related alteration

The changes in the compositions of fluids were estimated during albitisation using the isocon based mass balance estimates (Chapters 1.4.4 and 2.5). An average composition in albitisation fluid was rich Na, Th and Nb as shown in Figure 4.3 (changes of Na_2O from 4.2 to 7.6 wt%, Th from 18.9 to 41.7 ppm and Nb from 31.4 to 77 ppm). The element group of Al, Ga, Zr and Y and additional Sr can be interpreted to characterise the albitisation fluids. The three isocons giving details of progressive albitisation support the finding that the albitising fluids contained elevated-amounts of Na and minor amount of Al, Ga, Zr, Y and Sr. The additional evidence was that densities (ρ^O) of regional Proterozoic metasediments were in the range of 2.65-2.72 and albitised metasediment (ρ^A) in the range of 2.60-2.65 (Clark et al. 2005; Oliver et al. 2004). The porosities of all metasediments were consistent without obvious changes during albitisation. The fluid composition changed continuously during albitisation, exchanging the predominant Na, minor Th and Nb and the accessory group of Al, Ga, Zr, Y and Sr, for Mg, Fe, K; minor Zn, Ni, Cr, Ba, P and the REE elements. The changes of rock compositions imply that the average effects of albitisation may not be

appropriate for mass balance estimates. The evolution of fluid composition gradually resulted in metal rich fluids.

A large hypersaline source reservoir may be a good interpretation for evolution of compositions in the Migmatite Creek region. The regional metamorphism may provide a main albitisation fluid source (Clark et al. 2005, 2006). Migmatism may also produce minor part of albitisation fluid source as shown by Na-CO_2 rich inclusions in pegmatite (Lu & Lottermoser 1997).

The fluids involved in the brecciation process may have been derived from albitisation. Thus, the evolved fluid lead to mafic character of the breccia matrix and possibly produced the K-rich biotite alteration in suitable lithological and structural settings.

Fluid compositions of high intensity albitisation were Na-rich and hypersaline. During medium intensity albitisation, the K_2O contents of fluids were not greatly increased at low stable concentration of K_2O 0.07 wt% in sample Uap007 of psammite and 0.12 wt% in Mb006 of albitised psammite. However, the increase of K_2O was clearly in high and low intensity albitisation as demonstrated by K_2O of 0.36 to 0.14 wt% (L003 to Ab004) and 1.55 to 0.50 wt% (Ps008 to Ab010) from psammites to albitised psammites, respectively. The evolution of fluid compositions partly depended on the primary metasediments.

Mobility of Si was slight although concentrations were changed from SiO_2 85.90-86.45 wt% to SiO_2 74.99-77.01 wt% (Decrease of SiO_2 10 wt% was by brecciation mechanism - decrease of density and increase of porosity) during medium and high intensity albitisation (see mass balance estimates-Iscons). Yet the mobility of Si did not occur at low intensity albitisation with a quite stable range of 76.73-76.78 wt% SiO_2. However, mobility of Si was identified by change of concentrations from 76.68 to 61.20 wt% SiO_2 during biotite alteration.

Whole rock analyses thus allow us to determine the bulk composition of the alteration fluids. This can be further substantiated by including the detailed results from individual mineral analysis (see following 4.2.1.5).

4.1.7 Timing estimates

The OD_3 (1580Ma) structures in the Weekeroo Inliers are represented by networks of breccia pipes in both N-S and SW-NE (60 degree) directions. The networks were formed at or just after the formation of the folds as indicated by the apparent control of albitisation intensity by hinges.

The estimated timing of albitisation could be contemporaneous with breccias at the end of OD_3. The formation of the breccia matrix reflects the end of evolution of composition of albitisation fluids. Thus, albitisation occurred at or after the formation of the antiformal folds.

4.2 Mineralogy, Geothermometers and Chemistry

4.2.1 Minerals (Feldspar and Micas)

Most feldspar in the analysed samples of psammite and amphibolite as well as in their albitised equivalents is the Na-rich end member albite. However, the new albite in albitised psammite contains the lower Na/K ratios and the lower contents of Al_2O_3, MgO, Fe_2O_3, TiO_2 and BaO than primary albite in unalbitised psammite (Figure 4.13). In contrast, the new albite contains more SiO_2 contents than the primary albite does (Figure 4.13). Biotite in albitised psammite also contains more K and Ca. The changes in albite and biotite demonstrate that the new albite-rich assemblage is completely different from the quartz-biotite–albite assemblage. This indicates new geochemical equilibria of fluid/rock system. A high albitisation temperature is inferred by the changes, compared to metamorphism.

$Ab_{90.0-97.4}$ in the biotite alteration of pegmatite contains more K (Ca) than $Ab_{93.4-98.0}$ in unaltered pegmatite does. An exception was $Or_{2.1}An_{14.5}$ in unaltered pegmatite which was associated with variable unaltered pegmatite. Transitional compositions between albite and orthoclase were not found except in the least albitised amphibolite. Albite $_{(An_{4.0-12.4})}$ in the least albitised amphibolite indicates micro-albitisation.

K-feldspar was identified only in pegmatites. $Or_{93.9-98.8}$ in the biotite alteration of pegmatite indicates slight depletion of K in comparison to $Or_{98.2–99.2}$ in unaltered pegmatite. $Or_{86.1}Ab_{11.8}An_{2.1}$ also indicates biotite or Fe-alteration of pegmatite (Bp32) depleted K. The distinguishing behaviours of albite and orthoclase may result from equilibria of fluid/rock in pegmatite during biotite alteration or Fe-alteration.

The distribution of compositions of feldspar is inferred using both the system equilibria and T-P evolution. The first interpretation is that there was an uncommonly low content of Ca and K in the whole system of fluid/rock. An alternative interpretation suggests that anorthite can not have formed below temperatures of 550 °C. Minor oligoclase may have been formed at a peak temperature of magmatism (amphibolite) / migmatization (pegmatites) rather than albitisation. Albitisation forms at lower temperatures. The preferred interpretation is that these feldspars reflect exchange of elements in minerals in circulation of fluid flow and temperature was a minor additional factor.

4.2.1.1 Ratios of Na_2O/K_2O, SiO_2 and Al_2O_3 in albite

The ratios of Na_2O/K_2O and Al_2O_3 concentrations of albite in albitised psammite are commonly lower than in unalbitised psammite (Figure 4.13). This indicates that newly formed albite contain less Na and Al than primary albite phases. Al behaviour in low intensity albitisation is different from it in medium-high intensity albitisation. The changes of average Na_2O/K_2O ratio (150 to 100) implied that the albitisation fluids remain rich in Na and poor in K for chemical equilibria of system during albitisation. The change of SiO_2 concentration also indicates that new albite replaced primary albite, although the equilibria are incomplete (Figure 4.13). The incomplete equilibria of fluid/rock reactions resulted in the total relict about 20% of primary albite grains.

4.2.1.2 Na_2O/K_2O ratios, MgO, Fe_2O_3 and TiO_2

Figure 4.13 shows an increase in MgO concentration in the new albite phase but a general decrease in Na_2O/K_2O ratios. The removal of MgO may have relied on

continual dissolution of primary minerals rather than new albite phase during low intensity albitisation. An exception of albite has high Na_2O/K_2O ratios and MgO concentration in unalbitised psammite. This may have reflected part of albitisation condition in metamorphism. Figure 4.13 shows two groups of albite with low and high concentration of TiO_2 and Fe_2O_3. That implies the Na depletion and saturation of TiO_2 and Fe_2O_3 during low intensity albitisation.
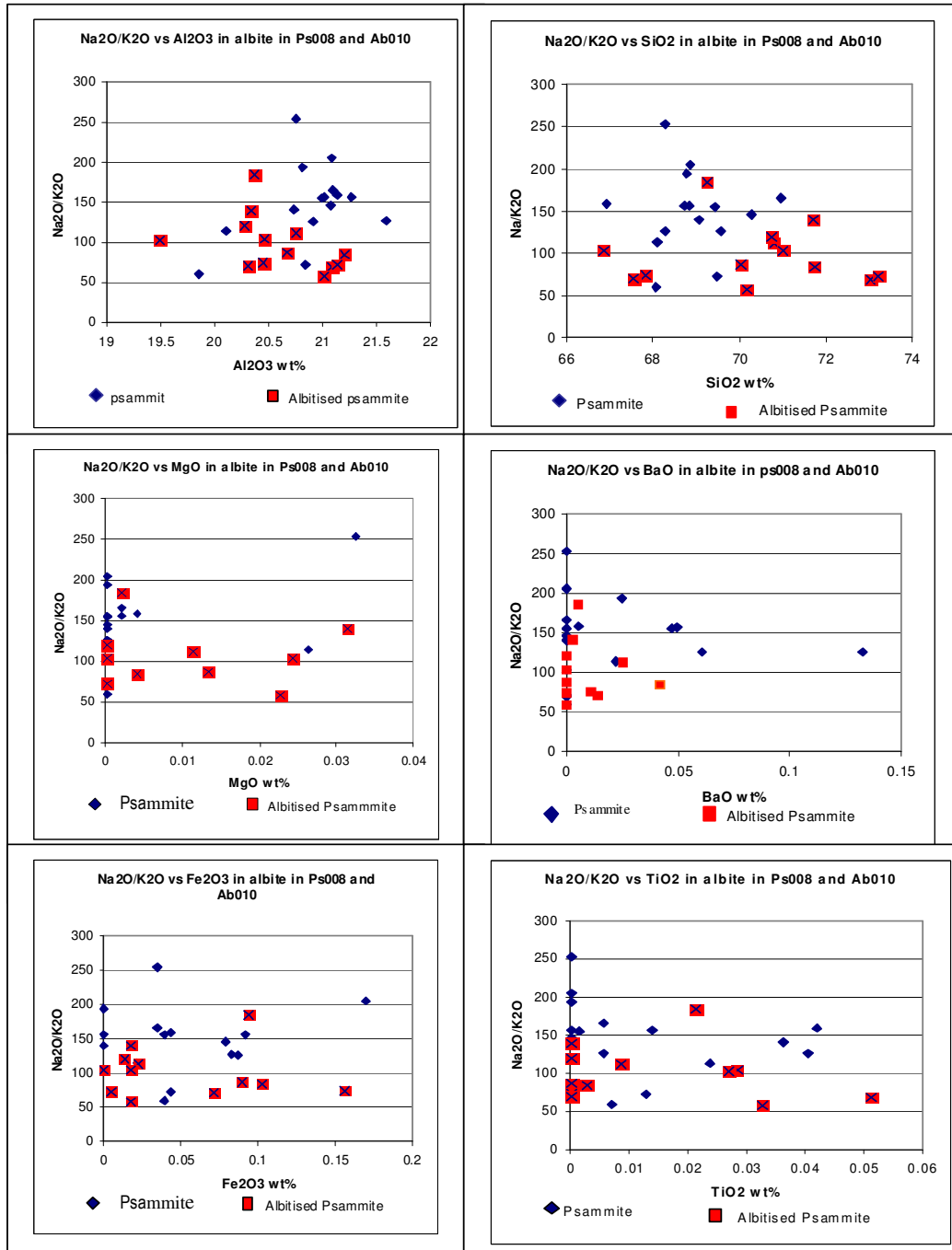
Figure 4.13 Na_2O/K_2O ratio vs Al_2O_3, SiO_2, MgO, BaO, Fe_2O_3 and TiO_2 in albite in psammite (Ps008) and albitised psammite (Ab010).
Notes: Albite is analysed by EMA.

4.2.1.3 Active elements: F and Ba

Seventeen and thirteen grains of albite were respectively analysed by the Electron Microprobe (EMP) in sample Ps008 (psammite) and Ab010 (albitised psammite) to determine the changes of F and Ba. The F concentration of albite in unalbitised psammite is on the average of 0.0791 wt% but on the average 0.116 wt% in albitised psammite albite. The average F contents were increased during albitisation. Figure 4.14 shows elevated F and Ba concentrations and the lower Na_2O/K_2O ratios in albite in albitised psammite compared to unalbitised psammite. This implies a depletion of Ba and concurrent enrichment of F during albitisation as well as a depletion of Na_2O and enrichments of K_2O in fluids during distal low intensity albitisation. This means that the albitisation fluids may contain elevated F. Fourteen grains of biotite in psammite had an average F content of 1.896 wt% and twelve grains of micas (six biotite grains and six muscovite grains) had an average F content of 0.8909 % in albitised psammite. These mean that F was removed into the fluid phase during the muscovite forming event.

The average BaO concentration of biotite was 0.041 wt% in unalbitised psammite and 0.064 wt% in albitised psammite. The average BaO concentrations of albite were 0.023 wt% in unalbitised psammite and 0.008 wt% in albitised psammite. The Ba was removed by mineral replacement reaction during the distal low intensity albitisation. Incomplete reaction resulted in total effect that is consistent with the whole rock data (Figures 4.14). Ba may continually be removed during progressive albitisation.

Forty three mineral grains were analysed by EMP in amphibolite (Am011 and AbA012) to determine variation of F and Ba. The average F concentration was 0.84 wt%. The F concentrations in albite were 0.0901 wt% in sample Am011 and 0.128 wt% in AbA012, respectively. This may imply that the rim of amphibolite had very low Na_2O before albitisation. The changes of F in albite in amphibolite were similar to that of albite in psammite and demonstrate the similarity of the behaviour of F in albitisation of amphibolite and psammite.

Thirty two grains of feldspars of pegmatite were analysed by EMP in samples Op031 (pegmatite) and Bp032 (altered pegmatite). The F concentrations were relative stable at 0.087 wt% in Op031 and at 0.098 wt% in Bp032. Eight grains of albite contained 0.087 wt% F in Op031 and seven grains of albite contained 0.090 wt% F in the sample Bp032. Thus, the biotite alteration may not affect the F of the feldspars (Chapter 4.1.1).

Increase of F in albite may imply depletion of F in fluids during low intensity albitisation of psammite and amphibolite. In turn, the decrease of F of biotite may indicate enrichments of F in the fluid during low intensity albitisation. Thus, the fluid may be continually enriched in F in the processes of break down of biotite and formation of albite. Stable concentration of F in feldspar may imply that F concentration of fluids may not change apparently during biotite alteration in pegmatites.
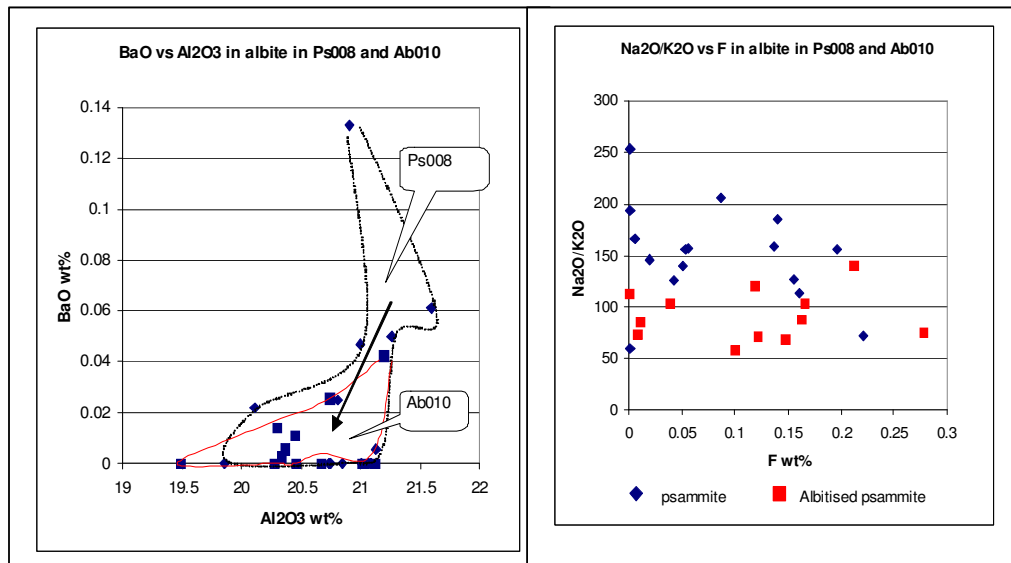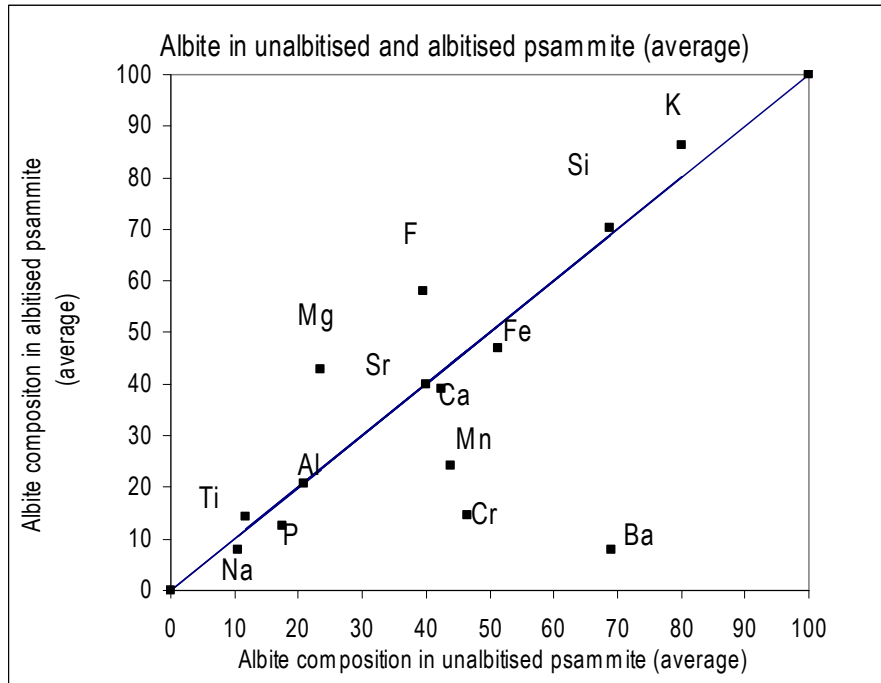
Figure 4.14 Concentration of F and BaO of albite grains in psammite and albitised psammite. Notes: 1. Decrease of Ba of albite in albitised (Ab010) compare to unalbitised psammite (Ps008) on the left. 2. Decrease of Na/K ratio but increase of F concentration in albite in albitised (Ab010) compare to unalbitised psammite (Ps008) on the right.

4.2.1.4 Compositional variation in mineral phases

Albite in albitised psammite has slightly more SiO_2 and less Al_2O_3 compared to albite in unalbitised psammite (Figure 4.15). The concentration of Fe_2O_3 and TiO_2 was at consistently low ranges of <0.15 wt% and <0.05 wt% respectively in albite in unalbitised and albitised psammite. However, the albite grains of albitised psammite have higher MgO concentration (0.00925 wt%) than albite in unalbitised psammite at 0.00469 wt% MgO. BaO concentrations decreased but F concentrations increased in albite grains in albitised psammite compared to albite in unalbitised psammite (Figure 4.15).

Figure 4.15 shows the variation of elements between albite (average) of unalbitised psammite and albite (average) of albitised psammite during low intensity albitisation. Albite grains of unalbitised psammite are replaced by albite in albitised psammite with increased F and Mg but much lower Mn, Cr and Ba contents (average) (Figure 4.15). However, Figure 4.16 shows that biotite grains in albitised psammite contains more Cr and Ba contents and less F and Mg compared to unalbitised psammite during low intensity albitisation.

Figure 4.15 Compositional variation of albite in unalbitised and albitised psammite.
Notes: 1. Element concentrations are the average of 15 albite grains of unalbitised psammite (Ps008) and the average of 13 albite grains of albitised psammite (Ab010). 2. Element concentration is arbitrarily scaled to aid in plots. 3. Albite grains are analysed by EMA.
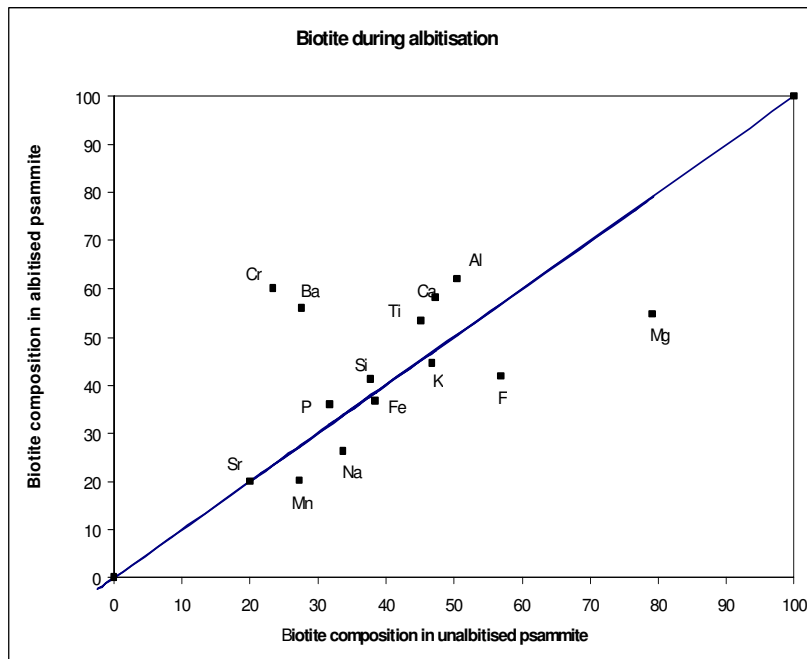
Figure 4.16 Compositional variation of biotite in unalbitised and albitised psammite.
Notes: 1. Element concentrations are the average of 14 biotite grains of psammite (Ps008) and the average of 6 biotite grains of albitised psammite (Ab010). 2. Element concentration is arbitrarily scaled to aid in plots. 3. Biotite grains are analysed by EMA.

4.2.2 Geothermometers

Geothermometers are widely applied using chemical equilibria of mineral pairs, e.g. the geothermometers of albite–biotite, albite–amphibole, and biotite–garnet (Yihunie et al. 2004; Wu & Zhao 2006; Marshall & Oliver 2006; Enami et al. 1994; John & Stormer 1975; Brown & Parsons 1981; Thompson & England 1984; Strauss 2003; Kiryukhin et al. 2002; Price et al. 1985). McCaig et al. (1990) suggested that muscovitisation occurred along a down temperature path but albitisation relates to an up–temperature path at about 400-450°C at fluid/rock ratio of 1000. Clark et al. (2005) calculated the temperature of mobility of Si for albitisation at 425°C. The variation of TiO_2 in biotite was suggested to be a reliable geothermometer at temperatures in the range of 450-550°C during regional metamorphisms (Marshall& Oliver. 2006). Enami et al. (1994) suggested the garnet zone temperature of 440°C and an albite-biotite zone temperature of 520°C. Niiranen et al. (2005) suggested the solubility of iron in saline fluids dropped drastically even in hypersaline fluids in the temperature interval of 300-500°C.

In recent studies, fluids of 29-58 wt% NaCl are suggested to have formed at temperatures of 270-330°C for dissolution of halite and 350-400 °C for dissolution of other minerals except hematite. Strauss (2003) suggested that the peak temperature of equilibrium of biotite-garnet was in the range of 486-600°C at a depth of 5 km. Kiryukhin et al (2002) applied the ratios of Na/K in feldspars as a geothermometer. The Log(Na/K) yielded temperatures of 150-200°C for albite dissolution and orthoclase precipitation. However, this situation was not observed in this study, although the ratio of Na/K was very high in the Migmatite Creek.

The minimum temperature of albitisation in Migmatite Creek is estimated as ≈ 400°C on the basis of partial dissolution of magnetite in saline fluids (Niiranen et al. 2005) and the absence of halite. The maximum temperature of albitisation is estimated as ≈ 520°C on the basis of the complete dissolution of magnetite and equilibrium of albite and low-Fe biotite. This was consistent with the temperature of biotite alteration with the equilibrium between garnet and biotite. The biotite-garnet geothermometer suggests a narrow temperature range of 450-520°C during biotite alteration, believed to represent the temperature at the end of albitisation.

Thus, during progressive albitisation, the fluid/rock interaction may vary with temperature in the range of 400-520°C for low intensity albitisation. The main mineral assemblage was albite - quartz - biotite ± amphibole in sample Ab010 of albitised psammite. Muscovite was identified only as a secondary phase in sample Ab010 but could not be identified in sample Ps008 of unalbitised psammite. The muscovite demonstrates the down-temperature path to 400°C and the albite-amphibole paragenesis defines the maximum temperature of 550°C. The albite + biotite + quartz + muscovite ± amphibole assemblage can be explained by fluctuating temperature between 400 and 520°C during low intensity albitisation. The composition of albitisation fluids continually decreased in Na content until possible saturation of fluids with respect to K, Fe and Mg for biotite or the end of albite formation. The equilibrium of albite-biotite was attained in the secondary mineral phases and some primary biotite remained, and even part of the primary biotite was replaced by small new biotite and muscovite or possibly newly formed amphibole.

During medium intensity albitisation, temperature increased slightly to 450-<500°C corresponding to the popular theory that albite formed during the up-temperature path. There are two pieces of evidence for this. Firstly, the formation of small euhedral magnetite crystals restricts the temperature to <500°C for equilibrium of albite-magnetite. Secondly, about 30% relict in primary high-Fe biotite remained in

sample Mb006. In other words, the primary biotite was only partly replaced by albite. The secondary mineral assemblage was albite-quartz-biotite. The equilibrium of albite-biotite was mostly attained at <500°C but magnetite was not precipitated above 450°C. The composition of fluids continually changed from K_2O, Fe_2O_3 and MgO poor to K_2O, Fe_2O_3 and MgO rich with a slight increase of Na_2O. The temperature of albitisation was variable in the narrow range of 450-500°C. At this temperature range, albite failed to replace biotite. This was quite different from the complete replacement of biotite by albite in high intensity albitisation. Normally, hydrothermal fluids result in temperature increase in a more stable in fluid/rock system during progressive albitisation. For low intensity albitisation, temperature may be much more variable because possible 550°C hot fluids encountered cooler 400°C rock.

During high intensity albitisation, temperature may have been at about 480-515°C because the mineral assemblage was completely changed from quartz-biotite-albite to albite–quartz assemblages. Biotite was completely replaced by albite. New accessory biotite was of quite fine-grain with light grey colour and was different from the primary coarse-gained dark grey biotite. The rough porosity and slightly light pink colour as seen in the cutting sides of specimen may be linked to Fe ions. This may relate to the mineral replacement reactions and extreme depletion of Fe. The geothermometer of equilibria of albite-quartz indicates temperatures around 515°C because most magnetite was dissolved during high intensity albitisation. Albite + quartz and accessory rutile and ilmenite indicate temperatures of albitisation below 515°C at mineral equilibrium in sample Ab004 of albitised psammite. The temperature of high intensity albitisation was restricted to the range of 480-515°C based on mineral assemblages.

In the case of complete albitisation, fine accessory magnetite and biotite grew within large euhedral albite crystals. Equilibrium may indicate a temperature of about 480 °C in the presence of hypersaline fluids and psammopelite. Thus, fluctuation of temperature may have gradually stabilised from low via medium to high intensity albitisation.

Another piece of evidence for this conclusion relates to decrease of SiO_2 in whole rock data (Chapter 3.5.1). Not only does evolution of temperature control mineral assemblages but also relate to changes of SiO_2 in the range of 400-500°C (decrease of 3.8-6.9% of SiO_2 content of rock for albitisation at 400-500°C (500 bar), Clark et al. 2005). Decrease of 5-10% SiO_2 content clearly happened during albitisation.

To summarise, the evolution of temperature was considered to be in the narrow range of 480-515°C during high intensity albitisation, 450-500°C during medium intensity albitisation and 400-520°C during low intensity albitisation in the Migmatite Creek area. The evolution of temperature was followed an up-temperature path from 400-520°C to 480-515°C during progressive albitisation and from large fluctuations to the stable narrow range.

4.2.3 REE characteristics of different rock types

Progressive albitisation was responsible for intense removal of all REE elements in metasediments (Figure 3.18). This may imply the same source for albitisation and breccia system or both of these belonged to the same fluid/rock reaction system.

The total REE content of the respective rock types can be plotted against to the concentrations of Na_2O (Figure 4.17) or Na_2O+K_2O. For example, psammite contains 3.96 wt% Na_2O with the high REE sum (559.83ppm) comparing to 4.83 wt% Na_2O (plus 1.55 wt% K_2O) with low REE sum (183.63ppm). Pegmatite has the lowest REE

sum (6.89 wt%) with the highest concentration of total of Na_2O (4.73wt%)+ K_2O (5.85 wt%) comparing to high REE sum (258.89ppm) with 3.98 wt% Na_2O and 1.65 wt% K_2O. Further considering the concentration of Al_2O_3, the REE sum in rock sequence may be actually contrasted to the alkaline feldspars (albite and K-feldspar). The ratios of Na_2O/Al_2O_3 and K_2O/Al_2O_3 are derived from alkaline feldspar ($NaAlSi_3O_8$ for albite and $KAlSi_3O_8$ for orthoclase relate to plagioclase) ignored the content of Na and K in other minerals (e.g. micas). This interpretation is also consistent with that of progressive albitisation.

Progressive albitisation continually removes the K_2O into alkaline fluids and the REE elements are either removed by the initial Na-rich fluid or by the changed Na+K rich fluid. The REE characteristic of low intensity albitisation is different from that of high intensity albitisation and they has the roughly same level of the REE sum (29.63ppm for LA and 22.67 ppm for HA). The slight difference is derived from the complete equilibrium (HA) and the incomplete equilibrium (LA) in new mineral assemblage formation or fluid/rock reaction or dissolution-precipitation. The chemical equilibria are also distinguished due to evolution of fluid composition related to conditions (e.g. T-P-pH, fluid/rock ratios).

Progressive albitisation results in extreme LREE depletion relative to slight or weak depletion of HREE. The oxidised Na-rich fluids removed the LREE from all rock types completely and formed the high charged LREE fluids (Figures 4.18 and 4.19). Although HREE are also depleted at a relative gentle content the distinguishing characteristics definitely depend on changes of the atomic radius and mass between LREE and HREE or partition. LREE have the bigger radius and less mass so that LREE are removed easier than HREE by alkaline fluids.

The minor opposite situation is possibly linked to variable spaces of rock pores (e.g. the micro-structures (amphibolites), porosity (part of breccias) and foliated structures). The factors may cause partly changes of REE characteristics but they are minor. Albitisation and alkaline feldspars are also frequently associated with the structure factors (foliation, brecciaing and porosity). Thus, the characteristic of REE is partly controlled by the structures (from microscope scale via visible to fault/antiformal scales) in rock sequences.

The total REE contents of amphibolite depend on the foliated microstructure rather than slight variation of Na contents. The REE characteristics of amphibolite were apparently not changed due to the lack of albitised gradient changes. However, the slight depletion of LREE is clear in amphibolites which are similar to other rock types. Thus, depletion of LREE is present in all rock types.

In breccias the total REE content averaged at 190.88 ppm with a range of 65.73–440.5 ppm. The REE characteristics could be dependent on the country rock which breccias belong to and also are dependent on progressive albitisation of clasts as well as the proportion of matrix material. The total REE varied from the 181.5 ppm in breccia pipes (sample BP005) to 65.73 ppm in breccia networks (sample B002). This was consistent with intensities of albitisation gradients. The REE pattern of breccias shows quite stable concentrations of REE elements with negative Eu anomaly, and was also similar to intense albitisation in shapes but rich in concentration (Figures 3.18B and 3.19). This may imply that the breccias were involved in both, the mixed mafic component and albitisation with the same system of fluid/rock reactions.

The total REE contents of the four types of pegmatite fall in 3 ranges of 6.89 ppm, 14.67–36.7 ppm, 88.41–258.89 ppm. The total REE contents of pegmatites correlate the Na_2O concentration and intensity of albitisation in reversion partition (Figures 4.17 and 4.18). The REE patterns of unalbitised pegmatite show the clearly

negative Ce anomaly without depletion of light REE (Figure 3.19). The light REE elements were removed in pegmatite during albitisation. The effect of albitisation for pegmatite could also be observed carefully in REE concentration 29.52 ppm (albitised PB034) and 155.73 ppm (unalbitised PS035). Biotite alteration of pegmatite has the highest REE content (258.89 ppm) and intensive albitisation of pegmatite has a low REE content (14.68 ppm).

Biotite alteration has the characteristic of decrease of REE contents (Figure 3.18). However, it does not cause extreme LREE depletion compare to albitisation. Its REE patterns are variable to a flat shape.

The REE contents firstly depend on the rock types or rock composition. Migmatism results in the overprinted Na-K alteration for pegmatite with decrease of REE. Albitisation overprinted in all rock types again results in the extreme LREE depletion with a distinguishing characteristic. Biotite alteration also results in decrease of REE but no apparent LREE depletion compare to albitisation. Breccias have been undertaken albitisation and partly involved in biotite alteration or matrix mechanism mixed. This results in the REE characteristic of breccias similar to that of biotite alteration locally. Extreme biotite alteration has a flat REE characteristic with possible previous low albitisation. Finally, albitisation (plus secondary biotite alteration with post-dating migmatism) produced a total effect of REE depletion in rock sequence. Albitisation of metasediments is predominate so that LREE remobilisation has obvious significant for the evolution from alkaline to metal rich fluid.
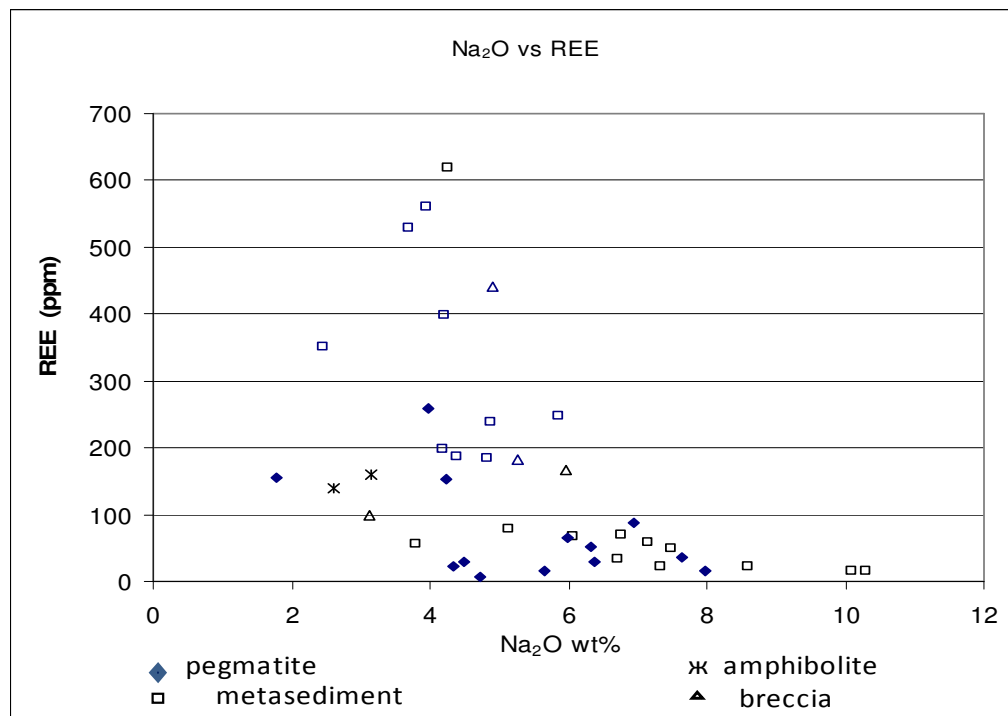

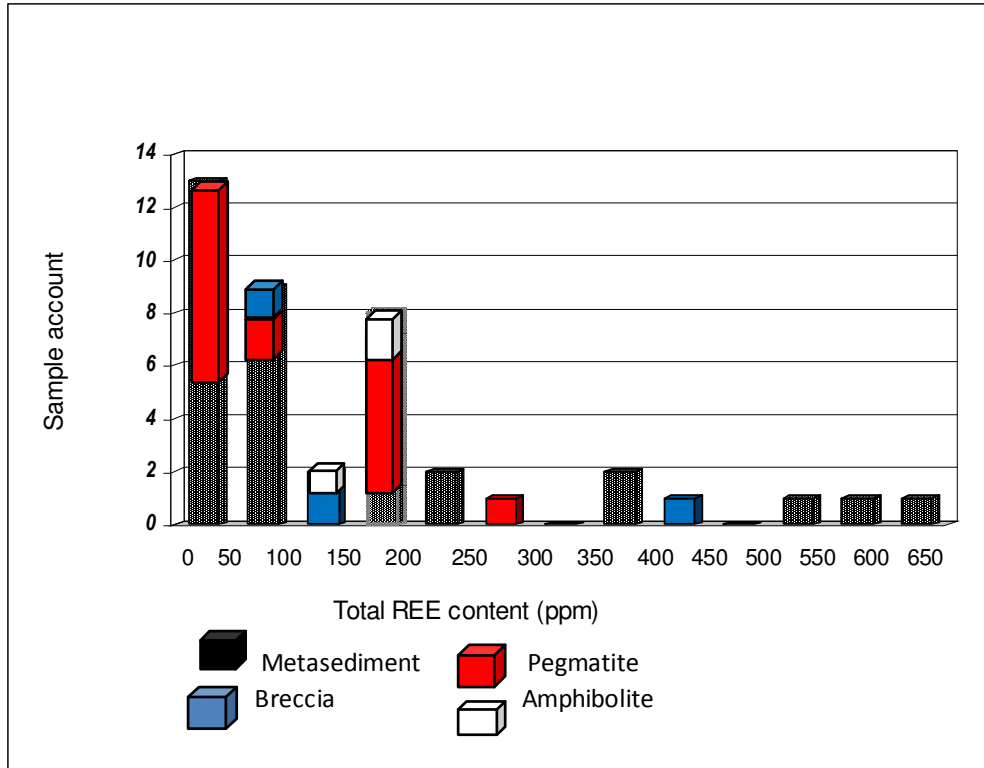**Figure 4.17** Total REE sums decrease while increasing of Na in rock types. N=40.

Figure 4.18 Showing total REE content of all samples (see also table 5). N=40.
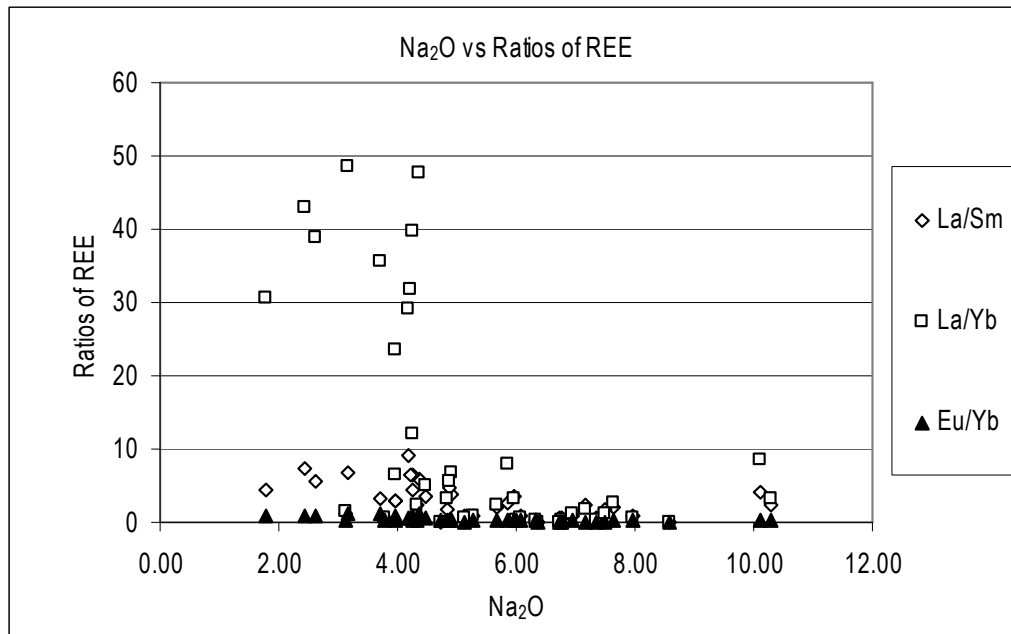

Figure 4.19 Ratios of La/Sm, La/Yb and Eu/Yb vary while increasing Na, N=40.

4.2.4 REE patterns of minerals

The REE patterns of minerals are clearly influenced by progressive albitisation with variable shapes (Figures 3.22, 3.23 and 3.24). For example, negative Eu anomalies in albite are respectively accompanied with Ce negative anomaly and U-shape patterns in unalbitised and albitised psammite. The negative Eu anomalies in albite may indicate the oxidised condition in both metamorphism and overprinted albitisation. LREE depletion in albite in albitised psammite may be derived from complete replacement of albite or just albite by new albite. Albitisation fluid has stronger ability for removal of LREE than that of HREE during changes of albite phases. However, in biotite of albitised psammite, the Eu negative anomaly partly remains and LREE is apparently enriched in biotite of albitised psammite. LREE has not apparently mobilised but HREE has. On other words, new biotite enriched slightly La and Ce but depleted Yb and Lu. Thus, the total REE contents decrease in both albite and biotite, in albitised psammite. The changes of REE patterns imply that new formation of albite and biotite result in the exchange of REE in fluid/rock system.

The changes of mineral REE patterns may infer that: 1) new formation of minerals has a complete change in different ways depending on alkaline oxidised fluid/rock system; 2) mineral assemblage either change or new minerals change themselves minerals; 3) REE patterns of minerals depend on the chemical equilibria in fluid/rock reactions; 4) dissolution-precipitation of minerals may be the basic processes rather than single mineral broken; 5) high fluid/rock ratios may exist in progressive albitisation; 6) penetrative structures (pathways) are important for albitisation linking to breccias, microstructures, foliated and porosity.

The REE patterns of minerals indicate a stable enrichment of LREE in unaltered pegmatite. The REE patterns of both albite and biotite have been completely changed in biotite alteration of pegmatite. The REE patterns of K-feldspars have been changed following up increase of total REE contents in biotite alteration of pegmatite. The negative Eu anomaly of biotite reflects the alkaline oxidized conditions in unaltered pegmatite. A REE pattern of albite indicated the enrichment of HREE in biotite alteration of pegmatite (Bp032). A positive Eu anomaly of orthoclase may imply possible change from albitisation to biotite alteration:

1) the pegmatites Op031 and Bp032 were affected by different alkaline fluid pulses derived from metasediments as suggested by Lu & Lottermoser (1997).
2) the REE patterns of biotite (Fe-K) alteration in pegmatite were related to the reduced conditions rather than the oxidised alkaline system.

The positive Eu anomaly of orthoclase is different from all REE patterns in minerals and rock types. This implies a short reduced condition related to K-alteration (pegmatite) and biotite (K-Fe) alteration in pre-dating or post-dating. Alkaline fluid/rock reaction system is major process but slight variation occurs in biotite alteration of pegmatite. Further, alkaline fluids may gradually vary to neutral condition in albitisation system.

4.2.5 The influences of albitisation on REE content and variation

The influence of albitisation on REE content and patterns was very intense and extremely variable in metasediments with LREE contents (La to Eu) decreasing by 99.71% (La 264 to 0.76 ppm) and 87.5% (Eu 40 to 5 ppm) (Chapters 3.5.2 and 3.6.2; Figures 3.18, L003). The decrease of REE content depends on intensity of the gradient of albitisation accompanying chemical equilibria of fluid/rock reaction:

1. REE elements were removed by the oxidised albitisation fluids on variable content;
2. new minerals and mineral assemblages have been formed in dissolution-precipitation processes. Albite + quartz ± accessory minerals (e.g. new magnetite) replacing the quartz + feldspar + micas + magnetite, results in remobilisation of REE;
3. REE variation reflects the incomplete to complete chemical equilibria of fluids/rock from LA to HA;
4. Extreme LREE depletion was involved in fluid/rock ratios linking to pathways (microstructures and porosity). Breccia network and antiformal hinges were the dominant sites for removal of REE.
5. High ratios of fluid/rock may be inferred from the REE evolution (depletion of LREE and decrease of REE content).

It can be inferred that REE were taken up by the oxidised albitisation fluids and only 12.5% Eu and 0.29% La remained in the albitised psammite (Ab004). HREE values (Gd to Lu) were reduced by 86.1% (Gd 63.4 to 8.3 ppm) and 52.87% (Lu 24.4 to 11.5 ppm). This satisfied the basic geochemical fluid dynamic principal related to element mass and radius in albite crystal structure sites or as inclusions. During low intensity albitisation, the LREE values (La to Eu) were reduced by 80.18% (Eu 22.7 to 4.5 ppm) to 98.75% (La 60.8 to 0.76 ppm) and HREE values were reduced by 76.5% (Gd 41.7 to 9.8 ppm) to 36.3% (Lu 25.9 to 6.5 ppm). The influence on REE is clearly correlated to the intensity of gradient albitisation form LA and HA (Figures 4.17, 4.18 and 4.19).

The influence of albitisation on the REE characteristics of metasediments (psammite) was similar to other lithological units in variable intensity for REE concentrations/nominated values. LREE was removed (La 30.5 to 5 ppm and Sm 7 to 1.5 ppm) in albitised pegmatite margin and slight LREE was removed (La 36.5 to 31 ppm and Nd 28.5 to 25 ppm) in albitised amphibolite margin and HREE has not been changed (Figures 3.18D and E). The influence of albitisation on REE producing the difference between metasediment (psammite) and amphibolite (also pegmatite), can be inferred that was derived from albitised gradient intensity in distinguished lithologies.

4.3 Mobility of Elements

4.3.1 Siderophile elements and Cu – Pb – Zn (Chalcophile)

The siderophile elements were mostly removed during albitisation (see Chapter 1.4.4). Mobility of siderophile and chalcophile (Cu, Pb and Zn) elements was found to be dependent on the chemical equilibria during fluid/rock interaction. The hypersaline albitisation fluids had distinct capacity to solve the trace metals. The composition of fluids was variable for metasediments and minor pegmatites (see Chapter 4.1). Figure 4.3 implies the average compositions of Na rich + Al(OH)x + Nb + Th + Ga ± Sr ± Zr ± Y in metasediments during progressive albitisation. The unalbitised (samples Pm017 and Op031) and albitised pegmatite (samples ABP022 and AP041) were taken to determine the evolution of pegmatite in the area. The composition of albitising fluids contained Na rich + Al + Ti + Ca ± Nb ± Zr ± Ga ± Th. Siderophile, Fe, Co and Ni were continually mobilised during progressive albitisation (Figure 4.20). Mn and Cr were mobilised during progressive albitisation (Figures 4.6, 4.7, 4.8 and 4.13). Chalcophile, Zn was completely mobilised; Pb was slightly mobilised or the least mobile during progressive albitisation (Figure 4.20).

4.3.2 Ba, U and REE

All rare earth elements were substantially removed and mainly LREE in all types of rock, metasediments and pegmatites during progressive albitisation (Figures 4.20 and 4.12). Differential mobilisation of REE elements lead to a change of the overall REE patterns between original rock and albitised rock. Mobility of light REE elements was distinctly stronger than that of heavy REE elements was. Intense mobility of light REE elements resulted in clear depletion in all albitised lithologies. Thus, progressive albitisation is a possible source process for the formation U – REE mineral deposits. The mobility of Ba and U was dependent on their concentrations in fluids of albitisation. In most cases, Ba and U were mobilised into the fluids during albitisation.
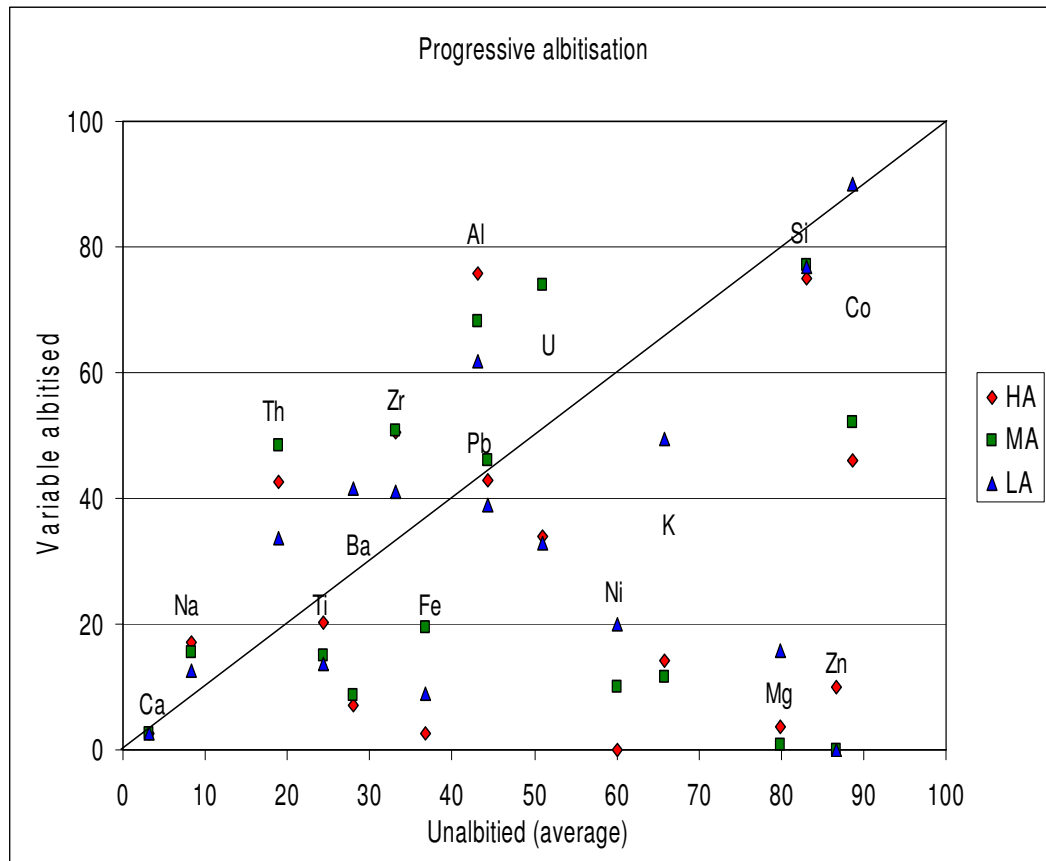

Figure 4.20 Mobility of elements during progressive albitisation.

5. CONCLUSIONS AND RECOMMENDATIONS

5.1 Conclusions

5.1.1 Albitisation System

Progressive albitisation was active during Olarian Deformation (OD_3, -1600-1580 Ma), post dating migmatization, and associated with breccia formation in the Migmatite Creek area. Using ArcGIS mapping, the intensity gradients of albitisation were shown to be structurally controlled by both the networks of breccias and the position of an antiformal OD_3 fold. Three levels of albitisation were distinguished as high intensity albitisation (HA), medium intensity albitisation (MA) and low intensity albitisation (LA) linked to breccia networks, pipes and un-brecciaed metasediments. Progressive albitisation affected the metasediments, part of pegmatites and amphibolite. Progressive albitisation resulted in the minor but intense biotite alteration as well as the late less minor quartz formation or silification/silication. The biotite alteration commonly was formed in the vicinity of synformal folds with thin-layer of metasediments or foliated structures. Silification or quartz veins formed in small faults or along cleavage planes in scales of centimetre to metre as the latest stages of fluid/rock reactions.

The fluid compositions were best modeled to be about 30 wt% NaCl equivalent salinity. The initial compositions would have been hypersaline with the H_2O-NaCl rich and KCl-$MgCl_2$-CO_2–$Al(OH)_3$–$Al(OH)_2$±$CaCl_2$.

HA formed in the temperature range 480°C-515°C for albite (An_{0-25}). LA occurred to a wider range of temperature and pressure mainly during increasing temperature and pressure from 400°C to 520°C. The variable temperature possibly resulted in the formation of muscovite.

Quartz + feldspar + biotite + magnetite assemblages in psammite were completely replaced by new albite + quartz + magnetite during medium to high intensity albitisation. The new albite-rich assemblages are likely to be the result of dissolution – repreciptation rather than single mineral break down. The main chemical equilibria were attained between albite and quartz due to the high concentration of SiO_2 (up to 86 wt%) of psammite.

The fluid/rock ratios of albitisation are consistent with the modelling results of Oliver et al (2004). The necessary fluid/rock ratios of 10^3 to 10^6 were achieved during medium to high intensity albitisation. However, the 1-10^3 of ratios of fluid/rock may be sufficient for low intensity albitisation zones. Quartz and magnetite formed in the final stages due to saturation of SiO_2 and Fe ions in parts of system after or during biotite alteration.

The albitisation system was characterised by Na-rich, high Na/K ratio (up to 60) linked to metasediment unit – psammite. This resulted in the low CaO-MnO content in range of 0.01-0.85 wt% with average 0.18 wt% CaO and <0.01-0.08 wt% MnO in all lithologies. K concentrations have two peaks in pegmatite (related to migmatism) and biotite alteration in metasediments.

5.1.2 Fluids and Mineralisation

Progressive albitisation indicates evolution of fluids associated with the hypersaline fluid sources in different ages consistent with (Lu & Lottermoser 1997; Clark et al. 2006). The alkaline-oxidised fluids have been changed by the diversity of rock sequence

in the pathway. The fluids are Na rich involved in the breccia event. CO_2 is a necessary oxidizing agent during albitisation as a component of fluids. CH_4 may not be a component of albitisation fluids but $Al(OH)_3$ is. Fluorine is a definite active component associated with biotite alteration. Fe-Mg-K-Ca has been removed from rock sequence to fluids.

Progressive albitisation resulted in changes of fluid compositions, leading to the minor biotite alteration and the latest weak quartz alteration. A highly charged, metal-rich fluid was generated during progressive albitisation bearing undeniable similarities to the potential mineralising fluids of Fe-oxide – U – REE mineralisation. The mobilisation of Zn (Pb) – Ni – Cr - V may imply the potential for Pb-Zn mineralisation and Ni – Co (– Cr – V) mineralisation after the albitisation fluids flows through some rich metal metasediments.

5.2 Recommendations

This project had an initial incentive to establish intensities of gradients of albitisation in the 3 km^2 mapping area. The extensive albitisation system identified with the Olary Orogeny may be linked to mineralisation in the Curnamona. The intense mobility of Fe, U and REE implies the realistic potential for exploration for IOCG mineralisation. Mobility of Zn – Ni – Ba – Sr clearly resulted in metal rich fluids.

Further research of albitisation is seen as highly valuable for the understanding of fluid/rock interactions and can support the current exploration in Curnamona for possible positioning of potential Fe –U – REE systems. Furthermore, three integrated-methods are recommended here: 1. Hch modellings for mineralisations related to albitisation in Curnamona; 2. Distribution characteristic of mineralisation gradients of Fe-U-REE in outcrops at or along main albitisation zone. 3. Re-integrating data of albitisation associated with breccias, possible Sr isotopes and fluid inclusions in accessory metal minerals. Three research titles are recommended to be: 1. Mobility of elements in the Curnamona Proterozoic basin. 2. Evolution of mineralisation from albitisation systems in the Curnamona. 3. Hch modelling of alteration systems in the Curnamona.

A conceptual model can produce the evolution from albitisation to mineralisation throughout the Olary Domain. Changes caused by albitisation in both the boundaries of redox and magnetic fields are valuable.