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Magmatic and Postmagmatic Evolution of the Kwandonkaya Anorogenic Ring Complex in the Nigerian A-type Granite Province.

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Abstract

The Kwandonkaya complex is one of the Mesozoic Nigerian Younger Granite complexes, which belong to the within-plate A-type granite spectrum. The complex, comprised of three main intrusive centers, contains subvolcanic rocks, plutonic rocks, and localized occurrences of altered granite. In this study, the field, petrographic and mineral chemical characteristics of the complex are investigated, a model of evolution of the subvolcanic rocks is formulated, and the mode of genesis of the parental melt are proposed.

The Fe-rich mafic minerals documented in the subvolcanic rocks suggest a comagmatic suite, and low $f(\text{O}_2)$ at the time of crystallization (ca. QFM). The subvolcanic rocks are inferred, from mica chemistry, to have equilibrated with a fluid phase characterized by low average $f_{(\text{HF})}/f_{(\text{HCl})}$ values ($10^{-1.74}$ at 773 K); values of the same ratio are higher in plutonic rocks ($10^{-0.67}$ at 673 K), and they are even higher values in the drusy and mineralized variants of the biotite granite. The occurrence of orthoclase, orthoclase + microcline, and microcline reflects decreases in the degree of undercooling, related to size of intrusive unit, and in the An content of sanidine_{ss}, which precluded complete ordering. The successful conversion of a monoclinic precursor to microcline indicates an increase in degree of volatile build-up and retention, and in extent of rock-fluid interactions in the subsolidus range of temperatures. The temperature of closure was between 450 and 300°C for orthoclase-bearing rocks, and between 450 and 200°C for those with microcline $\pm$ orthoclase.

This granite also contains dark enclaves, formed by mingling of a hybrid magma with the evolved, but not on a scale that noticeably affected its geochemical features. Petrologic modelling of the felsic subvolcanic rocks show that the rock suite could have evolved by 29% Rayleigh fractionation of liquidus phases (cumulate: sanidine_{ss}: 88%, fayalite: 6%, hedenbergite: 5%, ilmenite: 1%). Trace element evolution requires 50% fractionation of the same phases. This disagreement, between the major and trace element models, results from the role of zircon, chevkinite, and allanite, not required in the major-element model. Selective trace-element enrichment followed by nonmodal melting of a lower crustal rocks, of granulitic composition, could produce the required parental melt by 20 to 60% melting.

Resumé

Le complexe de Kwandonkaya est un des centres intrusifs de la province magmatique des "Younger Granites" au Nigéria, d'âge mésozoïque, et faisant partie de l'association intra-plaque de granites anorogéniques. Le massif se compose de trois centres intrusifs, et contient des roches subvolcaniques, des roches plutoniques et, à une échelle locale, des roches lessivées. Dans ce projet, les données de terrain, les observations pétrographiques et la composition des minéraux ont fourni les éléments essentiels pour déduire le mode d'évolution de la suite de roches subvolcaniques, et pour évaluer l'origine ultime du magma parental.

Les minéraux mafiques, très riches en fer dans cette suite subvolcanique, indiquent une suite comagmatique, et une faible fugacité d'oxygène lors de la cristallisation. D'après la composition du mica, la suite subvolcanique se serait équilibrée avec une phase fluide dont le rapport $f(\text{HF})/f(\text{HCl})$ était faible ($10^{-1.74}$ à 773 K), tandis que dans la suite de roches plutoniques, le même rapport est nettement plus élevé ($10^{-0.67}$ à 673 K). Il l'est encore davantage dans les unités à biotite qui sont minéralisées et pleines de miaroles. La distribution d'orthose, d'orthose + microcline, et de microcline dépend du taux de trempe, et donc de la dimension de l'unité intrusive, et aussi de la teneur en An de la sanidine primaire, facteur qui ralentit la mise en ordre. La transformation d'un feldspath précurseur monoclinique en microcline augmente avec le degré d'enrichissement et de conservation de la phase volatile dans le milieu, et avec la portée des interactions entre roche et phase fluide sous le solidus. Le système s'est fermé à ces interactions entre 450 et 300°C dans le cas des roches à orthose, et entre 450 et 200°C dans le cas des roches à microcline $\pm$ orthose.

Ces roches contiennent aussi des enclaves sombres, dont l'origine implique un mélange d'un magma hybride et du magma non contaminé, ce qui n'a pas sensiblement affecté les propriétés géochimiques du magma felsique prédominant. Un modèle pétrologique visant à expliquer la cristallisation du magma felsique indique un fractionnement de Rayleigh à un taux de 29% des phases sur le liquidus (sanidine_{ss}: 88%, fayalite: 6%, hédénbergite: 5%, et ilménite: 1%). Par contre, les éléments traces requièrent un taux de fractionnement de 50% des mêmes minéraux. Ce décalage entre les taux de fractionnement s'explique en considérant le rôle des minéraux primaires accessoires, comme le zircon, la chevkinite et l'allanite, non requis par le modèle fondé sur les éléments majeurs. Un enrichissement sélectif de la croûte granulitique inférieure, suivi d'une fusion partielle non modale, pourrait produire le magma parental requis suite à un taux de fusion entre 20 et 60%.

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Dedication

This thesis is dedicated to my mum and my late father

TABLE OF CONTENTS

Abstract	i
Resumé	ii
Acknowledgements	iii
Table of Contents	v
List of Figures	xii
List of Tables	xvii
Chapter 1: General Introduction	1
1.1 Preamble	1
1.2 Definition of problem, objectives, and methods of study	1
1.3 Previous work	4
1.4 Location and basis of choice of study area	6
1.5 Format of thesis presentation	7
1.6 Statement on the originality of the research	10
Chapter 2: Field geology, petrography, and geochemical characterization	12
2.1 Introduction	12
2.2 Geological setting of the NYG complexes	14
2.3 Field relations	14
2.4 Distribution of map units	15
2.5 Structural model	19
2.6 Petrography of the subvolcanic units	22
2.7 Petrography of the plutonic rocks	26
2.8 Minor rock types	26

2.9 Altered rocks	27
2.10 Geochemical Characteristics	29
2.10.1 Major elements	29
2.10.2 Trace elements	35
2.10.3 Rare-earth elements	42
2.11 Discussion	46
2.11.1 Magmatic evolution	47
2.11.2 Inferred evolution of biotite granite	51
2.11.3 Postmagmatic processes	54
2.12 Summary and conclusions	56
 Chapter 3: Olivine, pyroxene, and amphibole chemistry as a key to magmatic to postmagmatic history of the subvolcanic rocks	 62
3.1 Introduction	62
3.2 Distribution of subvolcanic rocks	64
3.3 Petrography of the subvolcanic rocks	66
3.4 Mineral Chemistry	70
3.4.1 Analytical methods	72
3.4.2 Fayalite	73
3.4.3 Pyroxene	73
3.4.4 Amphibole	76
3.5 Discussion	80
3.5.1 Magmatic evolution	82
3.5.2 Subsolidus re-equilibration	86
3.5.3 Inferred P-T conditions of formation	87

3.6 Summary and conclusion	89
Chapter 4: Mica chemistry as an indicator of magmatic to postmagmatic evolution, and halogen fugacities.	94
4.1 Introduction	94
4.2 Distribution of biotite-bearing rock units	96
4.3 Petrography of the subvolcanic rocks	98
4.4 Petrography of the plutonic and altered rocks	100
4.5 Mica Chemistry	105
4.5.1 Analytical methods	105
4.5.2 Variations in mica chemistry	106
4.5.3 Halogen content and Fe-F avoidance	113
4.5.4 Inferred halogen fugacities of fluids	116
4.6 Chlorite	120
4.7 Discussion	120
4.7.1 Late-magmatic to subsolidus mica	121
4.7.2 Secondary mica	124
4.7.3 Inferred halogen fugacities	125
4.7.4 Inferred T-f(O_2) conditions of crystallization.	126
4.8 Conclusions	128
Chapter 5: Alkali feldspars as monitors of magmatic evolution, and postmagmatic equilibration.	134
5.1 Introduction	134
5.2 Methods of Study	135
5.3 Petrography of subvolcanic rocks	135

5.4 Petrography of plutonic rocks	140
5.5 Degree of Al-Si order in the alkali feldspars	142
5.5.1 <i>Orthoclase</i>	145
5.5.2 <i>Orthoclase coexisting with microcline</i>	148
5.5.3 <i>Intermediate microcline</i>	149
5.5.4 <i>Low microcline</i>	151
5.5.6 <i>Sodic plagioclase</i>	152
5.6 Strain in the K-feldspars	153
5.7 Discussion	154
5.7.1 <i>Textural features</i>	154
5.7.2 <i>Implications of observed structural states of K-feldspar</i>	157
5.8 Conclusions	163
 Chapter 6: Petrographic and SEM study of chevkinite, allanite, and other accessory minerals and their significance in the evolution of the complex	167
6.1 Introduction	167
6.2 Petrography and SEM studies of accessory minerals	168
6.2.1 <i>Chevkinite</i>	168
6.2.2 <i>Allanite</i>	172
6.2.3 <i>Fe-Ti oxides</i>	174
6.2.4 <i>Zircon and thorite</i>	176
6.2.5 <i>Fluorite</i>	177
6.2.6 <i>Other accessory minerals</i>	179
6.3 Discussion	181
6.3.1 <i>Magmatic crystallization of accessory minerals</i>	182

6.3.2 <i>Local saturation</i>	186
6.3.3 <i>Zoning in some accessory minerals</i>	187
6.3.4 <i>Subsolidus - hydrothermal remobilization and enrichments.</i>	189
6.3.5 <i>Metamictization of chevkinite, allanite, zircon and thorite</i>	190
6.4 Summary and conclusions	191
Chapter 7: Evidence of magma mixing, fractionation, and commingling of magmas - a textural, mineral chemical, and petrochemical study of enclaves.	192
7.1 Introduction	192
7.2 Field relations of mixed rocks	194
7.3 Petrography of enclaves	196
7.3.1 <i>Enclaves in the subvolcanic rocks</i>	196
7.3.2 <i>Enclaves in biotite granite</i>	200
7.4 Mineral Chemistry	203
7.4.1 <i>Plagioclase</i>	204
7.4.2 <i>Pyroxene</i>	206
7.4.3 <i>Amphibole</i>	209
7.4.4 <i>Mica</i>	211
7.5 Geochemical characteristics	213
7.5.1 <i>Major elements</i>	214
7.5.2 <i>Trace elements</i>	216
7.6 Petrochemical model(s) of enclave evolution	219
7.6.1 <i>The restite model</i>	219
7.6.2 <i>Magma-mixing model</i>	221
7.6.3 <i>Fractionation or cumulate model</i>	222

7.6.4 Other modes of enclave modification	225
7.7 Discussion	226
7.7.1 Petrographic, mineral chemical and petrochemical features	226
7.7.2 Mechanism of hybridization, and structural implications	228
7.8 Summary and conclusions	232
 Chapter 8: Petrogenetic model of evolution of the subvolcanic felsic rocks of the Kwandonkaya complex.	 239
8.1 Introduction	239
8.2 Constraints on the evolution of subvolcanic rocks at Kwandonkaya	240
8.3 Petrogenetic modelling	245
8.3.1 Method of modelling	245
8.3.2 Major-element model (LSMC) for the subvolcanic rock suite	246
8.3.3 Trace-element evolution of subvolcanic rock suite	246
8.4 Review of models previously proposed for the origin of A-type granites	253
8.4.1 Mantle source for NYG (A-type) granites	253
8.4.2 Mixed mantle and crustal sources	255
8.4.3 The residual source (or restite) model	255
8.4.4 Partial melting model	256
8.5 Partial melting origin for the parental melt of the subvolcanic rocks	258
8.5.1 Equilibrium modal melting	258
8.5.2 Equilibrium nonmodal melting	261
8.5.3 Parental melt generation by modifying p^i values	264
8.5.4 Generation of parental melt by metasomatic enrichment followed by partial melting	265

8.6 Discussion	269
8.6.1 <i>Evolution of the subvolcanic suite</i>	269
8.6.2 <i>Genesis of parental melt for the subvolcanic suite</i>	270
8.7 Summary and conclusions	272
Chapter 9: Summary, contributions of this study, and recommendations for further studies	274
9.1 Summary and conclusions	274
9.1.1 <i>Generation of the parental melt</i>	274
9.1.2 <i>Implications of melting model to NYG and A-type granite magmatism</i>	275
9.1.3 <i>Structural and Magmatic Evolution of the Kwandonkaya complex</i>	278
9.1.3 <i>Postmagmatic evolution</i>	281
9.2 Contributions of this study	285
9.3. Recommendations for further studies.	286
References	287
List of Appendices	319

LIST OF FIGURES

Fig. Description	Page
1.1: Location of Kwandonkaya complex in the Nigerian Younger Granite province	2
1.2: a) Photo mosaic of aerial photographs of the southwestern part of topographic sheet 148	8
1.2: b) The western margin of Kwandonkaya complex	8
1.2: c) The eastern side of Kwandonkaya complex, as viewed from the Bauchi plain	8
1.3: A simplified geological map of Kwandonkaya complex (boxed area in Fig. 1.1)	9
2.1: A geological map of Kwandonkaya complex (boxed area in Fig. 1.1)	13
2.2a: Schematic diagram illustrating the structure of Kwandonkaya complex	16
2.2b: Structural map of the basement country rocks surrounding the Kwandonkaya complex	20
2.3: Petrographic features of subvolcanic rocks.	24
2.4: Minor rock types, with features indicative of magma mixing	28
2.5: Evidence of mineralization in the biotite granite	28
2.6: Plots of A/NK vs. A/CNK, Al_2O_3 , and Na_2O vs. SiO_2	31
2.7: Plots of K_2O , CaO , TiO_2 , and ΣFeO vs. SiO_2	33
2.8: Plots of FeO , Fe_2O_3 , and MnO vs. SiO_2 , and normative Ab-Qtz-Or plot	36
2.9: Plots of the concentrations of Sr, Ba, and Rb vs. Zr (in ppm), and triangular plot of Ba-Rb-Sr	38
2.10: Plots of concentrations of Ni, Pb, Th, and U vs. Zr, all in ppm	39
2.11: Plots of concentrations of Nb, and Y vs. Zr in ppm, and Log Nb vs. Log Y, and Log Rb vs. Log (Y + Nb)	40
2.12: Chondrite-normalized REE plots for rock units from Kwandonkaya complex	43
2.13: Trace-element abundances in topaz-bearing granites in this study compared to the low-phosphorus type of topaz granite from Mount Pleasant Ridge, New Brunswick	53

3.1: Geological map of the Kwandonkaya complex, with location of samples used in this study	63
3.2: Photomicrograph of anhedral (rounded) hedenbergite (Hed) inclusions within simply twinned orthoclase micropertthite	69
3.3: Photomicrograph of multigranular aggregates of hastingsitic hornblende (Amp) within hastingsite granite porphyry	69
3.4: Photomicrograph fayalite replaced principally by iddingsite	69
3.5: A SEM-BSE (scanning electron microscope - back-scattered-electron) image of replaced fayalite (Fa)	69
3.6: Petrographic features of amphibole in the Jubga-Ziem ring dyke	71
3.7: Plot of pyroxene data in the pyroxene quadrilateral En-Dp-Hd-Fs	75
3.8: A plot of the lower part of the Mg-Fe-Na diagram for pyroxene compositions	75
3.9: Plots of Mg/(Mg+Fe) vs. Si (apfu) for amphiboles	78
3.10: Plot of (Ca + ^{IV} Al) vs. (Si + Na + K) for amphiboles	78
3.11: Plots of a) F/Cl vs. F/FM [Fe/(Fe+Mg)], and b) F/Cl vs. (Si + Na + K) for amphiboles	81
3.12: a) A triangular plot of normative albite-orthoclase-quartz for the subvolcanic rocks of Kwandonkaya complex.	83
3.12: b) P-T diagram displaying the stability of ferromagnesian phases and the granite solidus at different water activities	83
4.1: Geological map of Kwandonkaya complex with location of samples whose mica was analyzed	97
4.2: SEM-BSE (Scanning electron microscope - back-scattered electron) images of mica in "hornblende"-biotite granite porphyry	101
4.3: SEM-BSE images of mica in biotite granite	101
4.4: SEM-BSE images of mica in drusy biotite granites (a, b), and altered rocks (c, d). . .	103
4.5: Plots of ^{VI} Al vs. Fe ²⁺ (a), F/FM vs. ^{IV} Al (b), ^{VI} Ti vs. Si (c), and Triangular plot of R ³⁺ - R ²⁺ - Li ⁺ (d).	108
4.6: Log X _{F/Cl} vs. X(Fe) (a), Log X _{F/Cl} vs. X _{Mg} (b), P-T diagram (c), and Log f(O ₂) - T diagram (d)	110

4.7: The distribution of mica types and general variations in inferred $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ of major rock units	122
5.1: Geological map of Kwandonkaya complex with location of samples used in this study	136
5.2: Petrographic characteristics of alkali feldspars in subvolcanic rocks	139
5.3: SEM - BSE images of alkali feldspar in plutonic rocks	141
5.4: a) Plot of $c(\text{\AA})$ versus $b(\text{\AA})$ for K-rich alkali feldspars	144
5.4: b) Enlargement of the b - c plot for the albite-rich composition (inset shows area of b - c plot enlarged)	144
5.5: Plot of β^* versus γ^* values for the sodic plagioclase with An contents indicated	146
5.6: Plots illustrating ordering paths for K-rich alkali feldspars	158
5.7: The distribution of alkali feldspars in major rock units of the Kwandonkaya complex	160
6.1: SEM-BSE (scanning-electron-microscope - back-scattered-electron) images of chevkinite in various mineral associations within subvolcanic rocks	171
6.2: SEM - BSE images of chevkinite (a - d), and allanite (b, e - f) in subvolcanic rocks	173
6.3: SEM - BSE images of ilmenite displaying different textural features and associated with a variety of accessory minerals	175
6.4: SEM-BSE images of other accessory minerals (ilmenite, xenotime, fluocerite, columbite-cassiterite, zircon-thorite, and an unknown Th-REE mineral)	178
6.5: Plots of REE and trace element data for cone-sheet units	184
6.6: Plots of REE and trace element data for ring dyke units	184
6.7: Plots of REE and trace element data for pristine biotite granite	184
6.8: Plots of REE and trace element data for altered biotite granite units	184
7.1: Location of mixed rocks analyzed in this study from the Kwandonkaya complex	195
7.2: Field photographs, photomicrographs, and SEM - BSE images of trachyandesite enclaves	198
7.3: Petrographic characteristics of microsyenodiorite enclaves	199

7.4: Field photographs and photomicrographs of enclaves in biotite granite	201
7.5: Plagioclase compositions in the enclaves in biotite granite	205
7.6: Compositional variation of pyroxenes in enclaves and their felsic host-rocks	207
7.7: Plots of amphibole composition in enclaves and their host rocks	210
7.8. Compositional variations of mica in enclaves and biotite granite	212
7.9: Plots of major element variations in enclaves and their host rocks	215
7.10: Plots of trace element variations for enclaves and their host rocks	217
7.11: Geochemical plots to qualitatively appraise the restite (a - b), and mixing (c - d) models	220
7.12. Plots of trace elements to qualitatively appraise the fractionation model	224
7.13: Illustration of mechanisms of mixing and structural implications	230
8.1: Triangular plots of normative Qtz-Ab-Or for NYG complexes (a) and the Kwandonkaya subvolcanic rocks (b)	241
8.2a - h: Log-Log trace-element plots, with ratios of concentrations indicated	242
8.2i - l: Plots of chondrite-normalized REE, and selected REE ratios that have trends consistent with fractional crystallization	243
8.3: Schematic log-log plot of concentration of a compatible element (e.g., Ba) vs. that of an incompatible element (e.g., Rb), to illustrate idealized paths of evolution for equilibrium (EC) and perfect fractional crystallization (FC)	251
8.4: Plots of evolution in trace-element concentrations expected as result of Rayleigh fractional crystallization, and equilibrium crystallization	252
8.5: Chondrite-normalized plots of trace-element concentrations in modal melts (generated by 20 to 60% melting of selected non-enriched source-rocks)	262
8.6: Chondrite-normalized plots of nonmodal melts generated by 20 to 60% melting of non-enriched trial source-rocks	263
8.7: Composition of source rocks in terms of selected trace elements and required enrichments (for generation of target melt by 20% and 60% modal melting)	266
8.8: Plots of enrichments required to modify trial source-rock (for the generation of target- melt compositions by 20 to 60% nonmodal melting)	268

9.1: A schematic lithosphere section (NNE-SSW) through the crust of the Jos Plateau region in northern Nigeria.	276
9.2: Framework geotherms for the continental crust	277

LIST OF TABLES

Table # Description

2.1 Petrographic features for the main rock types of Kwandonkaya complex.	23
2.2: Representative major and trace element data, and CIPW normative compositions . . .	59
2.3: Rare-earth-element data (in ppm) for selected rock samples of the Kwandonkaya complex	60
2.4: A comparison of characteristics of low-phosphorus subtype of topaz granites (cf. Taylor 1992) and topaz-bearing biotite granites in the Kwandonkaya complex . . .	61
3.1: Summary of petrographic features for subvolcanic rocks of Kwandonkaya complex	67
3.2: Average composition of fayalite	91
3.3: Average compositions of clinopyroxene	92
3.4: Average composition of amphibole	93
4.1: Petrographic features of mica-bearing rock types in the Kwandonkaya complex	99
4.2: Average composition of mica	131
4.3: Range of dark mica composition(s) in the various rock units	132
4.4: Inferred ratio of halogen fugacities $\{\text{Log}[f_{(\text{HF})}/f_{(\text{HCl})}]\}$ for Kwandonkaya suite.	133
5.1: Summary of petrographic features of alkali feldspars in rock units (Fig. 5.1)	137
5.2: Parameters calculated from unit-cell (XRD) measurements	166
6.1: Accessory mineral suite in the main rock types of the Kwandonkaya complex	170
7.1: Summary of petrographic characteristics of the enclaves	197
7.2: Results of electron-microprobe analyses of plagioclase in enclaves within biotite granite	234
7.3: Results of electron-microprobe analyses of pyroxene in enclaves and their host rocks	235
7.4: Results of electron-microprobe analyses of amphibole in enclaves and their host rocks	236

7.5: Results of electron-microprobe analyses of mica in enclaves within biotite granite of the Panshanu center	237
7.6: Whole-rock chemical data for enclaves and their felsic host rocks	238
8.1: Trace-element ratios (along with DI, ASI, and AI) for subvolcanic rocks and pristine biotite granites of the Kwandonkaya complex	244
8.2. Least-squares major-element calculation for the subvolcanic rocks of Kwandonkaya complex	247
8.3: Table of crystal/liquid partition coefficients used in fractional crystallization model	250
8.4: Range of concentrations of major and selected trace element for the Pan-African (Older) Granites and subvolcanic rocks of the Kwandonkaya complex	257
8.5a: Source mode and trace-element composition of source rocks (protoliths) used in partial melting model	259
8.5b: Partition coefficients used in evaluation of partial melting of trial source-rocks	260

Chapter 1: General Introduction

1.1 Preamble

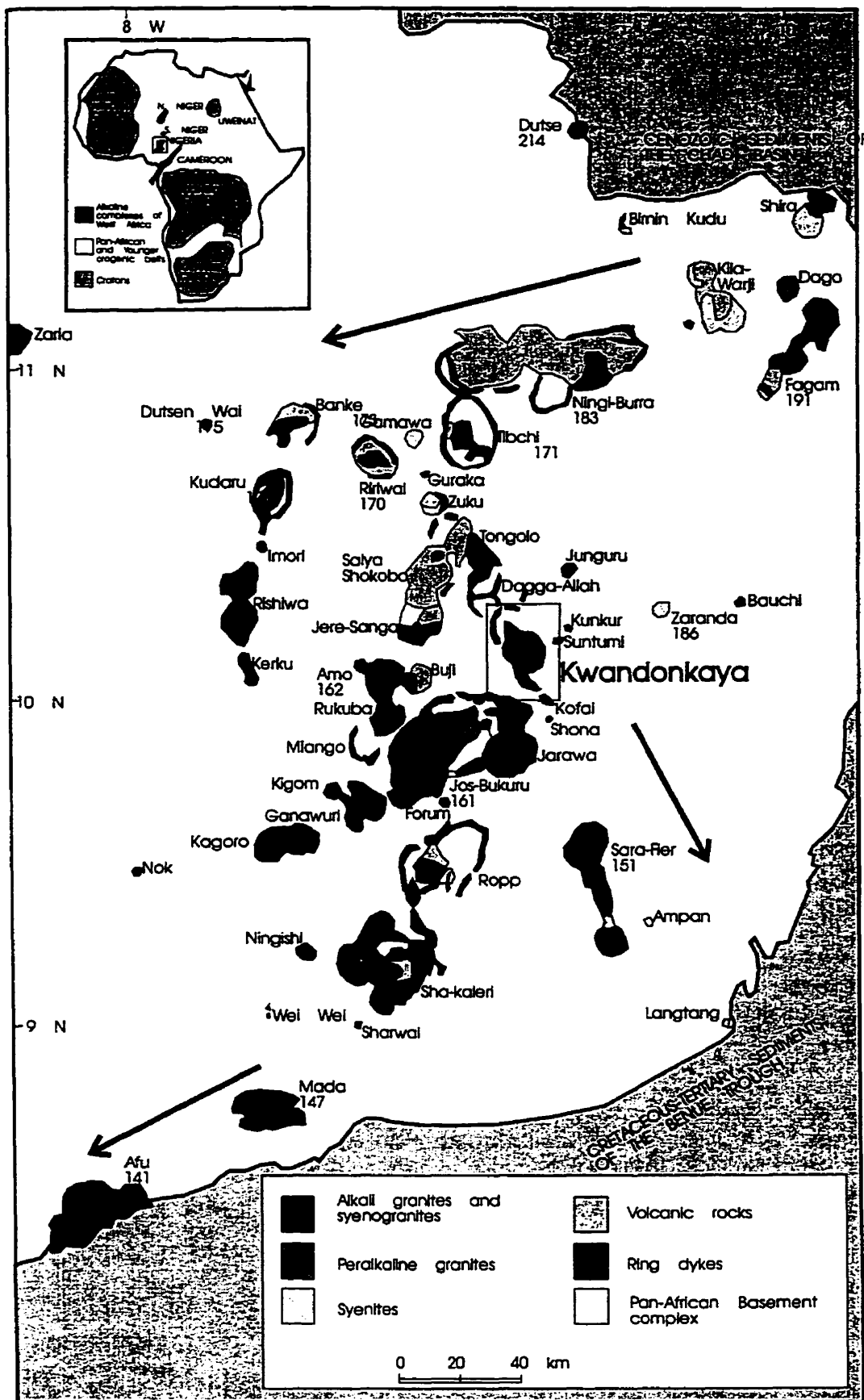
Despite the over half century of mapping and scientific studies of the NYG (Nigerian Younger Granite) ring complexes, questions still remain with respect to their origin and evolution. The level of erosion of the different complexes is variable, even in adjacent ring centers of multicentered complexes, although several rock types are noted to recur (Bowden et al. 1984). Overall, each of the NYG complexes exposes a unique combination of rock types, and hence provides bits and pieces of the history of the NYG. Therefore, the current study proposed for the Kwandonkaya complex contributes to a complex puzzle with respect to the NYG. The identification of the NYG as an important metallogenic province (i.e., the provenance area for Sn, Nb, and W) at the end of the 19th century (Falconer 1911), and their recognition as a distinct petrographic province initiated a flurry of activity that included mining, research, and growth of mining towns (for example, Jos, also known as Tin City, which developed initially to provide accommodation and ancillary services to the mining communities).

The Kwandonkaya complex lies on the western edge of the Mesozoic (ca. 191-144 Ma) NYG province (Fig. 1.1). The name "Kwandonkaya" is derived from two Hausa words, *Kwandon* meaning "basket of" and *kaya*, meaning "load" in allusion to the topographically distinct Kwandonkaya hills, which break the monotony of the Bauchi plains.

1.2 Definition of problem, objectives, and methods of study

Peraluminous biotite granite seems to be relatively insignificant in most A-type granite provinces. However, such a rock type constitutes about 60% (in areal extent) of the NYG, compared to peralkaline rocks (15%). This contrasts with the predominance of peralkaline rocks

Fig. 1.1: Location of Kwandonkaya complex in the Nigerian Younger Granite province. The map (after Kinnaird 1981, 1985) shows the location of Kwandonkaya complex in a boxed area. Arrows indicate direction of migration of ages of complexes. Numbers beside some complexes are ages in million years (after Rahaman et al. 1984). Inset in the top left corner shows location of the Niger and Nigerian A-type magmatic provinces in Africa.



(45%) relative to peraluminous biotite granites (3%) in the collinear but somewhat older A-type granites of the Niger Republic. The cause of this interesting contrast is beyond the scope of this study; the absence of peralkaline rocks in the Kwandonkaya complex limits this study to the evolution of metaluminous-peraluminous rocks.

The objectives of this study are a) to characterize the Kwandonkaya anorogenic complex in terms of its field geology, petrography, mineralogy, mineral chemistry, and whole-rock chemistry, and b) to define the most plausible petrogenetic model for the evolution of the predominantly peraluminous Kwandonkaya complex. Such a model of evolution could be applicable to similar, predominantly peraluminous ring complexes in the NYG, as well as in other A-type granite provinces. The model also intends to further the understanding of the unique association of some peraluminous biotite granites with Sn-W-Nb mineralization.

The detailed methods of study and analytical techniques are presented in introductory parts of the various chapters, and only an outline of the methods is provided here. This study involved field mapping and sampling of the Kwandonkaya complex during the summers of 1989 and 1990. The previous geological map on the scale of 1:100,000 (after Buchanan 1971) and a topographic map (1:50,000) were used as base maps. Subsequently, petrographic descriptions of hand specimens, and thin sections (about 98 polished and 60 covered) of selected samples were undertaken. Scanning electron microscopy with energy-dispersion spectroscopy (SEM-EDS) allowed the identification and the characterization of zoning in accessory phases. Electron-microprobe analyses were undertaken with wavelength-dispersion spectroscopy (WDS) technique using a Cameca - CAMEBAX electron microprobe, with four spectrometers. The ferromagnesian minerals, along with products of their alteration, were analyzed to define magmatic compositions and postmagmatic overprints. X-ray diffraction (powder method) was useful to characterize the alkali feldspars structurally. Whole-rock chemical analyses were undertaken using X-ray

fluorescence (XRF) and atomic absorption spectroscopy (AAS) for the major and trace elements (Geochemical Laboratory of McGill University, T. Ahmedali and G. Keating, analysts), and inductively coupled plasma - mass spectrometry (ICP - MS for the REE (Memorial University of Newfoundland, S. Jackson, analyst). The data collected are presented in Chapters 2 to 8.

1.3 Previous work

Loiselle and Wones (1979) first defined a class of granites as being "A-type", comprising of alkali feldspar granites (+ peralkaline and peraluminous variants), syenites, gabbros, and anorthosites (e.g., in Niger Republic, Bowden et al. 1987, and southeastern Australia, Collins et al. 1982). The A-type granites have been characterized by Clemens et al. (1983, 1986), and Whalen et al. (1987): i) they lie in the granite field in the QAP (quartz - alkali feldspar - plagioclase) plot of Streckeisen (1973, 1976), ii) they are invariably anorogenic, iii) they usually occur in subvolcanic-plutonic complexes, associated with lava flows, and less commonly ash flows, iv) they are usually comprised of metaluminous Ca-poor granites, with abundances of Sr, Ba, and Rb similar to those of S- and I-type granites; the metaluminous A-type granites are usually coeval with peralkaline and peraluminous rock-types, and v) the suite of A-type granites is typically enriched in high-field-strength elements (HFSE), e.g., Zr, Nb, Y, the rare earth elements (REEs), with high Fe/Mg, (Na + K)/Al, K/Na ratios, high concentrations of F, Ga, and Zn, and low levels of Mg, Ca, Al, Cr, and Ni. The use of the term A-type granites has been questioned (Creaser et al. 1991) on the basis of overlapping features with other types of granites. However, recent debates have established these granites as a distinct class (Eby et al. 1992, Bowden & Kinnaird 1992). The uncertainty on the mode of origin and evolution of this class of granites is greater than that on the validity of this suite of rocks. The proposed models of origin (Loiselle & Wones 1979, Collins et al. 1982, White & Chappell 1983, Clemens et al. 1986,

Chappell et al. 1987, Creaser et al. 1991, Skjerlie & Johnston 1992, 1994) will be reviewed in discussions of the evolution of the Kwandonkaya complex.

The NYG complexes have received concerted attention since their distinction from the Pan-African Older Granites by Falconer (1911). Hossain & Turaki (1983) compiled a bibliography restricted to articles written after 1911, on the geology and mineralization of the NYG. Furthermore, Kinnaird (1987) undertook a comprehensive review of studies on geology, mineral chemistry, petrochemistry, age, and ultimate origin. A review of previous work, with respect to the various aspects studied in this work, is undertaken in individual chapters; only highlights of previous work are given in this section.

The first composite geological map of the complexes on the scale of 1:500,000 was compiled by Kinnaird (1981). Research into the origin and evolution of A-type granites in Africa is contained in specialized publications (Jacobson et al. 1958, Buchanan et al. 1971, Bennett et al. 1984, Kinnaird & Bowden 1987). The regional geology and mineralization of the NYG are reviewed by Bowden et al. (1981, 1984, 1987). Trends in sequential age of emplacement were documented by van Breemen & Bowden (1973) and van Breemen et al. (1975), with southward migration (younging) of ring complexes from Niger to Nigeria. Further research indicated that the age migration occurred in 60-80 Ma cycles: Ordovician - Silurian and Devonian - Carboniferous in Niger, and Triassic - Jurassic in Nigeria (Rahaman et al. 1984). Borley (1973, 1976) undertook the only known regional studies of the mineralogy of ferromagnesian phases in the NYG, and defined peralkaline and aluminous evolutionary trends for pyroxenes and amphiboles. Several models have been proposed for the origin and evolution of the NYG, including fractionation from a mildly alkaline, orthopyroxene-normative olivine basalt to produce either aluminous or peralkaline trends (Jacobson et al. 1958, Black 1965, Bowden et al. 1976a, Turner & Bowden 1979, Bowden 1982). An independent source is suggested for the voluminous felsic magma

(Bowden 1982). Brown and Bowden (1973) modeled the felsic metaluminous and peraluminous rocks as being derived by partial melting of charnockitic rocks. Isotopic studies so far have indicated the Kila Warji syenite to be the least evolved of the felsic units in the NYG; it is considered the parental composition from which other syenites and granites evolved (van Breemen & Bowden 1973, Bowden et al. 1976b, Bowden & Kinnaird 1992). In a review of the geology and origin of the Niger - Nigerian A-type granite province, Bowden et al. (1987) showed that the Niger subprovince illustrates well the origin of A-type granites by fractionation from a mafic melt, whereas the NYG subprovince illustrates the importance of postmagmatic overprints and mineralization.

Despite the large size of the Kwandonkaya complex, and its closeness to the best-studied complex in the NYG (i.e., the Jos-Bukuru complex), it has received minimal attention. Bain & Russ (1923) first surveyed the complex and interpreted its highly variable texture as an indication that the apical part of the pluton is exposed. During wartime efforts to locate tungsten resources, the Kwandonkaya complex was resurveyed (Jacobson & Jacques 1944), and an updated geological map published from the field survey started by R. Black in 1956, and completed by Buchanan (1971). Imeokparia (1984) undertook a geological-geochemical study of the complex, but did not include the Buri ring dyke and cone sheets, documented in this study.

1.4 Location and basis of choice of study area

The Kwandonkaya complex is located on the east-central margin of the Jos plateau, in the Toro local government area of Bauchi state of Nigeria. The complex is accessible by the Jos to Bauchi road, about 43 km northeast of Jos town (not shown on Fig. 1.1) and with Bauchi being about 75 km to the northeast. The complexes that surround the Kwandonkaya complex include: Dagga Allah complex to the north, Kunkur and Sutumi complexes to the northeast, Zaranda

complex to the east, Kofai to the south and Jos-Bukuru complex to the southwest (Fig. 1.1). The area of study lies between longitudes 9°6' and 9°18'W, and between latitudes 10°00' and 10°18'N.

The geological setting and occurrence of one of the largest exposures of peraluminous biotite granite in the NYG make the complex an appropriate choice for this study. A detailed discussion of the geological setting is undertaken in Chapter 2. The quality of exposure is excellent (Fig. 1.2a - c), with vertical sections reaching 600 meters above the ground level (or 1400 meters above sea level) and preserving the transition from the subvolcanic (ring-dyke level), to the apical or roof zone of the mass of biotite granite. The complex is therefore comprised of texturally and petrographically diverse units (Fig. 1.3). The accessibility and lack of major settlements enabled unimpeded observation of geological features. The complex has been known for alluvial and eluvial mining of cassiterite and columbite since 1928, and is currently of interest for its potential for gem-quality topaz and beryl (aquamarine). There are no data on production or on potential reserves available for the area.

1.5 Format of thesis presentation

The thesis is presented in three parts. The first part (this chapter and chapter 2) is introductory. The second chapter provides a synoptic geological and geochemical characterization of the Kwandonkaya complex. Analytical data and relevant explanations that are complementary to the main parts are given in appendices.

The second part of the thesis (chapters 3 to 6) covers the chemical characteristics of the rock-forming minerals and their role in the magmatic to postmagmatic evolution of the complex. Chapter 3 deals with the principal ferromagnesian minerals (olivine, pyroxene, and amphibole) within subvolcanic rocks in cone sheets and ring dykes. The patterns of chemical variation are

Fig. 1.2: a) Photo mosaic of aerial photographs of the southwestern part of topographic sheet 148. This includes the Kwandonkaya complex, shown in outline. It shows a prominent NNE-SSW trend of lineaments, with the lineaments perpendicular to this direction being less continuous. The lineaments represent a set of joints whose orientation indicates the main direction of extension upon cooling and unroofing. Such fractures are passageways for the percolation of hydrothermal fluids, and meteoric as well as surface waters. Jos is about 43 km to the southeast, and Bauchi is 75 km to the northeast, along the Jos to Bauchi road. The Ziem peak (at 1400 m above sea level), in the eastern part of the complex, is the highest point. Scale: 1 cm represents 1.4 km (or 1:140,000)

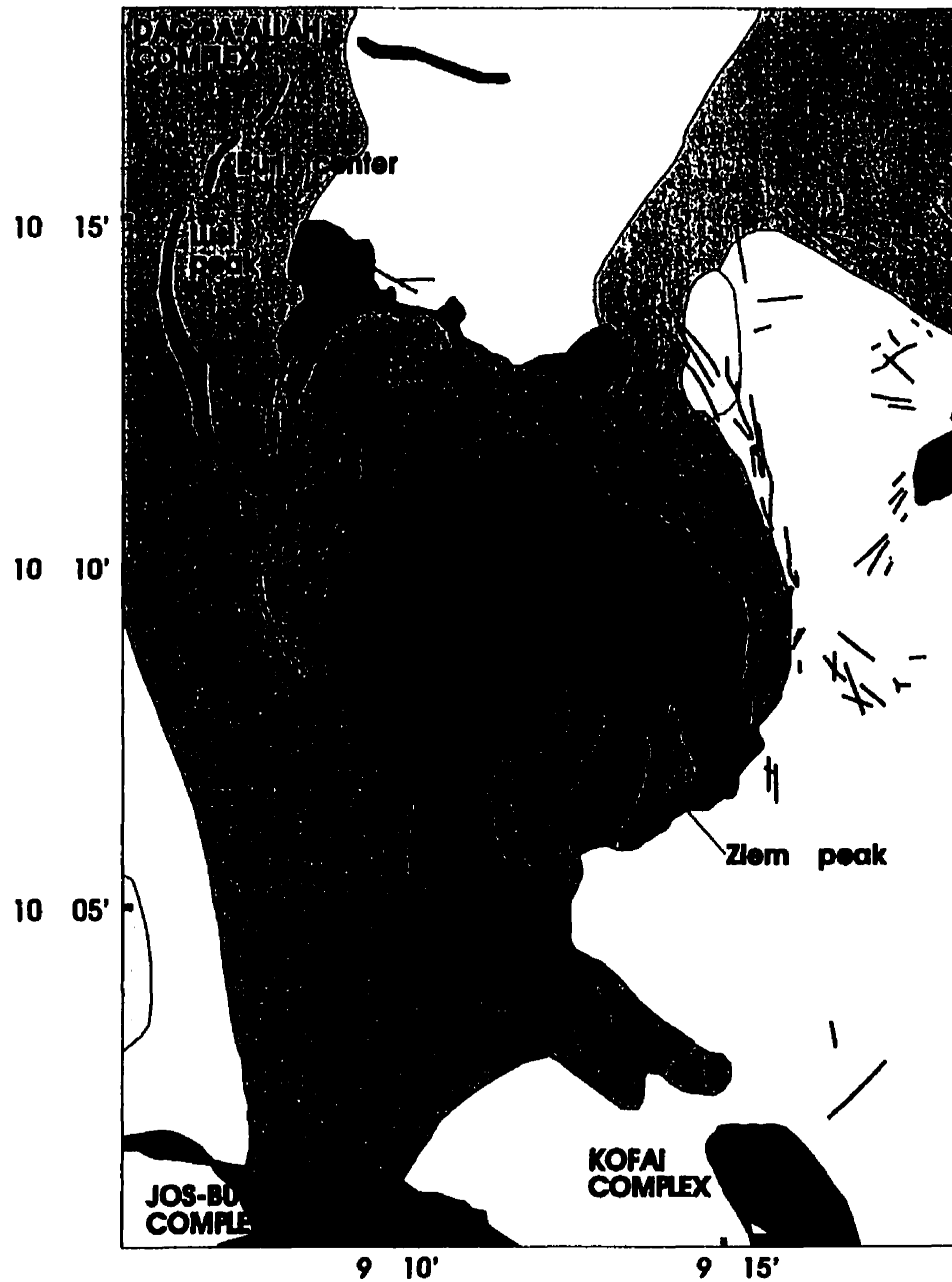
b) The western margin of Kwandonkaya complex, i.e., the side toward the Jos plateau (about 150 m above the Bauchi plains in the vicinity of the Kwandonkaya complex).

c): The eastern side of Kwandonkaya complex, as viewed from the Bauchi plain. The Ziem peak (650 m above the Bauchi plains) is indicated by an arrow in this photograph and the previous one.





Fig. 1.3: A simplified geological map of Kwandonkaya complex (boxed area in Fig. 1.1). The information needed to compile this map was partly taken from Buchanan (1971), and supplemented by field mapping in the summers of 1989 and 1990.

**EXPLANATION****MESOZOIC YOUNGER GRANITES**

G: Post-joint greisens

/ Basalt dyke

■ Fine- to medium-grained biotite granite

■ Coarse-grained biotite granite

■ Porphyritic biotite microgranite

■ "Homblende"-biotite, biotite granite porphyry, +/- fayalite syenogranite

■ Hedenbergite fayalite granite porphyry

■ Rhyolite porphyry

/ Felsite porphyry, granite porphyry

■ Gabbro

PAN-AFRICAN BASEMENT COMPLEX

□ Biotite homblende granite

■ Gneissic granite

□ Migmatite

1 0 2 4 6 8 km

used to document the magmatic evolution, and the products of alteration of these minerals are also employed to document the pattern of deuteritic alteration. In chapter 4, the mica in both subvolcanic and plutonic rocks is compared, and also used as an indicator of magmatic to postmagmatic evolution. In particular, halogen fugacities, inferred from mica compositions, are related to the mineralization potential of the granites. In Chapter 5, I present data on the alkali feldspars, used as monitors of degree of evolution via correlation between the structural state and the degree of subsolidus equilibration of their host rock. In Chapter 6, I present petrographic and SEM studies of accessory minerals and an assessment of their petrogenetic significance.

In the third part (Chapters 7 and 8), the possible processes that account for the observed geochemical variations in the complex are assessed. In Chapter 7, features of magma mixing and the role of hybridization in the observed geochemical variation in the complex are assessed. In Chapter 8, the composition of the least evolved felsic rock is used to assess whether or not fractionation of liquidus phases can account for the observed geochemical variations. Furthermore, a variety of protoliths are assessed to identify the most likely source-rock composition for the least evolved felsic rock composition in the complex.

The final part (Chapter 9) presents a summary of the scientific inferences made on the basis of the data presented. References cited in each chapter are listed at the end of the thesis.

1.6 Statement on the originality of the research

The research was conducted by me; wherever necessary, relevant cited work is fully acknowledged. This study represents the first detailed characterization of the Kwandonkaya complex. Studies of the compositions of mica and of the structural state and composition of constituent feldspars are the most detailed up to now, of the subvolcanic - plutonic rocks in the NYG. The use of chemical data on the mica to infer halogen fugacities is a first attempt to relate

the magmatic to postmagmatic evolution of NYG complexes to the changes in their volatile content. The study of mafic enclaves and their implication to the role of mixing of magmas also represents the first effort in this direction. The model of evolution, involving partial (batch) melting and in situ fractionation, represents the first semi-quantitative assessment of the most likely mode of evolution of the Kwandonkaya complex (and, by inference, other NYG complexes).

Chapter 2: Field geology, petrography, and geochemical characterization.

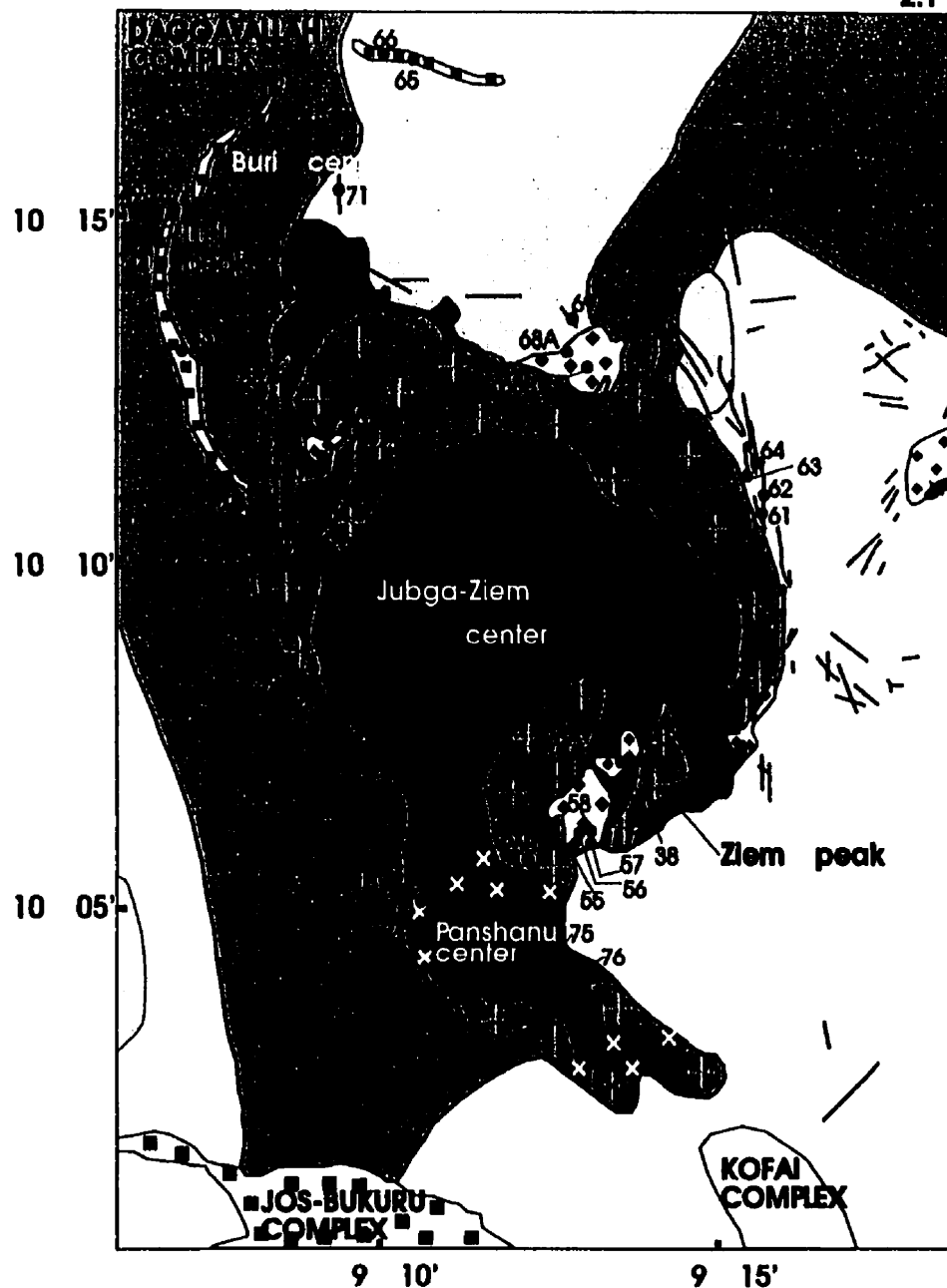
2.1 Introduction

The Kwandonkaya igneous complex is located on the eastern edge of Jos plateau, in Nigeria (Fig. 1.1). It forms part of a chain of overlapping volcano-calderas that include Zuku, Tongolo, and Dagga-Allah complexes (Fig. 1.1). This series of NNW-SSE-trending complexes repeats a province-wide pattern of distribution of centers (Bowden et al. 1987). Predominantly peralkaline and volcanic rocks are restricted to the northern complexes, whereas peraluminous intrusive rocks dominate in the southern part of the chain.

The Kwandonkaya complex was first described by Bain & Russ (1923); they commented on the highly variable texture of the granite and implied that recent unroofing had exposed the apical part of the pluton. The complex was resurveyed during wartime efforts to locate tungsten resources (Jacobson & Jacques 1944), and an updated geological map was published from the field survey started by R. Black in 1956, and completed by Buchanan (1971). Imeokparia (1984) undertook a geological-geochemical study of the complex, but did not include the Buri ring dyke and cone sheets in the Jubga-Ziem center (Fig. 2.1). In this chapter, field-based studies, petrography, and geochemical information are used to provide preliminary interpretation of the evolution of the complex. Features of magma mixing, briefly reported previously in other complexes (Turner 1963), are documented for the first time in the Kwandonkaya complex. A structural model is proposed to explain the evolution of the Kwandonkaya complex. This study is part of an ongoing investigation of the geology, geochemistry and mineral chemistry of the Kwandonkaya complex, intended to give further insight into the evolution of metaluminous to peraluminous granites in the A-type granite province of Nigeria.

Fig. 2.1: A geological map of Kwandonkaya complex (boxed area in Fig. 1.1). The map is modified from that of Buchanan (1971). The samples on which this investigation is based are located on the map, and were taken from all the map units.

2.1



• Sample location

MESOZOIC GRANITES

G: Grelsens

▨ Basalt dyke

■ Fine- to medium-grained biotite granite

▨ Coarse-grained biotite granite

⊗ Porphyritic biotite microgranite

◊ "Homblende"-biotite, biotite granite porphyry, +/- fayalite syenogranite

▣ Hedenbergite fayalite granite porphyry

▣ Rhyolite porphyry

▨ Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

□ Biotite homblende granite

▨ Gneissic granite

□ Migmatite

1 0 2 4 6 8 kms

2.2 Geological setting of the NYG complexes

The Kwandonkaya complex is one of the stocks of within-plate A-type granites emplaced during Mesozoic uplift (van Breemen & Bowden 1973, Bowden 1985). These ring complexes exhibit a range in age of emplacement (van Breemen & Bowden 1973). In a province-wide review of radiometric studies, Rahaman et al. (1984) showed that emplacement occurred in three 50-60 Ma cycles: complexes in Niger were emplaced during Ordovician-Silurian and Devonian-Carboniferous times; and those in Nigeria (Fig. 1.1) were emplaced during the Triassic-Jurassic (244 - 141 Ma). Changes in the direction of movement of the African plate are considered to have provoked reactivation of north-south-trending shear zones (Bowden et al. 1987, Bonin 1990) developed during a Proterozoic (Pan-African) ensialic tectonothermal event (Kröner 1981), and led to locally incipient within-plate rifting. This caused conjugate oblique (NNW-SSE and NNE-SSW) sets of transcurrent faults, originally under compression, to open and propagate as tensional faults, causing the release of pressure, a channelling of volatiles, partial melting and generation of parental magma(s) for the NYG, and the localization of intrusive centers (Bowden et al. 1987, Bonin 1990). The NYG intrusive complexes represent eroded volcano-caldera complexes exposed in their root zones, at subvolcanic (epizonal) levels (Turner & Bowden 1979, Bowden 1985). In fact, volcanic rocks are preserved within inner parts of ring dykes owing to cauldron subsidence in some complexes. In the Kwandonkaya complex, no volcanic rocks were encountered, owing to the deeper levels of erosion. The age, geological setting, structures, and rock types in the Kwandonkaya and other NYG complexes are comparable to those in the White Mountain Magma series (Greenwood 1951, Eby 1990, Eby et al. 1992).

2.3 Field relations

The Kwandonkaya complex is composed of three overlapping intrusive centers that have

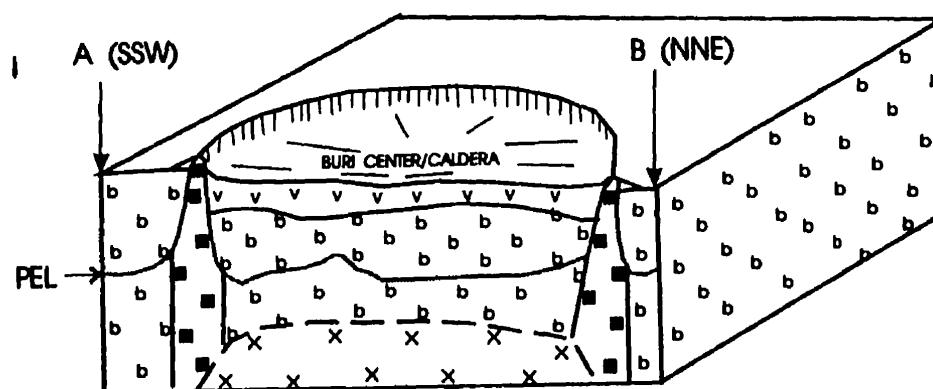
a NNW-SSE alignment: Buri, Jubga-Ziem, and Panshanu (Figs. 2.1). This direction subparallels that documented for the multicentered Sara-Fier complex (Turner 1963) (Fig. 1.1). The Buri center (northernmost) is comprised mainly of a discontinuous, subcircular ring-dyke with a diameter of 10 km, separated from the main mass of the Kwandonkaya hills (i.e., the Jubga-Ziem center) by an intervening screen of Pan-African basement gneisses and granites. The Buri ring dyke (made up of rhyolite porphyry and granite porphyry) rises up to 150 m (Lusi peak) above the plains, which are underlain by the poorly exposed Pan-African basement. The Jubga-Ziem intrusive center, the main one, is composite, comprised of different textural variants of biotite granite, with minor relics of its ring dyke preserved along the margin, and as roof pendants. The main mass of the Jubga-Ziem center attains 650 m (Ziem peak) above the migmatite gneiss complex of the Bauchi plains; it has a north-to-south diameter of 12 km. Swarms of arcuate cone sheets composed of felsite porphyry characterize the northwestern and northeastern areas, and define a subcircular outline beyond the main intrusive mass of the Jubga-Ziem center (Fig. 2.1). The Panshanu center is a curvilinear composite intrusive body of about 8x4 km, with minor cone-sheets to the east of this body. This third center is relatively lower-lying compared to the Jubga-Ziem center, from a few metres to over 210 m above Bauchi plains. The western contact (toward the Jos Plateau) is exposed at a higher altitude (900 m) than the eastern contact, toward Bauchi plains (about 750 m above sea level).

2.4 Distribution of map units

A geological section (Fig. 2.2a) provides a summary of relations between major rock units in Kwandonkaya complex. Present work shows the complex, including the Buri ring-dyke, to cover about 196 km² (Fig. 2.1). Cross-cutting relationships indicate a relative migration of intrusive centers: the Buri center is truncated by the Jubga-Ziem center, and the latter is cut by

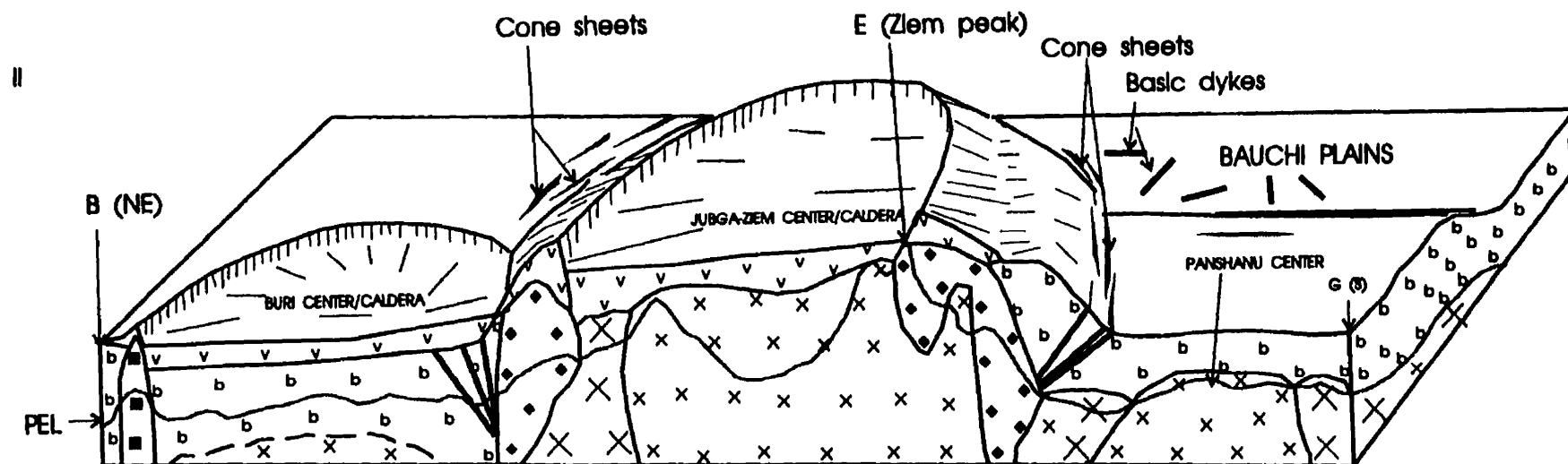
Fig. 2.2a: Schematic diagram illustrating the structure of Kwandonkaya complex. i) A northeast (B) - southwest (A) section across the Buri center. ii) A north (B) - south (A) section across the whole of the Kwandonkaya complex. Symbols used in this figure conform to the legend used in Fig. 2.1.

2.2a



Horizontal Scale 1:165,000

PEL = present level of erosion
 X Coarse-grained biotite granite
 x Fine- to medium-grained biotite granite
 ♦ Jubga-Zlem ring dyke
 ■ Buri ring dyke
 v Volcanic rocks (inferred)
 b Pan-African basement complex



Horizontal Scale 1:170,000

the Panshanu center. The sequence of description of rock units is: Buri ring dyke, Jubga-Ziem (cone sheets and ring dyke), and plutonic rocks in both Jubga-Ziem and Panshanu centers.

Rhyolite porphyry (4% of the Buri ring dyke exposure) is limited in occurrence to the Buri center, in exposures north of Lusi peak along an extension of the main granite porphyry ring dyke (Fig. 2.1). The rhyolite porphyry exposures, with a striking development of enclaves (20 to 40% of the rock), occur in a relatively lower-lying ridge (about 10 to 30 m above the plains), compared to the granite porphyry in the same ring-dyke. Hedenbergite fayalite granite porphyry constitutes the larger part (96%) of the Buri ring dyke. This rock unit has sharply defined contacts with the basement, and an irregular distribution of enclaves (up to 20% in localized areas). The Buri ring dyke originated by intrusion of degassed crystal-rich felsic magma along a ring fracture, most likely at a stage of intracaldera volcanism (Fig. 2.2).

The cone sheets vary from less than one meter (individual sheets) to swarms covering a width of over 100 m. The rocks range from felsite porphyry to granite porphyry, and the variants may occur as bands in a single cone sheet. Hedenbergite hastingsite granite porphyry and hastingsite granite porphyry occur as minor intrusive bodies in low-lying ridges, in association with the swarms of felsite porphyry. Overall, the cone sheets are considered the oldest units of the Jubga-Ziem center, since they developed as a product of positive magmatic pressure (cf. Ike 1983), by the injection of small volume of melts into a cone-shaped system of fractures (Fig. 2.2).

The emplacement of the Jubga-Ziem ring dyke, which cross-cuts the southern part of the Buri ring dyke, represents the next event. Hypersolvus (fayalite) syenogranite, "hornblende"-biotite granite porphyry, and biotite granite porphyry are preserved in remnants of this ring dyke in marginal parts of the Jubga-Ziem center (Fig. 2.1, 2.2a). The term "hornblende" is used for calcic amphibole that is further characterized in Chapter 3. This usage conforms with previous descriptions of this rock type in the Kwandonkaya complex and NYG province. The hypersolvus

(fayalite) syenogranite occurs in inner parts of some bodies of "hornblende"-biotite granite porphyry, with biotite granite porphyry capping such bodies. Typically, the "hornblende"-biotite granite porphyry grades to drusy variants along contact zones (i.e., near the contact with the basement complex or biotite granite). The "hornblende"-biotite granite porphyry occurs in direct contact with most of the rock types (Fig. 2.1), whereas biotite granite porphyry forms the upper parts of the relics of the Jubga-Ziem ring dyke.

Coarse-grained biotite granite occurs as arcuate units along margins in the Jubga-Ziem and Panshanu centers. In the Jubga-Ziem center, the exposures of such granite define a subcircular outline whose continuity is breached by cross-cutting, fine- to medium-grained biotite granite. In the Panshanu center, the coarse-grained biotite granite also occurs along the margin of the curvilinear composite body, in sharp contact with the basement, and it shows a similar relationship to fine- to medium-grained biotite granite, as in the Jubga-Ziem center (Fig. 2.1).

Medium- to fine-grained biotite granite occupies the central parts of the Jubga-Ziem center, in contact with the coarse-grained biotite granite, and the basement country-rock. The contact with the coarse-grained biotite granite generally is sharp, but is poorly defined in the case of Ziem ridge. The albitized variants of the biotite granite occur in upper sections or marginal parts of fine- to medium-grained biotite granite. Post-joint greisens occur predominantly in the porphyritic and fine-grained biotite granite (predominant trend: 285° to 330° , prominent trends: 050° to 075°), in the Jubga-Ziem center. The greisen zones occur either singly or in subparallel groups, producing superposed zones of alteration. These zones, millimetric to decimetric, have a central quartz vein, a quartz-rich greenish to grey greisen zone, and an outer zone of reddened granite grading into the porphyritic biotite granite host.

In marginal parts of the Panshanu center, mafic enclaves occur within porphyritic and fine-grained biotite granite. The enclaves vary from rounded to subangular fragments restricted

to localized zones that range from less than 20 m² to over 500 m², but the occurrence is extensive in the southernmost part of this center. In most cases, the enclaves do not show a preferred orientation or sorting, but seem randomly distributed.

2.5 Structural model

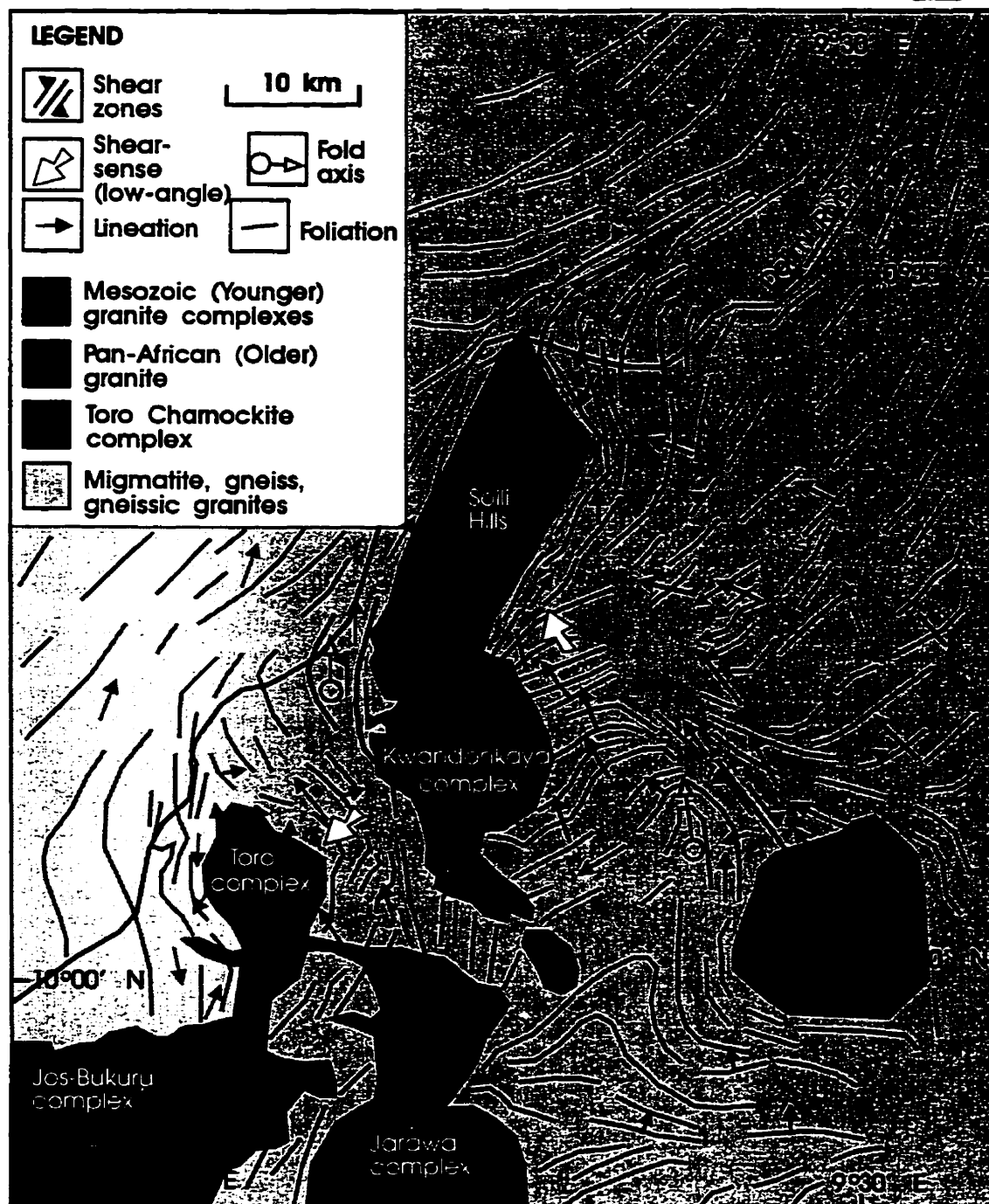
The effect of ring faulting and cauldron subsidence and the occurrence of the Kwandonkaya complex along the margin of the Jos Plateau, with attendant high rates of erosion, have combined to give a rare glimpse of three levels of juxtaposed volcano-caldera - plutonic centers. Figure 2.2a illustrates a composite section and structural model of the Kwandonkaya complex. The location of the complex in relation to structures in the surrounding basement complex is shown in Figure 2.2b (cf. Ferré et al. 1995). The NNW-SSE alignment of the Kwandonkaya complex and the southeastward direction of migration of magmatic centers were controlled by the direction of fold axes and shear zones (Fig. 2.2b).

The evolution of the complex, from field and photogeological evidence at the present level of erosion, involved:

i) Initiation of the Buri center by ring-fracturing, cross-cutting the Dagga-Allah structure. The ring fractures were filled with crystal-rich batches of felsic melt quenched at various degrees of crystallization, possibly as a result of cauldron subsidence and intracaldera volcanism. Rock types produced were rhyolite porphyry and hedenbergite fayalite granite porphyry in ring dykes; the volcanic edifice produced has been completely eroded. There is evidence, in form of mafic enclaves, that mafic melts were available, in different proportions, throughout the duration of the ring-dyke emplacement.

ii) Emplacement of cone sheets in a regime of positive magmatic pressure to produce curvilinear or radial small-volume intrusions including felsite porphyry, hastingsite ± hedenbergite

Fig. 2.2b: Structural map of the basement country rocks surrounding the Kwandonkaya complex (modified from Ferré et al. 1995). The structural patterns were interpreted from radar imagery followed by field studies that focused on the Solli hills lying immediately north of the Kwandonkaya complex. This map shows the alignment of the Kwandonkaya complex to be sub-parallel to fold axes and shear zones in the Pan-African basement. Ferré et al. (1995) did not make any inferences on the role of the basement structures on the localization of NYG complexes in their study area.



granite porphyries. The batches of magma that led to the different textural types seem to have been emplaced simultaneously.

iii) Initiation of ring fracturing in the Jubga-Ziem center, which cross-cuts the Buri center, could have been in part contemporaneous with the Buri ring dyke, located to the northwest. The ring-fracturing was most likely followed by intracaldera volcanism, with the only evidence preserved as ring dykes composed of hypersolvus (fayalite) syenogranite, "hornblende"-biotite and biotite granite porphyries. The intracaldera volcanism was followed by emplacement of a composite pluton of biotite granite. The coarse-grained biotite granite most likely originated by preferential accretion of nucleating crystals on walls of the magma chamber (either comprised of the granite porphyry ring dyke or Pan-African basement). The medium-grained biotite granite seems to have formed from a batch of relatively volatile-poor magma, as part of a composite hypabyssal pluton, along with the overlying fine-grained biotite granite that formed a volatile-rich carapace.

iv) In the Panshanu center, there is no evidence of volcanism or ring-dyke emplacement. However, there is evidence, in form of mafic enclaves and disrupted basaltic dykes, that the felsic magma reservoir was underlain by dacitic and basaltic magmas. Both dacitic and basaltic magmas were mixed to different extents and emplaced along margins of the reservoir of felsic magma. The occurrence of mafic enclaves in some rock types indicates the availability of mafic melt throughout the history of the Kwandonkaya complex, and lends credence to the structural models of Hildreth (1981) and Huppert & Sparks (1988) on the generation of felsic magma by lithospheric extension, injection and ponding of basaltic magma, and local mixing of the two magmas. The occurrence of basaltic volcanism is mainly supported by preserved basaltic map units in some NYG ring complexes such as Ningi-Burra (Turner & Bowden 1979), and Tibchi (Ike 1983). The structural model of the Kwandonkaya complex is similar to that presented by

Turner & Bowden (1979) for the Ningi-Burra complex, except that the Kwandonkaya complex studied has only three centers, and is eroded to a deeper level. Furthermore, the complex provides an insight to the transition from subvolcanic to apical plutonic levels of volcano-caldera complexes analogous to the features of the White Mountain Magma series (Greenwood 1951, Eby et al. 1992, Creasy & Eby 1993).

Petrographic descriptions are presented according to the relative sequence of emplacement (Table 2.1). The nomenclature follows Streckeisen (1973, 1976), with textures and mafic minerals used to distinguish the various units.

2.6 Petrography of the subvolcanic units

The subvolcanic rocks include porphyries (of rhyolite, felsite, and granite) that occur in ring dykes and cone sheets. These are termed volcanic-feeder intrusions by Kinnaird (1985, 1987). Table 2.1 provides a summary of petrographic characteristics of the main units, and formulae of accessory minerals are given in the footnotes.

Rhyolite porphyry in the Buri ring dyke typically has the appearance shown in Fig. 2.3a. The sequence of crystallization seems to be: chevkinite, zircon, ilmenite, fayalite (<0.5%), hedenbergite (1 - 5%), sanidine_{ss} (20 - 25%), β -quartz (now α -quartz) (15-25%), and fine-grained groundmass (40 - 60%). The groundmass is composed of microcrystalline to locally spherulitic intergrowths of feldspar and quartz: the assemblage probably represents devitrified glass. Examples of *hedenbergite fayalite granite porphyry* in the same ring dyke are distinguished from other cases of granite porphyry in containing mafic enclaves. The sequence of crystallization, in this case, is: chevkinite (0.2%), zircon (0.3%), fayalite (<0.5%), hedenbergite (3 - 5%), sanidine_{ss} (40 - 45%), β -quartz (now α -quartz) (20 - 25%), and fine-grained groundmass (30 - 50%).

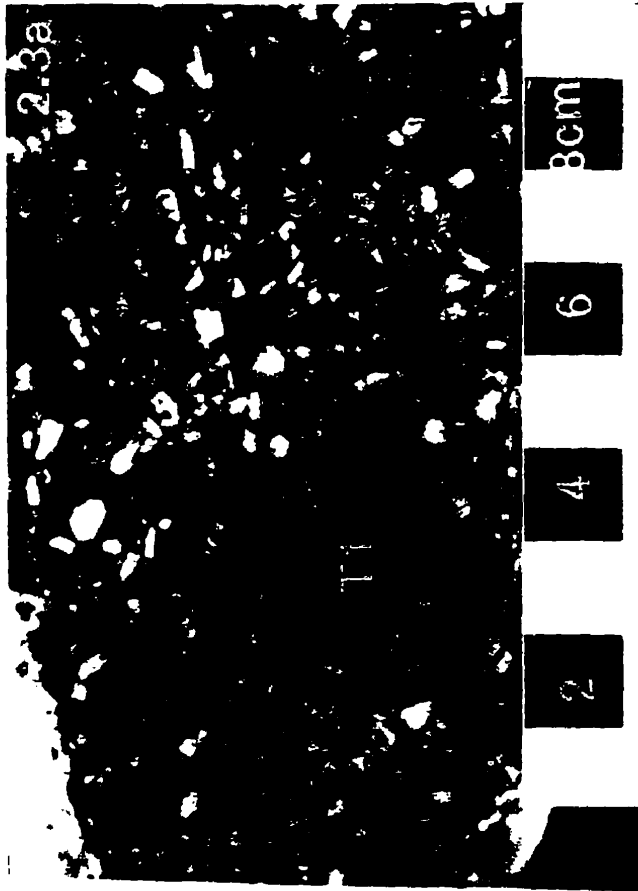
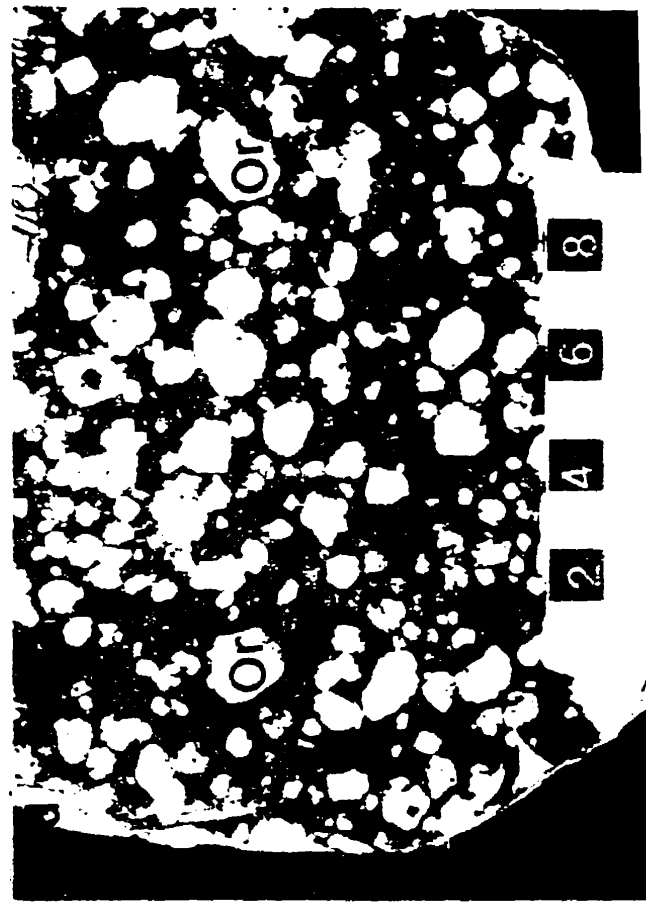
The felsite porphyry cone sheets of the Jubga-Ziem center have phenocrysts of quartz and

Table 2.1 PETROGRAPHIC FEATURES FOR THE MAIN ROCK TYPES OF KWANDONKAYA COMPLEX.

Rock Type	Texture	Essential minerals	Mafic Minerals	Accessory minerals
Rhyolite porphyry (B)	Porphyritic	Or, Qtz phenocrysts	Hd, Fa, & Fe-Ed microphenocrysts, (Fe-Act)	Ckt, Zrn, Ilm, Mag, Ttn, Ap, Bhl, Aln
Hd Fa granite porphyry (B)	Glomeroporphyritic	Or, Qtz phenocrysts	Hd, Fa, & Fe-Ed microphenocrysts, (Fe-Act, Bt, & Chl)	Zrn, Ilm, Ckt, Mag, Ap, Fl, Aln
Felsite porphyry, Hd Hs, & Hs Granite porphyries (CS: J-Z)	Aphanitic, porphyritic, spherulitic	Or, Qtz phenocrysts	Hs, $\pm$ Hd, $\pm$ Bt, (Chl), mostly leucocratic	Zrn, Aln, Ilm, Mag, Ckt, Ap, Ttn
Hypersolvus (Fa) syenogranite, Hbl-Bt, Bt granite porphyry (J-Z)	Drusy, aphanitic, glomeroporphyritic, coarse-grained	Or, Qtz phenocrysts	Fa, Hbl, & Bt phenocrysts	Zrn, Aln, Ilm, Mag, Fl, Mnz, Fst, Xnt, Tht, Fct
Coarse-grained biotite granite	Coarse granular	Or, Mc, Ab, Qtz	Bt, (Chl)	Ilm, Zrn, Tht, Aln, Mnz, Hem, Ap, & Mag
Fine- to medium-grained biotite granite	Fine- to medium-grained, porphyritic, drusy, miarolitic	Or, Mc, Ab, Qtz	Bt, (Chl)	Ilm, Mag, Fl, Toz, Mnz, Zrn Ap, Tht, Xnt
Albitized biotite granite	Fine granular or sugary	Mc, Ab, Qtz	Bt, (Chl)	Col, Cst, Zrn, Ilm, Mag, Mnz, Fl, Aln, Toz, Tht
Post-joint greisens	Even grained, or drusy	Qtz, Ab, Or, Micas	(Bt, Srp, Chl)	Cst, Toz, Col, Zrn, Tht, Rt, Aln, Hem

Mineral abbreviations used (mostly after Kretz 1983) are: **Ab**: albite, **Aln**: allanite {Ca(Ce,LREE,Y)FeAl₂(SiO₄)₃(OH)}, **Ap**: apatite {Ca₅[(Si,P)O₄]₃(OH,F)}, **Bhl**: britholite {(Ce,Y,Ca)₅[(Si,P)O₄]₃(OH,F)}, **Bt**: biotite, **Cam**: calcic amphibole, **Chl**: chlorite, **Ckt**: chevkinite {(LREE,Y,Ca,Th)₄(Ti₄Fe)Si₄O₂₂}, **Col**: columbite {(Nb,Ta)₂(Fe,Mn)O₆}, **Cst**: cassiterite {SnO₂}, **Fa**: fayalite, **Fct**: fluocerite {(La,Ce)F₃}, **Fe-Act**: ferroactinolite, **Fe-Ed**: ferroedenite, **Fl**: fluorite {CaF₂}, **Fst**: fergusonite {(Ce,La,Nd,Y)NbO₄}, **Hbl**: hornblende, **Hd**: hedenbergite, **Hem**: hematite, **Hs**: hastingsite, **Ilm**: ilmenite {FeTiO₃}, **Mag**: magnetite {Fe₃O₄}, **Mc**: microcline, **Mnz**: monazite {(La,Ce,Nd,Y,Ca,Th)PO₄}, **Qtz**: quartz, **Rt**: rutile {TiO₂}, **Srp**: siderophyllite, **Tht**: thorite {ThSiO₄}, **Toz**: topaz {Al₂SiO₄(F,OH)₂}, **Ttn**: titanite {CaTiSiO₅}, **Xnt**: xenotime {YPO₄}, **Zrn**: zircon {ZrSiO₄}. Secondary mafic minerals are in brackets. Other abbreviations: **CS**: cone sheets, **B**: Buri ring dyke, and **J-Z**: Jubga-Ziem ring dyke; other rock types occur in both Jubga-Ziem and Panshanu centers.

Fig. 2.3: Petrographic features of subvolcanic rocks. Ring dykes: a) Photomicrograph of rhyolite porphyry composed of euhedral to angular phenocrysts of orthoclase microperthite, rounded to subrounded, and embayed quartz phenocrysts. The amoeboid form of quartz indicates arrested growth forms, whereas the presence of broken quartz along with complexly rounded to subangular alkali feldspar phenocrysts indicate transport and attrition effects on entrained minerals within a moving body of melt, before its degassing and freezing along the volcanic feeder. Mafic enclave (T1) has inclusions of rounded to subrounded orthoclase microperthite. **Cone sheets: b)** Photomicrograph of euhedral phenocrysts of orthoclase microperthite (Or) in a granophyric to spherulitic groundmass of felsite porphyry. **c)** Photomicrograph of sector-zoned phenocryst of orthoclase microperthite in felsite porphyry with a relatively fine-scale granophyric core (Or-Qtz), a microperthitic intermediate zone (Per), and a relatively coarse granophyric margin (Or-Qtz). The phenocryst is set within microcrystalline quartz and feldspar; the unusual development of "granophyric phenocrysts" suggests rapid cotectic crystallization (Carstens 1983). **d)** A hand specimen of hastingsite granite porphyry, characterized by a glomeroporphyritic or synneusis texture formed by aggregation of alkali feldspar and quartz phenocrysts. Synneusis may be produced by turbulence in a crystal-rich melt (Stull 1979), and is also enhanced by inhomogeneous nucleation and preferential drifting together of phenocrysts of the same minerals. Orthoclase microperthite phenocrysts (Or) have a relatively colourless core and a turbid microperthite rim. The mantle of turbid microperthite is highly irregular and follows fractures, suggestive of an origin by replacement via a fluid phase. Width of field of view in all cases (photomicrographs) is 3.1 mm.



turbid orthoclase microperthite set in a granophyric to spherulitic groundmass (Fig. 2.3b, Table 2.1). The constituents of felsite porphyry include: orthoclase microperthite phenocrysts (20-30%), quartz phenocrysts (15-25%), and aphanitic groundmass (45 - 70%). In some samples, the feldspar phenocrysts themselves are granophyric, i.e., they contain alkali feldspar and quartz arranged in sectors (Fig. 2.3c). Although the quartz seems platy in some sectors, and thus possibly formed as tridymite later inverted to quartz (cf. Smith 1974), the shape of the quartz domains in the central sector suggests a series of rods of common orientation within a cellular crystal of alkali feldspar. Such "granophyric" phenocrysts are not a common feature in the Kwandonkaya complex. Also encountered in cone sheets are *hedenbergite hastingsite granite porphyry* and *hastingsite granite porphyry* (in which hastingsite is present as the only mafic mineral at only one locality: 40A). Both variants of granite porphyry have orthoclase phenocrysts with a clear microperthitic core and turbid albite-rich margins (fine white rim visible on original but seems largely lost here). The crystallization sequence seems to have been chevkinite, zircon, ilmenite, (+hedenbergite), sanidine_{ss}, quartz, hastingsite, and allanite.

The *hypersolvus (fayalite) syenogranite in the Jubga-Ziem ring dyke* varies from fine grained in chilled margins to coarse grained in the interior. The sequence of crystallization is: fayalite (1 - 3%), sanidine_{ss} (60 - 70%), quartz (20 - 25%), and interstitial micrographic material (2 - 5%). "*Hornblende*"-*biotite granite porphyry*, also in the Jubga-Ziem ring dyke and gradational with the syenogranite, is texturally variable from drusy or miarolitic to a nondrusy porphyry inward from the contact. The sequence of crystallization, in this rock unit, is: sanidine_{ss} (40 - 45%), quartz (30 - 35%), "hornblende" (2 - 5%), allanite (0.5 - 1%), biotite (< 1%), Fe-Ti oxides (<0.5%), interstitial groundmass (10 to 30%). *Biotite granite porphyry* is transitional to porphyritic biotite microgranite, and is composed of: orthoclase microperthite phenocrysts (35 - 40%), quartz (25 - 30%), biotite (1 - 3%), and a microcrystalline groundmass (20 - 45%).

2.7 Petrography of the plutonic rocks

These rock types predominate in the Jubga-Ziem and Panshanu centers, both being composed of different textural varieties of biotite granite (Fig. 2.1). Each center has three or more textural varieties that include coarse-grained (78 km²), and fine- to medium-grained biotite granites (117 km²).

The *coarse-grained biotite granite* differs from other biotite granites in having discrete grains of albite, a coarse grain-size (up to 2.5 cm), and preferential location toward the external contact of the pluton (Fig. 2.1). The rock unit contains orthoclase and microcline microperthite with intragranular albite (Table 2.1) and varies from quartz-rich alkali granite to syenogranite. The predominance of microperthitic alkali feldspar indicates a hypersolvus mineralogy, whereas the interstitial albite is suggestive of late-stage build up of volatiles to result in incipient, late-magmatic to subsolidus overprint and incipient development of a subsolvus texture. This rock unit does not show any evidence of post-joint alteration typical of other types of biotite granite.

The *medium-grained biotite granite* grades into *fine-grained biotite granite*. Textural variants of these rocks include: equigranular, porphyritic, drusy, and albitized types. The drusy variants are developed on marginal parts of the intrusions, with miarolitic cavities containing K-feldspar, dogtooth quartz, biotite, $\pm$ topaz, $\pm$ greenish beryl (aquamarine). Orthoclase microperthite occurs in the essentially hypersolvus biotite granites. In most other cases, the rock contains either microcline perthite, or grid-twinned microcline along with intra- and extragranular albite, i.e., a subsolvus texture produced by late-magmatic to subsolidus overprint on an initially hypersolvus mineralogy.

2.8 Minor rock types

The minor rock types include mafic enclaves within subvolcanic units in the Buri ring

dyke and plutonic biotite granite, as well as basaltic dykes.

Within the Buri ring dyke, *trachyandesite enclaves* in the rhyolite porphyry range in size from microscopic to over 10 x 25 cm, and the larger enclaves may contain xenocrysts of orthoclase microperthite, and quartz (Fig. 2.4a). These enclaves are composed of hornblende, biotite, alkali feldspar, ilmenite, apatite, and magnetite. *Microsyenodiorite enclaves* within the hedenbergite fayalite granite porphyry are composed of, in addition to the foregoing, ortho- and clinopyroxene. *Dacitic* and *andesitic enclaves* in marginal parts of the medium-grained biotite granite are globular to subangular (Fig. 2.4b). The enclaves are composed of oligoclase-andesine, hornblende, biotite, quartz, and accessory Fe-Ti oxides and apatite.

The basaltic dykes vary from 20 to 150 cm wide and intrude all rock units, except the coarse-grained biotite granite. The dykes vary from aphanitic to sparsely porphyritic types. The dominant phenocryst phases are plagioclase, clinopyroxene, and hornblende. Titanite, apatite, and Fe-Ti oxides are accessory minerals, whereas sericite, chlorite, epidote, and magnetite replace the primary minerals. The undisrupted basaltic dykes postdate granite emplacement, whereas the disrupted types are interpreted as synplutonic. A more detailed characterization of these units is presented in Chapter 7.

2.9 Altered rocks

The samples of *albitized biotite granite* are fine- to very fine-grained, and chalk white to hematite-speckled. The albite seems identical to that in the less altered zones, whereas microcline occur as relics only, indicating preferential replacement by albite. Biotite is chloritized to various extents. This rock type is associated with disseminated columbite-cassiterite mineralization (Fig. 2.5a).

Post-joint greisens develop a variety of textural types depending the degree of acid

Fig. 2.4: Minor rock types, with features indicative of magma mixing. **a)** Field photograph of rhyolite porphyry (QPr) in the Buri ring dyke, with dispersed enclaves of trachyandesite (T1 and T2); **b)** field photograph of enclaves in marginal part of fine- to medium-grained biotite granite. The rounded to subrounded enclaves are trachybasaltic, trachydacitic (T3), and andesitic (And).

Fig. 2.5. Evidence of mineralization in the biotite granite: **a)** a SEM-BSE image of zoned grains of columbite (Col) with a dark grey Nb- and Mn-rich core, and a lighter grey Ta- and Ti-rich margin, within albitized biotite granite. The columbite grains show evidence of four stages of growth. Cassiterite (Cst) grains are considered youngest, and are relatively homogeneous. **b)** A hand specimen of zoned greisen, with a quartz-rich inner part (Qtz-rich), a dark greenish grey quartz-mica greisen zone (Greisen), and a porphyritic brick-red, K-metasomatized host in the outer part (PGr) bearing phenocrysts of microcline (Mc).



0002 20KU 10PM MD39

Cst

Col

Cst

Col

1 cm
2
3
4
5

metasomatism, and quartz enrichment (Fig. 2.5b). Some of the greisen veins have drusy cavities lined by dogtooth quartz, microcline, minor albite, and hematite (Table 2.1). Anhedral quartz constitutes 40 to 80% of such greisen. Microcline is turbid, and ranges from 0 to 25%. Siderophyllite and annitic biotite both are present, as is chlorite.

2.10 Geochemical Characteristics

Representative samples were analyzed for major elements by X-ray fluorescence (XRF) of fused beads prepared from ignited samples. For the trace elements (Nb, Zr, Y, Sr, Rb, Pb, Th, U), pressed powder pellets were used. The concentration of sulfur was obtained using a LECO induction furnace and an ALPHA sulfur titrator. FeO was determined by ammonium metavanadate titration. Concentrations of the rare earth elements (REE) were obtained by inductively coupled plasma - mass spectrometry (ICP - MS) (solution mode). The average composition of each rock type is presented in Table 2.2, whereas the REE data are given in Table 2.3. The geochemical characteristics are discussed in the following order: major elements, trace elements, and rare earth elements.

2.10.1 Major elements

The DI (differentiation index = normative orthoclase + albite + quartz), and ASI [aluminum saturation index = molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$] of the rocks are used as indicators of degrees of evolution (Table 2.2). There are significant overlaps in both the DI and the ASI for the various rock types (Table 2): the Buri ring dyke ($92.2 \leq \text{DI} \leq 94.4$, $0.94 \leq \text{ASI} \leq 1.07$), cone sheets ($91.2 \leq \text{DI} \leq 94.9$, $0.99 \leq \text{ASI} \leq 1.08$), the Jubga-Ziem ring dyke ($93.1 \leq \text{DI} \leq 95.1$, $0.99 \leq \text{ASI} \leq 1.1$), biotite granites ($92.8 \leq \text{DI} \leq 95$, $1.05 \leq \text{ASI} \leq 1.1$), as well as albitized biotite granite ($\text{DI} = 94.4$, $1.07 \leq \text{ASI} \leq 1.17$). These values indicate highly evolved

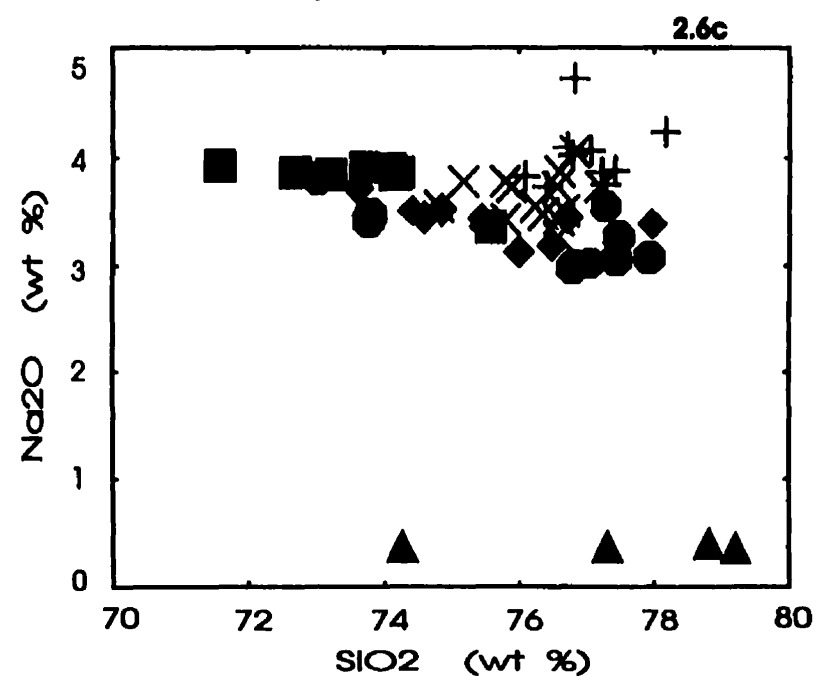
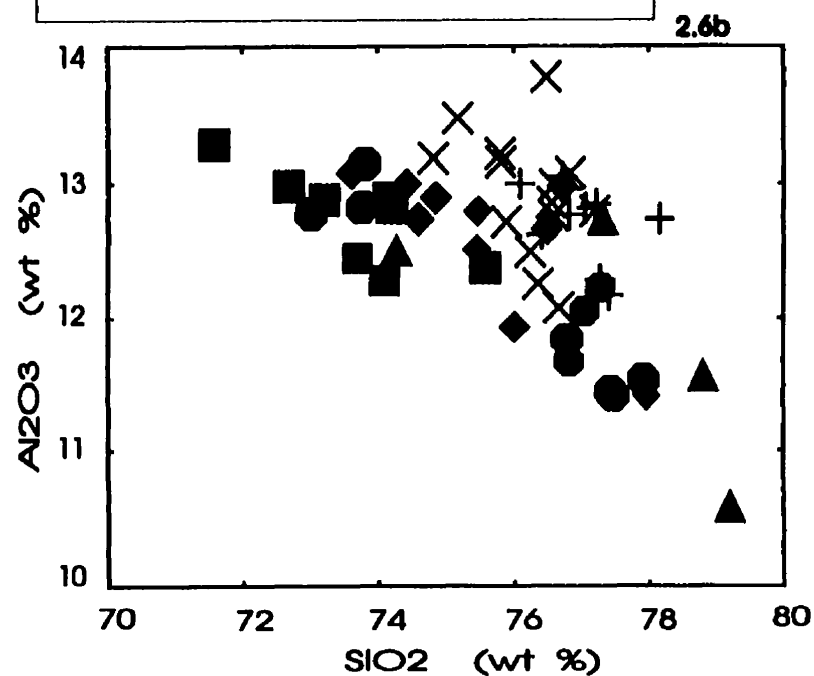
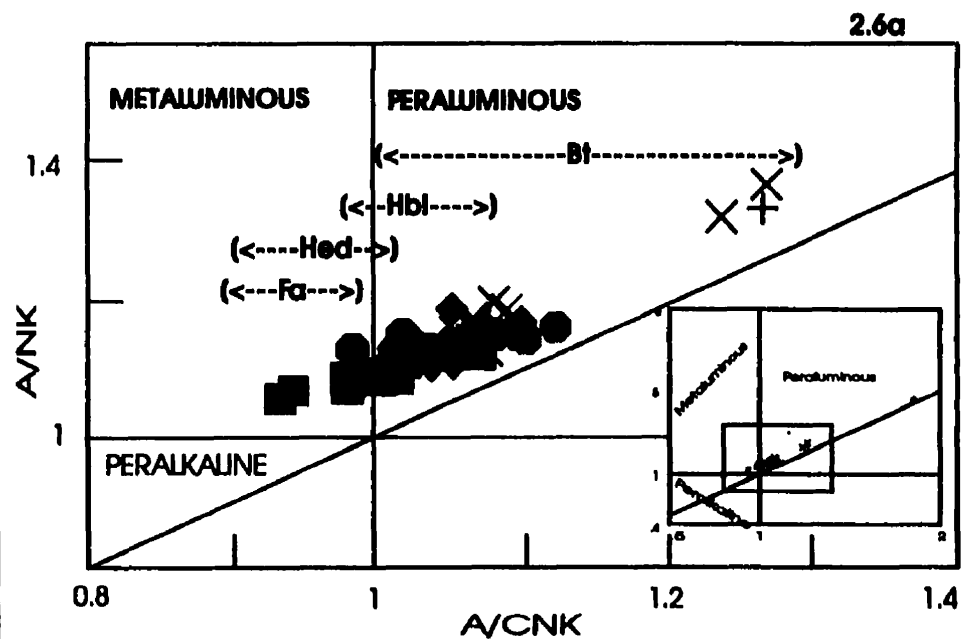
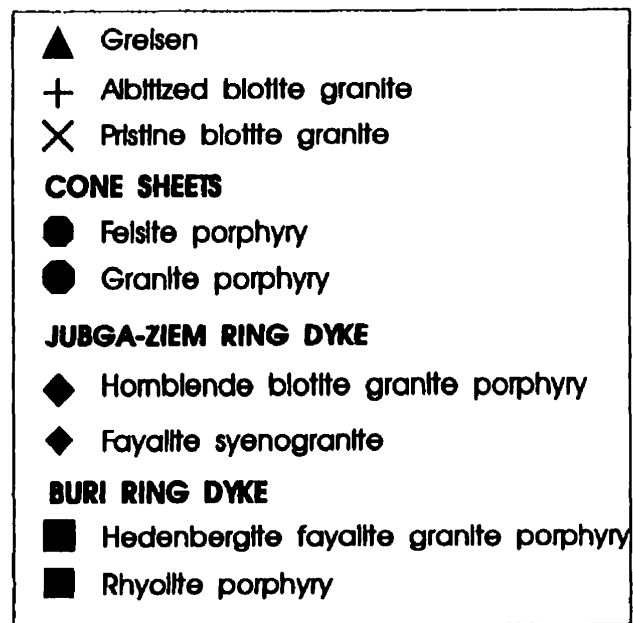
compositions that have a consistently narrow span of evolution, in terms of both the DI and ASI. The rock suite is also Fe-rich ($0.97 \leq \Sigma\text{FeO}/\text{MgO} \leq 1$).

The principal geochemical variation documented in the Kwandonkaya complex is the transition from metaluminous to peraluminous compositions (Fig. 2.6a). The Buri ring dyke shows a transition from metaluminous rhyolite porphyry (ASI = 0.96) to hedenbergite fayalite granite porphyry (ASI = 0.99). This is inferred to be due the fractional crystallization of fayalite, clinopyroxene, sanidine_{ss}, Ca-amphibole, chevkinite, and allanite. The reddened granite porphyry is peraluminous (ASI = 1.07). In this case, the peraluminous composition developed by postmagmatic alteration, which resulted in the destruction of mafic phases and alkali feldspar.

In cone sheets, which surrounds the main mass of the Jubga-Ziem center, there is a progressive increase in ASI from metaluminous hastingsite ($\pm$ hedenbergite) granite porphyry ($0.99 \leq \text{ASI} \leq 1.02$), through mildly peraluminous leucocratic granite porphyry (ASI = 1.05), to peraluminous felsite porphyry (ASI = 1.08). In this example, like in the case of the Buri ring dyke, there is no linear correlation between DI and ASI. This is consistent with some alkali loss by the cone sheets during subsolidus equilibration. In the ring dyke of the Jubga-Ziem center, the ASI seems to reflect the mafic mineral present: from fayalite syenogranite ($0.99 \leq \text{ASI} \leq 1.04$), through "hornblende"-biotite granite porphyry (ASI = 1.04); to biotite granite porphyry (ASI = 1.1). The transition from metaluminous to peraluminous compositions in this case seems to be related to the crystallization of fayalite, and Ca-amphibole ($\pm$ biotite, $\pm$ allanite). Loss of volatiles in marginal and apical parts of the ring dyke also probably contributed to enhance the peraluminous character due to the loss of alkalis.

Among the plutonic rocks, a notable overlap in the ASI is evident between coarse-grained biotite granite ($1.06 \leq \text{ASI} \leq 1.08$) and porphyritic and fine- to medium-grained biotite granite, including the host-rocks to the greisens ($1.05 \leq \text{ASI} \leq 1.1$). The superposed effects of buildup and

Fig. 2.6: Plots of A/NK vs. A/CNK , Al_2O_3 , and Na_2O vs. SiO_2 : **a)** $Al_2O_3/(Na_2O+K_2O)$ (molar basis) vs. $Al_2O_3/(CaO+Na_2O+K_2O)$ (molar basis) or aluminum saturation index (ASI) for rocks of the Kwandonkaya complex (Fa = fayalite, Hd = hedenbergite, Hbl = hornblende, and Bt = biotite); **b)** Al_2O_3 vs. SiO_2 (wt %); and **c)** Na_2O vs. SiO_2 (wt %).



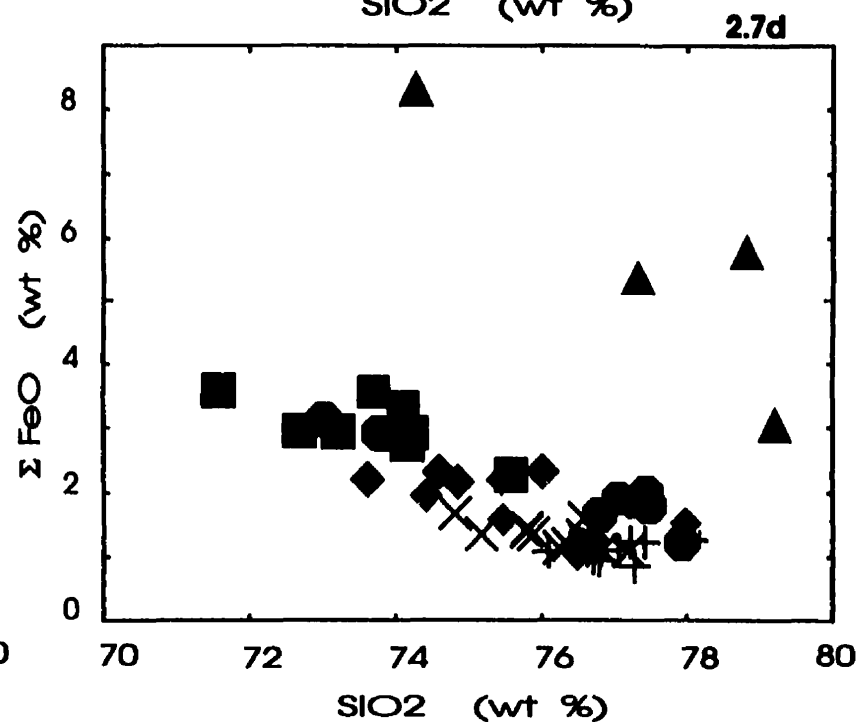
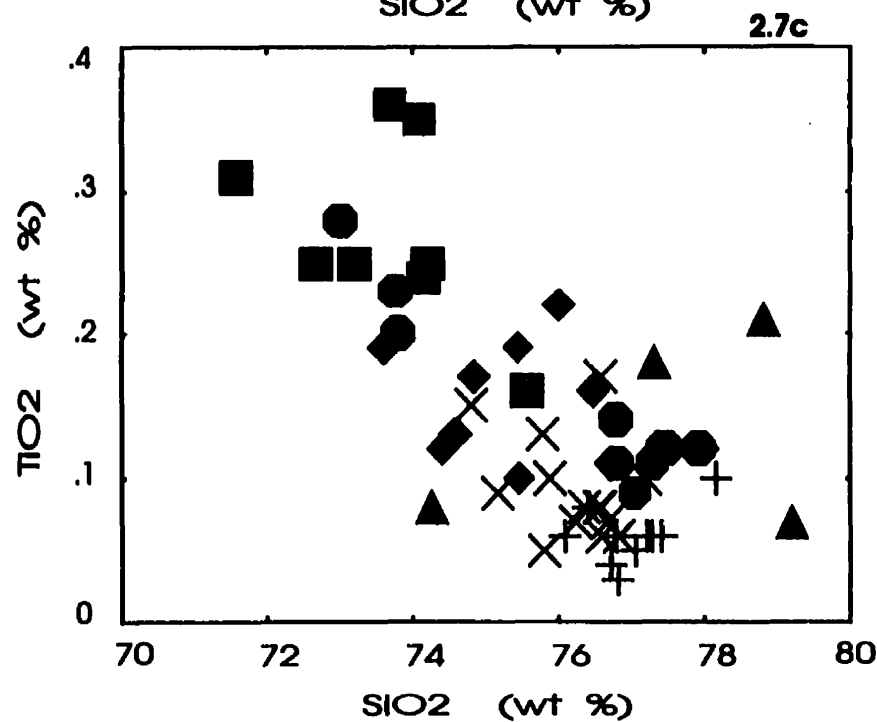
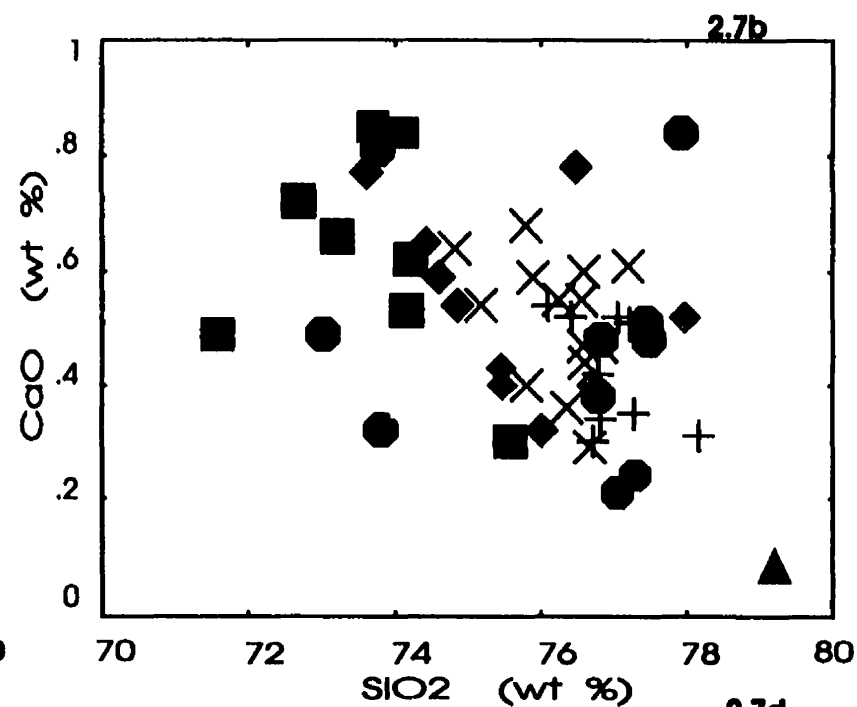
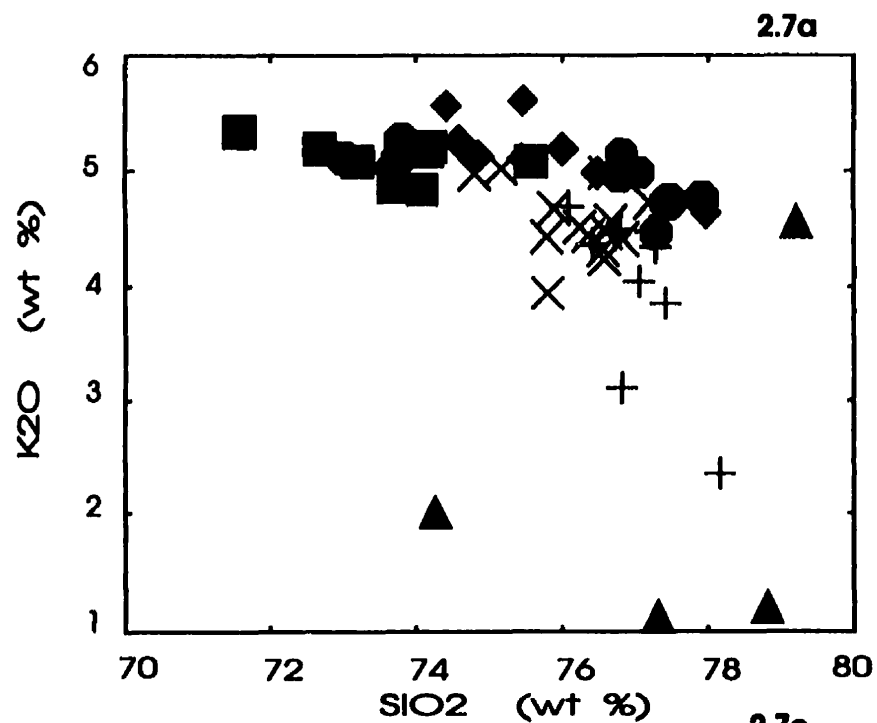
loss of volatiles and subsolidus modification most likely account for the development of peraluminosity in biotite granites in the NYG (cf. Martin & Bowden 1981). This is best illustrated in the albitized fine-grained biotite granite (relatively high aluminosity: $1.07 \leq \text{ASI} \leq 1.27$), and post-joint greisens (the most strongly aluminous units: $1.77 \leq \text{ASI} \leq 4.47$). The ASI of the greisens is two to four times that of the biotite granite host.

The SiO_2 content is lowest (for the whole suite) in the Buri ring dyke (72 to 74 wt.%), whereas both the cone sheets and Jubga-Ziem ring dyke have a bimodal distribution of SiO_2 contents (73 to 75%, and 75 to 77%) as displayed in Figure 2.6b. Some of the cone sheets have very high silica contents, comparable to levels in high-silica rhyolites (Christiansen et al. 1986). The biotite granites and their altered variants also have high SiO_2 contents (75 to 78%), which reflect the effects of subsolidus addition of quartz. The greisens generally have high silica contents ($\geq 75\%$), also as a result of silicification.

The Al_2O_3 decreases on average from the Buri to Jubga-Ziem subvolcanic rocks (Fig. 2.6b), with the biotite granites and their altered variants having a slightly broader range of Al_2O_3 compared to the subvolcanic rocks, possibly a reflection of selective mobility of the alkalis. A plot of Na_2O versus SiO_2 (Fig. 2.6c) shows the Buri ring dyke to be characterized by slightly higher Na_2O than the Jubga-Ziem cone sheets and ring dyke. The biotite granites have a slightly higher Na_2O than the coeval subvolcanic rocks of the Jubga-Ziem center. Whereas the albitized biotite granites have the higher levels of Na, the greisens are characterized by consistently very low levels of Na (<1 wt% Na_2O). The latter observation establishes the importance of localized Na-enrichment during albitization and Na-depletion during greisenization.

A plot of K_2O versus SiO_2 (Fig. 2.7a) indicates the subvolcanic rocks to be characterized by a similar range of K_2O , whereas the plutonic rocks contain slightly less K. Albitized biotite granite is characterized by a lower K_2O content that overlaps with values typical of the biotite

Fig. 2.7: Plots of K_2O , CaO , TiO_2 , and ΣFeO vs. SiO_2 , all in wt %: a) K_2O vs. SiO_2 , b) CaO vs. SiO_2 , c) TiO_2 vs. SiO_2 , and d) ΣFeO vs. SiO_2 . Symbols are the same as in Fig. 2.6.



granite. The greisens also have a lower level of K (mostly less than 2 wt% K_2O), again the same pattern as with Na. The range of CaO is similar in both subvolcanic rocks and plutonic rocks and their altered variants (Fig. 2.7b). This is consistent with the absence of any significant effect of fractionation of Ca-rich phases (e.g., plagioclase, clinopyroxene, and Ca-amphibole). Most of the greisen samples do not have detectable amounts of Ca.

TiO_2 content, depicted in Figure 2.7c, decreases from the Buri ring dyke via the subvolcanic rocks in cone sheets, and Jubga-Ziem ring dyke to the biotite granite. This suggests fractionation of Ti-rich accessory and mafic minerals (ilmenite, magnetite, biotite and Ca-amphibole). The albitized biotite granite contains the lowest average TiO_2 . In this case, the mobility of Ti in the subsolidus-hydrothermal stages is suggested, possibly due to increases in fluorine content (Keppler 1993). The greisen zones are characterized by enhanced Ti contents, compared to albitized biotite granites, possibly due to hydrothermal enrichment.

The ΣFeO decreases from subvolcanic to plutonic rocks (Fig. 2.7d), in a pattern similar to that of TiO_2 . This further stresses the role of fractionation of Fe-Ti oxides. There are no definite trends in the FeO and Fe_2O_3 contents of the subvolcanic and plutonic rocks (Fig. 2.8a, b). The absence of any definite trends in the FeO and Fe_2O_3 contents of the rock suite suggests the local importance of oxidation of Fe^{2+} to Fe^{3+} in the late-magmatic to subsolidus stages. It seems that Fe_2O_3 increased during the cooling of some subvolcanic rocks (Fig. 2.8b), presumably because of loss of H_2 . The albitized biotite granite contain Fe_2O_3 contents similar to that of their host rocks, and the greisens are enriched in Fe_2O_3 relative to other biotite granites (Fig. 2.8b). Albitization appears to have occurred under reducing conditions (where fluid compositions were buffered by the host biotite granite), whereas greisenization clearly occurred in oxidizing conditions, as hematite is prominent. The MnO distribution is similar to that of FeO in the subvolcanic rocks; in plutonic rocks and their albitized equivalents, MnO contents are close to or

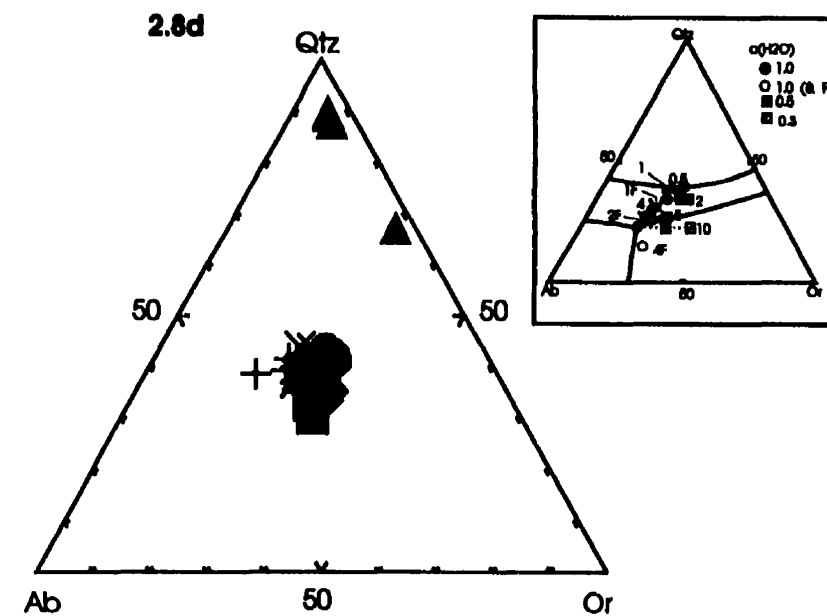
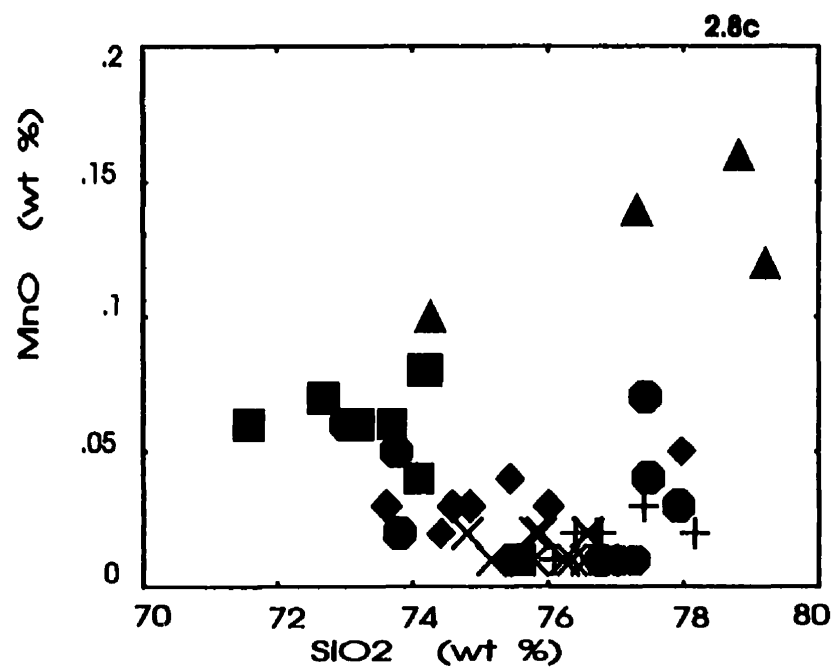
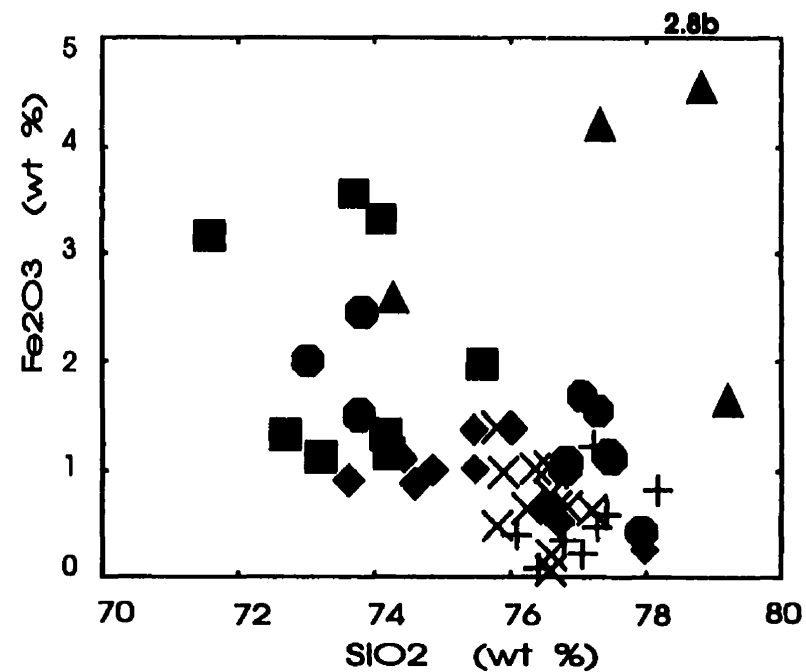
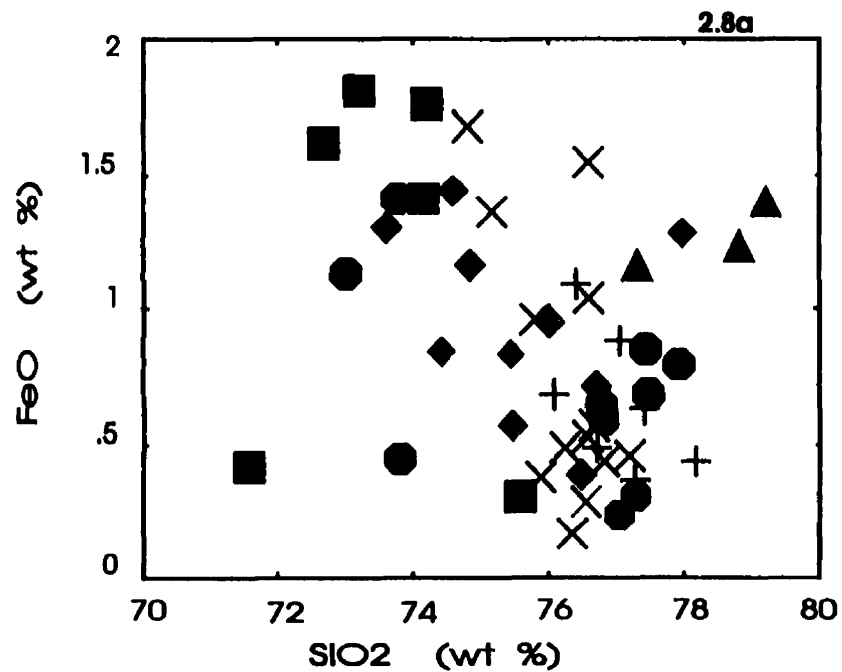
below the detection limit (Fig. 2.8c). Greisen samples, on the other hand, are the most enriched in Mn (and ΣFeO : Fig. 2.7d). The phosphorus content is at or below the detection limit in this suite. This suggests that apatite was a residual phase in partial melting events that produced the Kwandonkaya suite.

In a triangular plot of normative quartz-albite-orthoclase, rhyolite and granite porphyries define a linear trend, along a path consistent with water-undersaturated crystallization (Fig. 2.8d). Metaluminous rhyolite and granite porphyries in the Buri ring dyke, bearing fayalite and hedenbergite, seem to have evolved transitionally to "hornblende" $\pm$ biotite granite porphyry, between the paths of crystallization inferred for $0.3 \leq a(\text{H}_2\text{O}) < 0.5$, between 2.5 to 0.5 kbar (Fig. 2.8d, Appendix II.2), as defined by Ebadi & Johannes (1991). Felsite porphyry and granite porphyries in the cone-sheets seems to have crystallized at a minimum of about 0.5 kbar $P(\text{H}_2\text{O})$, from a melt on the quartz-alkali feldspar cotectic. The biotite granites, on the other hand, plot away from the 0.5 kbar minimum, slightly displaced toward the albite apex, consistent with loss of K and Si via a fluid phase. The greisens plot close to the quartz-orthoclase join and are highly enriched in normative quartz, owing to acid metasomatism and accompanying relative enrichment in silica. From Figure 2.8d, the variations in normative composition can be explained by: 1) fractionation of liquidus phases, 2) increasing ASI, 3) increasing water (volatile) contents, and 4) open-system subsolidus changes. These factors are discussed further in later sections.

2.10.2 Trace elements

The trace element data are presented in Table 2.2. Zr decreases from the subvolcanic rocks of the Buri ring dyke, via the Jubga-Ziem subvolcanic rocks, to plutonic rocks and their altered variants. The albitized biotite granites have the lowest Zr contents (Fig. 2.9a). In view of the inferred lack of mobility of Zr, the concentration of this element is selected as the abscissa

Fig. 2.8: Plots of FeO, Fe₂O₃, and MnO vs. SiO₂, and normative Ab-Qtz-Or plot: **a)** FeO vs. SiO₂ (wt %), **b)** Fe₂O₃ vs. SiO₂ (wt %), and **c)** MnO vs. SiO₂ (wt %), and **d)** normative albite-quartz-orthoclase for Kwandonkaya rocks [inset shows granite minima at different total pressures (P) at water-saturated conditions, i.e., $a(\text{H}_2\text{O}) = 1$ (from Tuttle & Bowen 1958, Luth et al. 1964); at P = 1 kbar, $a(\text{H}_2\text{O}) = 1$, and different wt.% F (Manning, 1981), for $a(\text{H}_2\text{O}) = 0.5$ and 0.3 at P = 2 kbars (Holtz et al. 1989, 1992a, b), and $a(\text{H}_2\text{O}) = 0.5$ & 0.3, P = 5 and 10 kbars (Ebadi & Johannes 1991): see Appendix II.1 for details from these sources; an enlarged version of the inset is shown in Appendix II.2]. Symbols are the same as in Fig. 2.6.



in a series of binary plots. The Sr content of the subvolcanic rocks is low and defines a spread of values: 10 to 45 ppm in the Buri ring dyke, and 5 to 75 ppm in the Jubga-Ziem cone sheets and ring dyke (Fig. 2.9a). The biotite granites ($3 \leq \text{Sr} \leq 35$ ppm) have a range of Sr contents similar to the Buri ring dyke, whereas Sr seems to have been removed from the albitized biotite granites ($3 \leq \text{Sr} \leq 10$ ppm). The average Ba content decreases from the Buri ring dyke via the Jubga-Ziem subvolcanic rocks to biotite granites and their altered variants (Fig. 2.9b). The Rb content is relatively constant in the Buri and Jubga-Ziem subvolcanic rocks (100 to 270 ppm), whereas the biotite granites have a higher range of Rb (260 to 600 ppm). Albitized biotite granites and greisens have an even higher average Rb content (Fig. 2.9c). A summary of the variations of Ba, Rb and Sr is provided in a triangular plot (Fig. 2.9d). The Kwandonkaya suite is relatively evolved, with the subvolcanic rocks in the field of normal granites, and biotite granites with a minor proportion of subvolcanic rocks in the field of strongly differentiated granites of El Bouseily & El Sokkary (1975). The hydrothermally altered plutonic rocks are characterized by low levels of Sr and Ba, and are enriched in Rb. It seems that alteration emphasized the trend of magmatic fractionation in the case of Rb, which behaved, predictably, as an incompatible element.

The Ni content of the different subvolcanic rocks (10 to 30 ppm) is low, as expected, whereas the biotite granites and their altered equivalents have a slightly higher Ni content (Fig. 2.10a). The Ni may either have been added to the biotite granites by diffusive exchange of felsic magma with a mafic magma, or by a hydrothermal fluid interacting with mafic material. The Pb content of both the subvolcanic and plutonic rocks is similar (2 to 10 ppm) and suggests a low input from the host basement complex (Fig. 2.10b), known to be Pb-enriched (Dickin et al. 1983, Kinnaird et al. 1985a, b, Kinnaird 1987). The greisen samples, on the other hand, are characterized by a high Pb content, suggesting a locally high flux of meteoric water that interacted with the basement complex. The Th and U contents vary in a pattern similar to Rb (Figs. 2.10c,

Fig. 2.9: Plots of the concentrations of Sr, Ba, and Rb vs. Zr (in ppm), and triangular plot of Ba-Rb-Sr: a) Sr vs. Zr, b) Ba vs. Zr, and c) Rb vs. Zr, and d) a triangular plot of Ba-Rb-Sr (I: diorites and gabbros, II: granodiorites, III: normal granites, and IV: strongly evolved granite. The fields are as defined by El Bouseily and El Sokkary (1975). Symbols are the same as in Fig. 2.6.

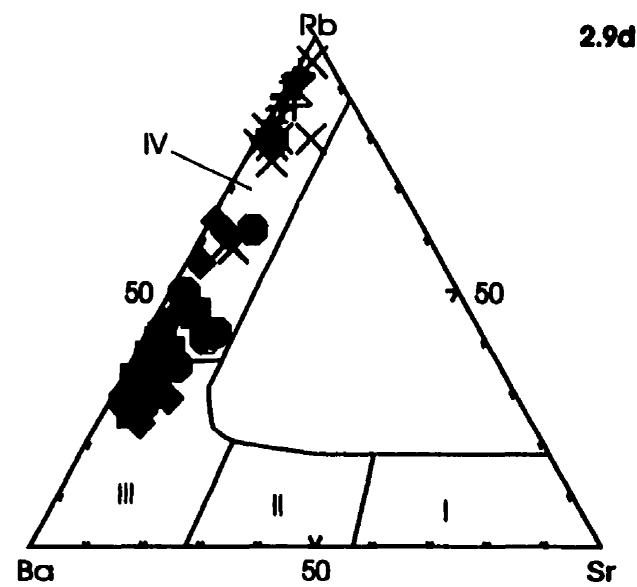
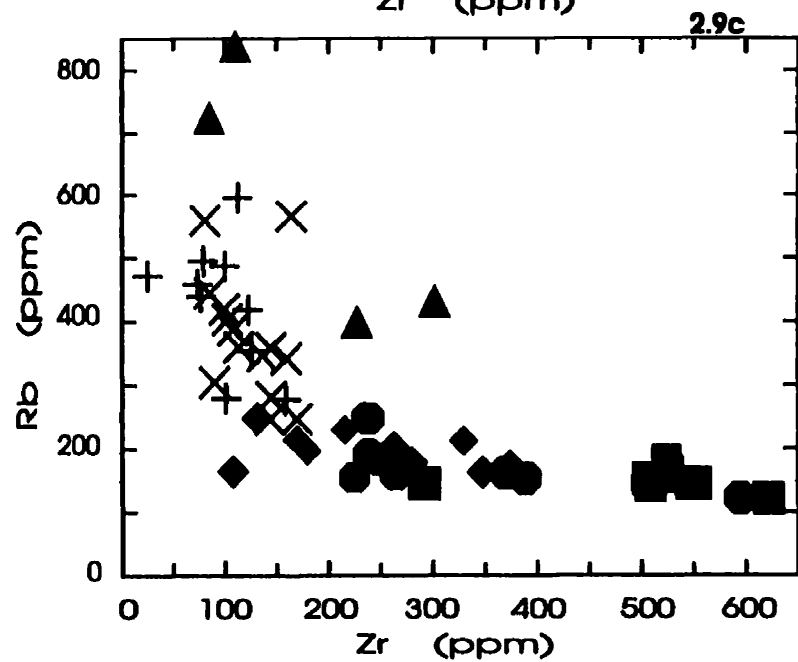
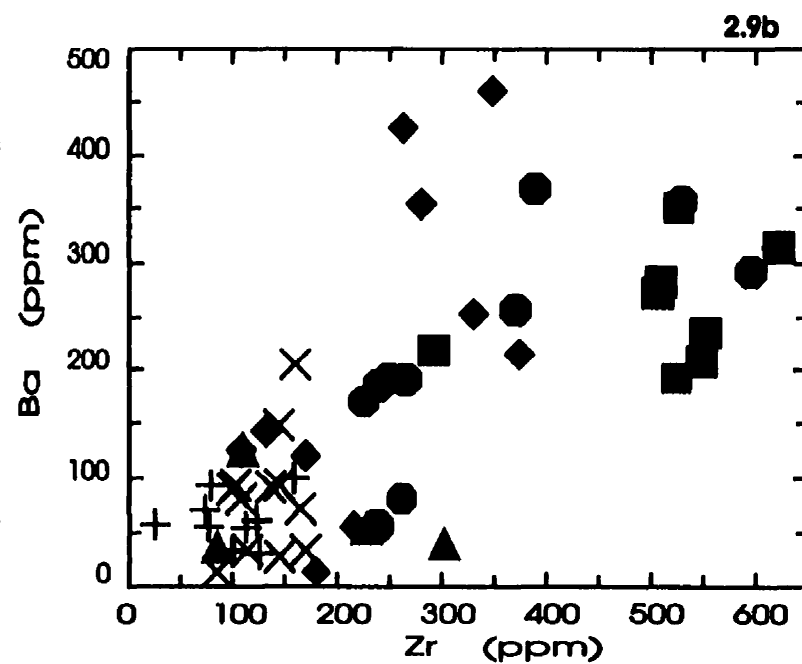
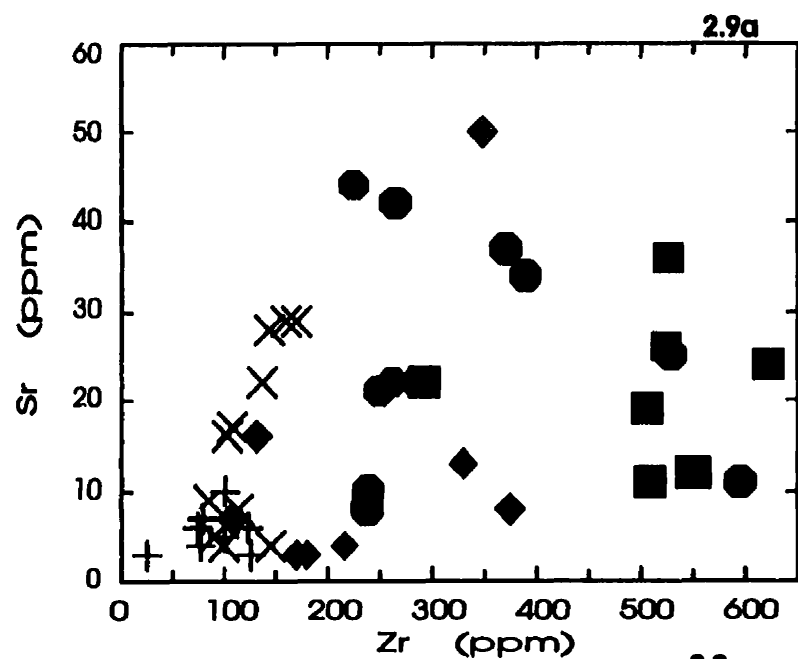


Fig. 2.10: Plots of concentrations of Ni, Pb, Th, and U vs. Zr, all in ppm: **a)** Ni vs. Zr, **b)** Pb vs. Zr, **c)** Th vs. Zr, and **d)** U vs. Zr. Symbols are the same as in Fig. 2.6.

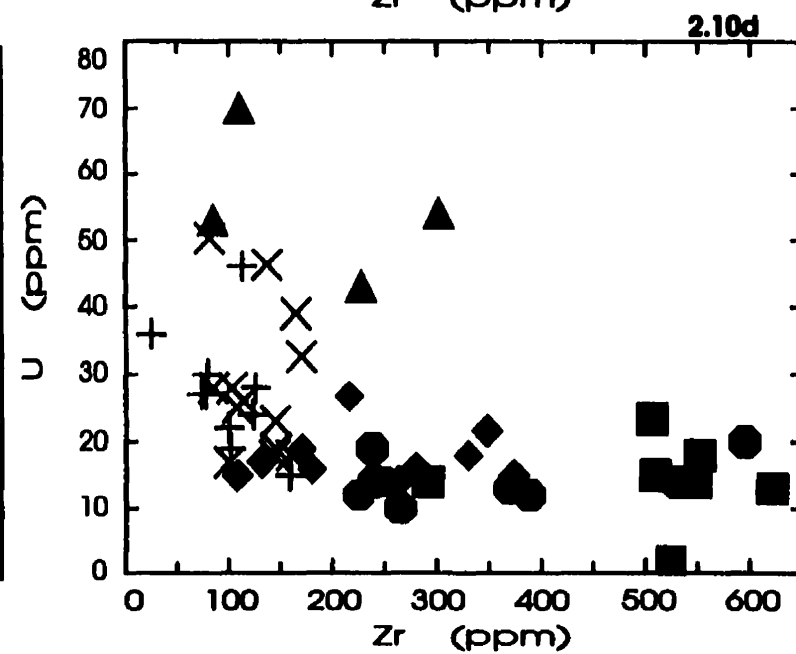
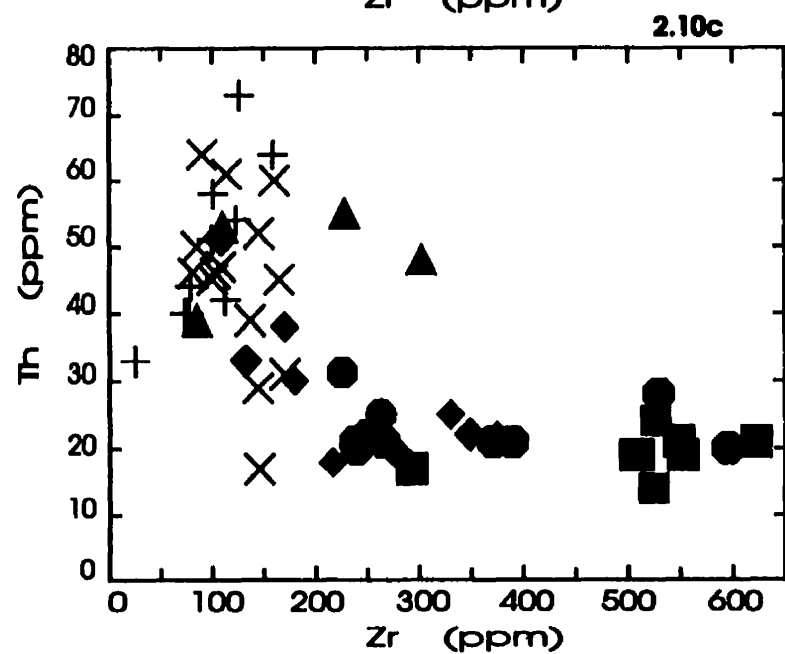
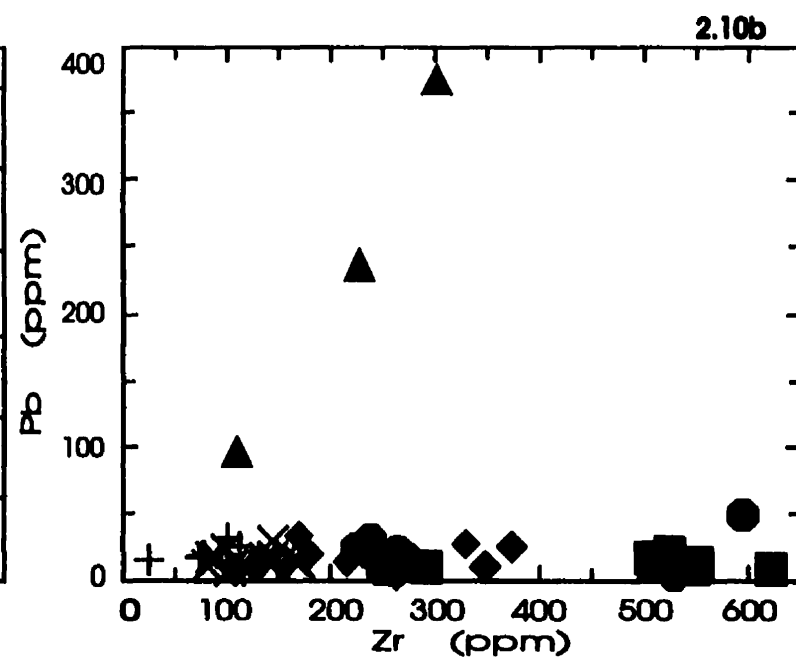
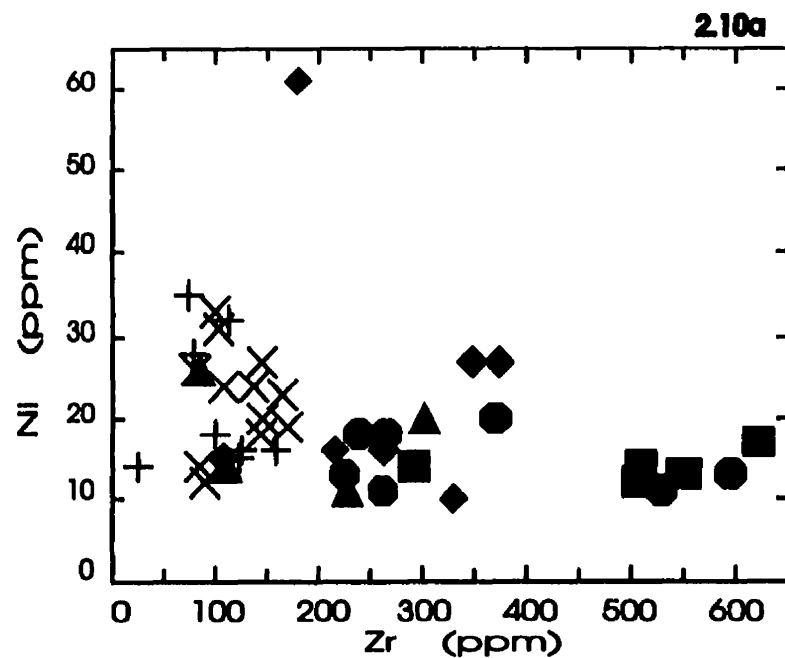
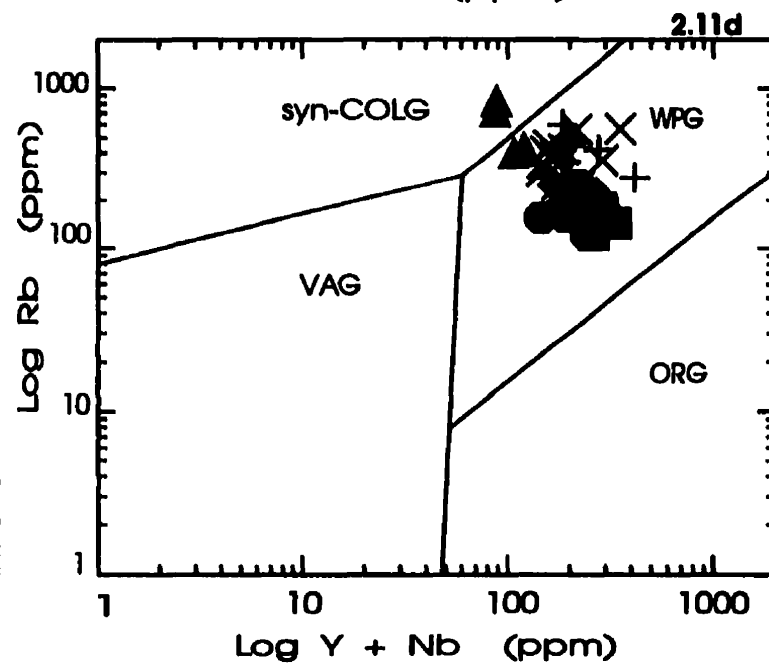
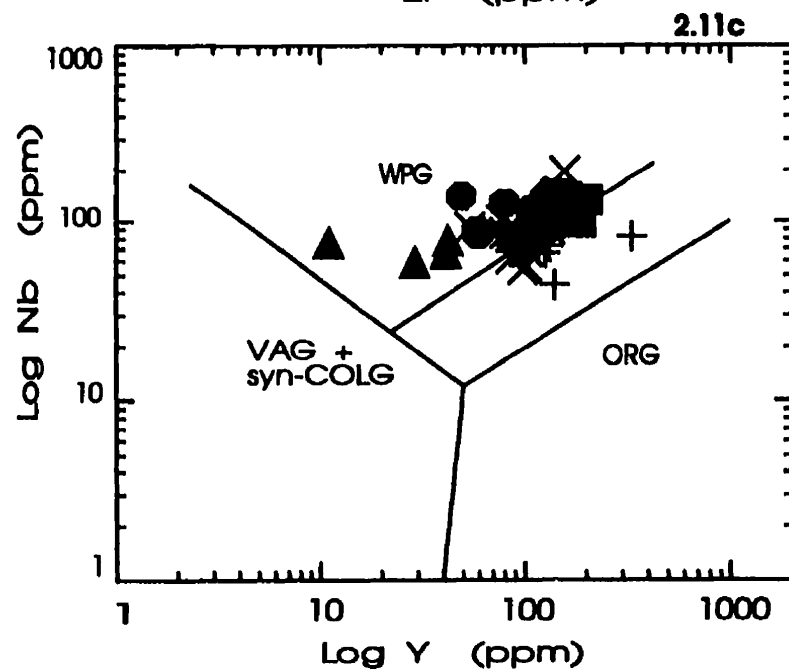
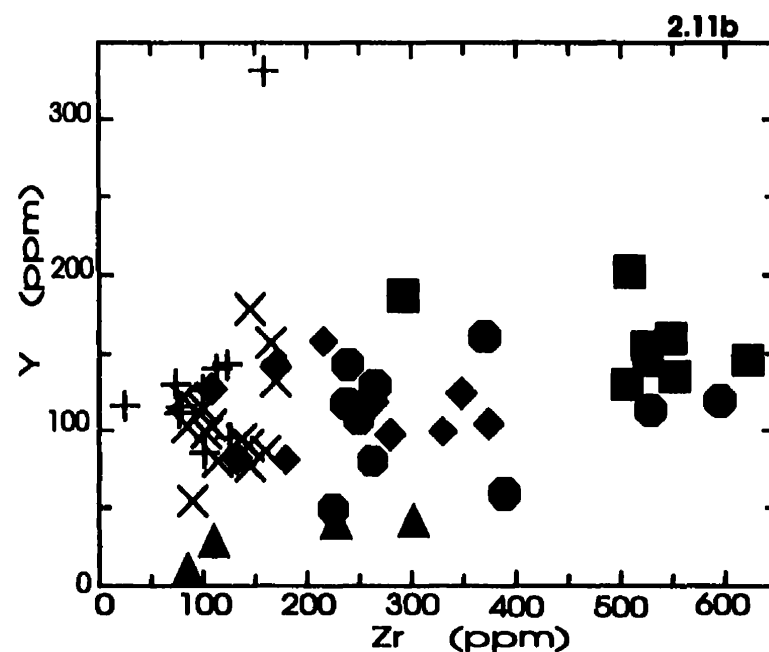
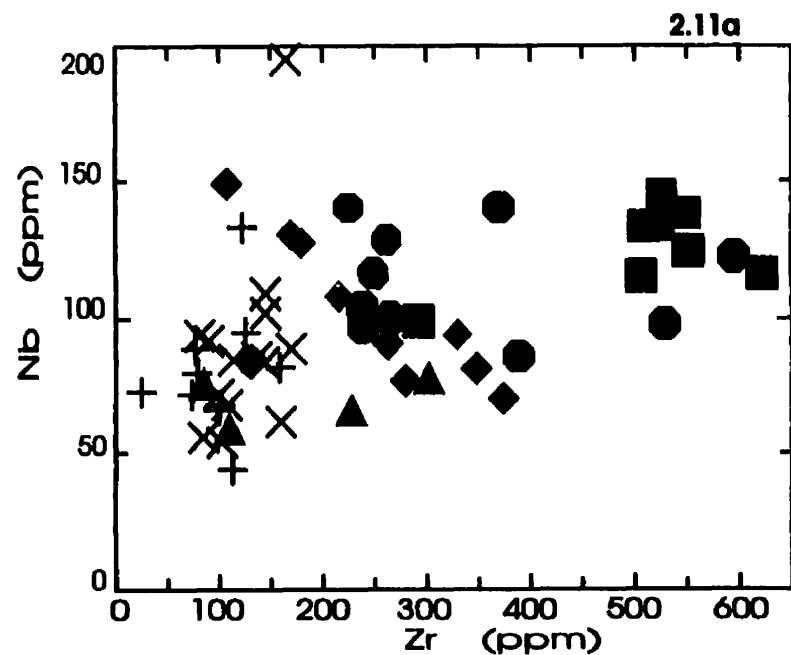


Fig. 2.11: Plots of concentrations of Nb, and Y vs. Zr in ppm, and Log Nb vs. Log Y, and Log Rb vs. Log (Y + Nb): **a)** Nb (ppm) vs. Zr (ppm), and **b)** Y (ppm) vs. Zr (ppm), **c)** Log Nb vs. Log Y, and **d)** Log Rb vs. Log (Y + Nb) for tectonic classification of rocks (WPG = within plate granite, VAG = volcanic arc granite, syn-COLG = syncollisional granite, and ORG = orogenic granite fields: after Pearce et al. 1984). Symbols are the same as in Fig. 2.6.



d). The greisen samples are the most enriched in U (40 to 70 ppm).

The Nb contents of the subvolcanic rocks is relatively constant (Fig. 2.11a), with the biotite granites and their albitized variants having a lower average concentration of Nb and Y (Figs. 2.11a, b). Plots of Log Nb vs Log Y (Fig. 2.11c) and Log Rb vs Log (Y + Nb) in Figure 2.11d establish a within-plate setting for the Kwandonkaya complex.

A decrease in levels of Ba, Ti, Zr, Y, and Nb with increasing concentrations of Si, Rb, U, and Th accompanies the transition from the subvolcanic to plutonic rocks (cf. Sakoma & Martin, 1991, 1992). These variations and the overlaps between less altered rock units displayed in these diagrams (Figs. 2.9 to 2.11) suggest that the rock units are comagmatic, and possibly related by fractionation from a single parental magma, although the patterns are complicated by a postmagmatic overprint in most cases.

Inferences from major and trace element patterns

From the major and trace element data, the following observations are made:

1) Increase in ASI with attendant variation from metaluminous to peraluminous composition in the Kwandonkaya complex is due to an interplay of factors that include fractional crystallization (during magmatic evolution), volatile loss, and hydrothermal alteration (at postmagmatic stages). The postmagmatic effects seems to have had an overriding effect in the development of peraluminosity in the larger bodies of biotite granite.

2) The plot of normative quartz -albite - orthoclase indicates that the subvolcanic rocks evolved by fractionation along a path of water-undersaturated crystallization, whereas plutonic rocks were mostly near-minimum melts that were overprinted by postmagmatic processes.

3) Magmatic evolution by fractional crystallization is indicated by a decrease in TiO_2 , ΣFeO , and Zr, and an increase in Rb. Ba, Sr, Nb, and Y decrease in a pattern consistent with

fractional crystallization, but the scatter in the data suggests an overprint by postmagmatic processes. Also, the concentration of Th and U increases in a pattern suggestive of evolution by fractionation, but these elements seem to have been mobile at the postmagmatic stage, particularly in plutonic rocks and their altered variants.

4) A majority of the subvolcanic rocks plot in the field of normal granites, with respect to Rb-Sr-Ba plot, whereas the plutonic rocks and their altered variants are typically highly evolved.

5) Overall, the rock suite is enriched in LILE (large-ion lithophile elements) such as K, Rb, Ba, Th, and U, and the high-field-strength elements (HFSE), such as Zr, Nb, Y, whereas it is relatively depleted in Sr, Ti, Mg, Ni, Cr, and Ca. These trends, combined with a relatively high Fe/Mg, are A-type signatures.

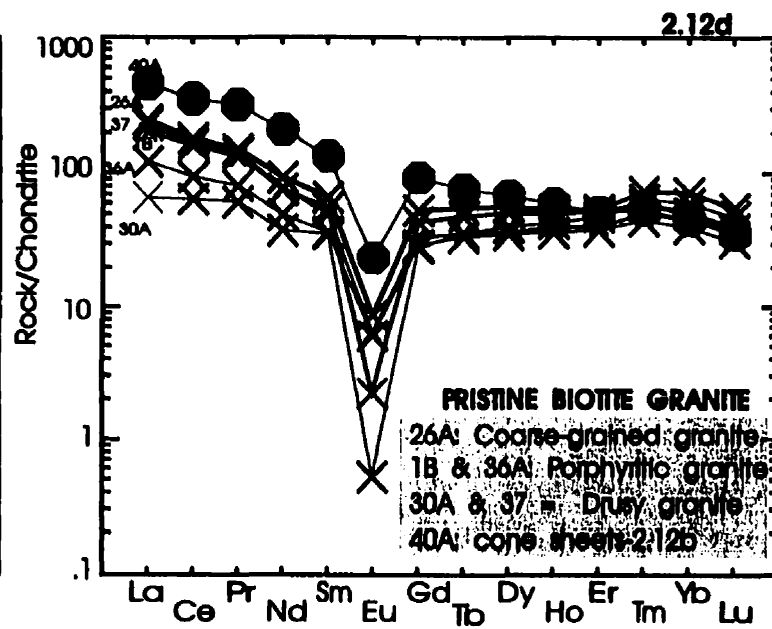
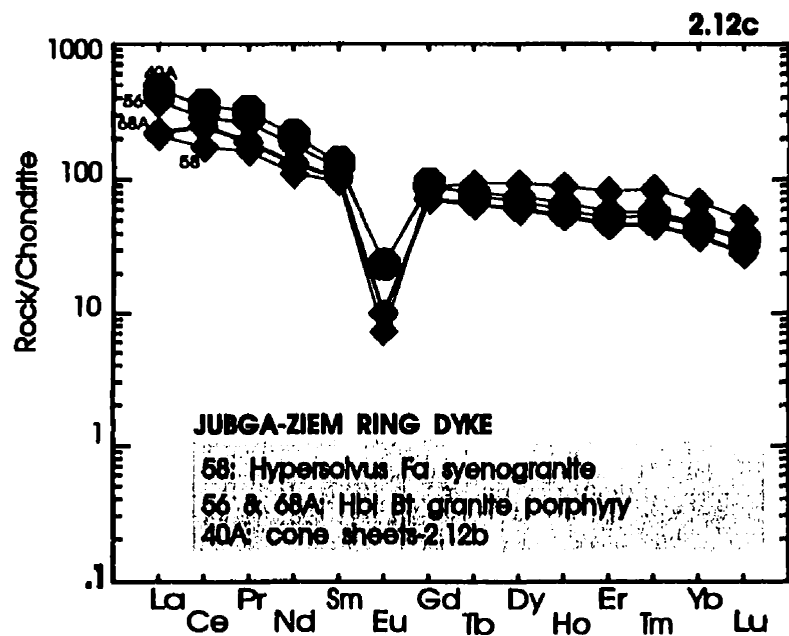
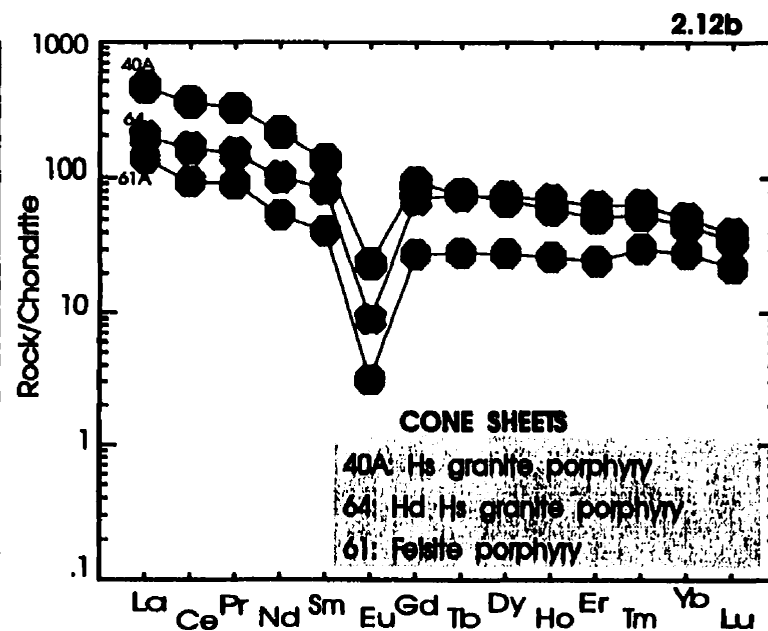
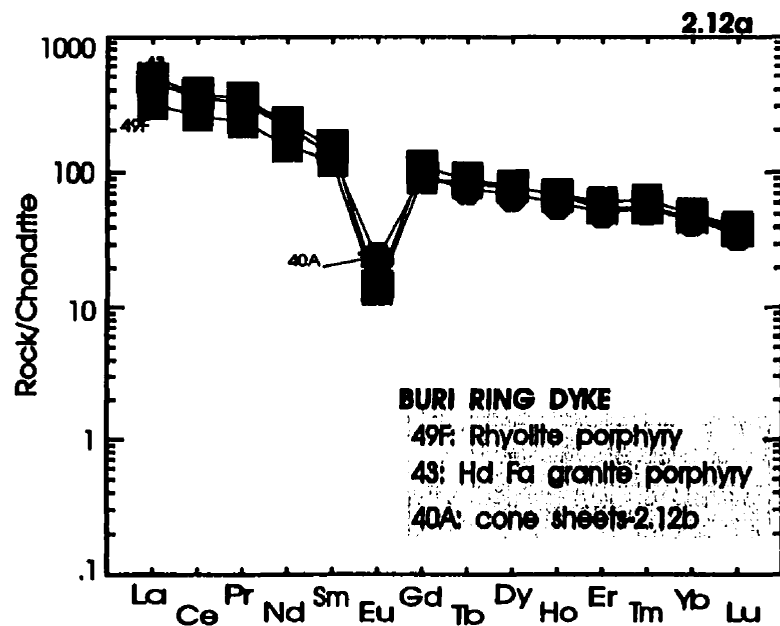
6) Postmagmatic loss of alkalis is indicated by decrease in $\Sigma(\text{Na}_2\text{O} + \text{K}_2\text{O})$, and a scatter in the Al contents. Further indications of mobility of some elements are given by the scatter in the CaO, FeO, and Fe_2O_3 contents of the rock suite, and the enhanced SiO_2 , Fe_2O_3 , and MnO contents of the greisen zones.

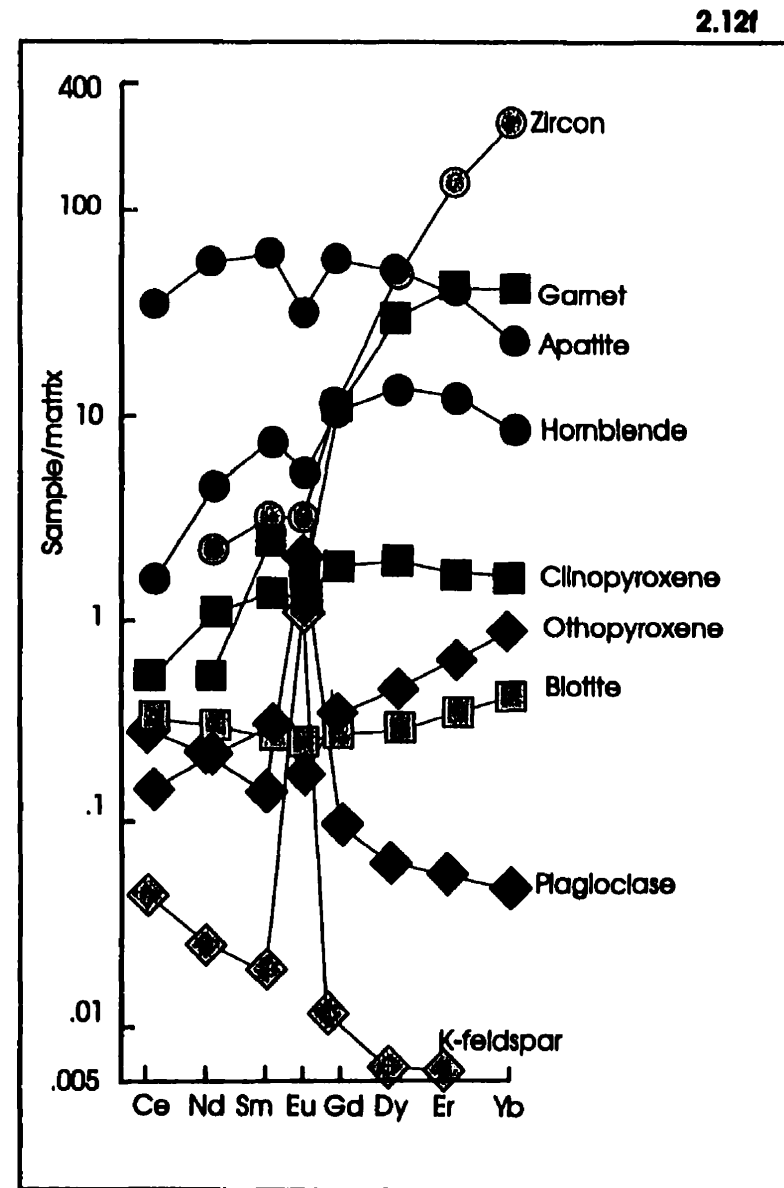
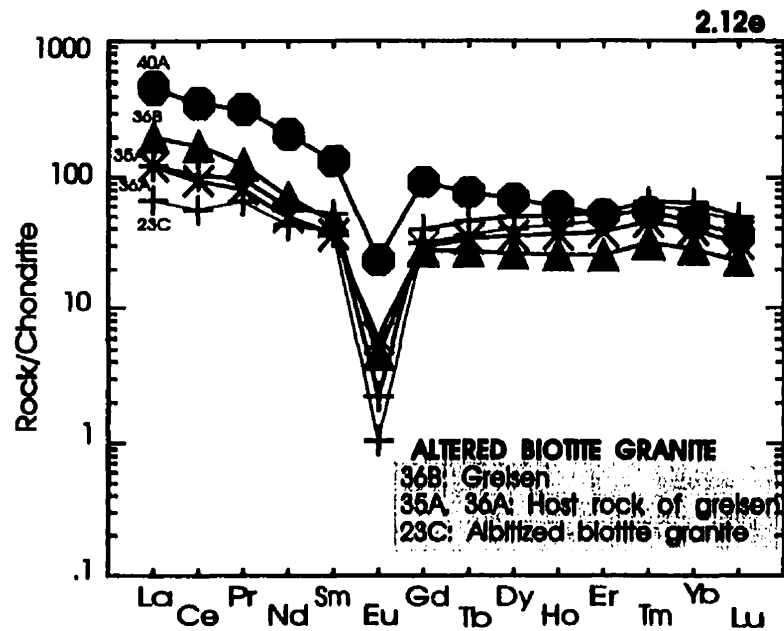
2.10.3 Rare-earth elements

The data on rare-earth-element concentrations are presented in Table 2.3, with chondrite-normalized plots displayed in Figures 2.12a - e. The expected partitioning of REE among selected minerals and the matrix is presented in Figure 2.12f. In the following discussions, La/Lu_N refers to the ratio of chondrite-normalized La and Lu concentrations, whereas Eu/Eu^* is the ratio of observed chondrite-normalized Eu concentration to the amount of Eu expected by interpolation between chondrite-normalized Sm and Gd concentrations.

Both the rhyolite porphyry and hedenbergite fayalite granite porphyry (in the Buri ring

Fig. 2.12: Chondrite-normalized REE plots for rock units from Kwandonkaya complex: **a)** Rock units from the Buri ring dyke: 49A (rhyolite porphyry), 43 (hedenbergite fayalite granite porphyry). Sample 40A (hastingsite granite porphyry) is plotted for comparison. **b)** Cone sheet units: 40A (hastingsite granite porphyry), 64 (hedenbergite hastingsite granite porphyry), and 61A (felsite porphyry). **c)** Subvolcanic rocks in the Jubga-Ziem center: 58 (fayalite syenogranite), 56 and 68A ("hornblende"-biotite granite porphyry). Sample 40A (same as in Figs. 2.13a, b). **d)** Biotite granite: 26A (coarse-grained biotite granite), 30A and 37 (drusy fine-grained biotite granite), 1B and 36A (porphyritic biotite granite), and 40A (as in Figs. 2.13a - c). **e)** Altered biotite granite: 23C (albitized biotite granite), 35A (albitized porphyritic granite host rock of greisen), 36A (unaltered porphyritic granite, same as in 2.13d and is also host rock to greisen), 36B (Greisen). **f)** Phenocryst/matrix plot for various minerals (after Hanson 1978) to indicate their likely role in explaining the REE trends of Kwandonkaya complex. Symbols are the same as in Fig. 2.6.





dyke) share a similar REE pattern with the hastingite granite porphyry (Fig. 2.12a). However, the hedenbergite fayalite granite porphyry ($\text{La/Lu}_N = 13.02$, $\text{Eu/Eu}^* = 0.11$) has a slightly higher LREE enrichment and lower HREE than the coeval rhyolite porphyry ($\text{La/Lu}_N = 7.73$, $\text{Eu/Eu}^* = 0.13$). This presumably is due to an interplay of factors such as presence of residual zircon, and garnet, or different extents of fractionation of zircon, clinopyroxene, and Ca-amphibole (cf. Fig. 2.12f, Hanson 1978). Petrographic observations on subvolcanic rocks favour the fractionation hypothesis.

In cone sheets, the LREE pattern of hastingite granite porphyry is similar to that of hedenbergite fayalite granite porphyry, but that rock shows a slightly lower degree of HREE enrichment compared to both rock units in the Buri ring dyke (Fig. 2.12a). The hastingite granite porphyry ($\text{La/Lu}_N = 12.87$, $\text{Eu/Eu}^* = 0.21$) is, on the other hand, relatively REE-enriched over both felsite porphyry ($\text{La/Lu}_N = 6.16$, $\text{Eu/Eu}^* = 0.09$) and hedenbergite hastingite granite porphyry ($\text{La/Lu}_N = 4.94$, $\text{Eu/Eu}^* = 0.12$), also in cone sheets. The implications of these findings are that the cone sheet units are comagmatic and evolved by fractionation of LREE-rich phases like chevkinite and allanite, noted in petrographic descriptions.

Within the Jubga-Ziem ring dyke, the "hornblende"-biotite granite porphyry ($7.83 \leq \text{La/Lu}_N \leq 10.43$, $0.12 \leq \text{Eu/Eu}^* \leq 0.23$) is LREE-depleted and has a similar HREE pattern to that of the hastingite granite porphyry (Fig. 2.12c). The hypersolvus (fayalite) syenogranite ($\text{La/Lu}_N = 4.16$, $\text{Eu/Eu}^* = 0.08$), also in the Jubga-Ziem ring dyke, is slightly depleted in LREE but is enriched in HREE content compared to granite porphyry in the same ring dyke (Fig. 2.12c), and the other subvolcanic rocks (Fig. 2.12a, b). These subtle differences may be due to a combination of local factors, which include: i) lesser extent of fractionation of Ca-amphibole and clinopyroxene in fayalite syenogranite, ii) enrichment of LREE in marginal F-rich hornblende biotite granite porphyry, and of HREE in F-poor inner zones that crystallized fayalite

syenogranite; according to Keppler (1993), the solubility of the REE in a F-bearing medium increases from La to Lu, and iii) fractional crystallization of allanite (LREE-rich) in the "hornblende"-biotite granite porphyry.

The biotite granite ($1.38 \leq \text{La}/\text{Lu}_N \leq 6.29$, $0.02 \leq \text{Eu}/\text{Eu}^* \leq 0.18$) has a lower LREE content, more strongly negative Eu anomaly, and slightly higher HREE than the subvolcanic rocks (Fig. 2.12d). Biotite granite associated with tin mineralization (30A, 36A) is depleted in LREE compared to the unmineralized granite (1B, 26A, 37), whereas the HREE do not seem affected by the acid metasomatism overprint. On the other hand, albitized biotite granites ($1.04 \leq \text{La}/\text{Lu}_N \leq 2.21$, $0.30 \leq \text{Eu}/\text{Eu}^* \leq 0.48$) display a similar LREE pattern but a slightly higher HREE compared to their less altered equivalents (Fig. 2.12e). The greisen zones ($\text{La}/\text{Lu}_N = 8.62$, $\text{Eu}/\text{Eu}^* = 0.12$), on the other hand, have a higher LREE and lower HREE than the albitized biotite granites. The plutonic rocks generally have a slightly higher HREE enrichment than subvolcanic rocks. One of the likely reasons is the enrichment in F in biotite granite and its altered variants relative to the subvolcanic rocks. This factor is discussed further in the following section.

Inferences from REE patterns

The following inferences may be made on the basis of the REE data (Table 2.3), plots (Figs. 2.12a -e) and above discussion:

- 1) There is an overall similarity in REE patterns of all samples in this suite, despite different extents of evolution; all are characterized by a negative slope in chondrite-normalized plots. This similarity suggests a comagmatic suite. The fractionation of HREE-rich phases (e.g., zircon, garnet, Ca-amphibole, and clinopyroxene) or their retention in the residue probably had an overwhelming effect compared to the effect of fractionation of LREE-rich phases (e.g., chevkinite, allanite), so as to maintain a LREE-rich composition, particularly in the subvolcanic

rocks.

2) The LREE in the more evolved subvolcanic rocks show a slight depletion compared to the less evolved rock units. This is attributed to fractional crystallization of allanite and chevkinite.

3) The least evolved subvolcanic rocks have a prominent negative Eu-anomaly, with similarities in magnitude in the individual subvolcanic intrusions (e.g., Buri ring dyke). This suggests that the subvolcanic rocks suite crystallized from a parental magma that had undergone plagioclase (or ternary feldspar) fractionation, or that the rock suite was formed from batches of magma that arose by a similar extent of partial melting. It could also partly be due to the fact that the $f(\text{O}_2)$ of the source-region was buffered (i.e., remained nearly constant) so as to result in consistent Eu-retention as Eu^{2+} in residual plagioclase (cf. Hanson 1978).

4) The middle REE appear to be slightly depleted relative to both the LREE and HREE (Figs. 2.12a - e). The observed depletion in the middle REE could be due to their retention by clinopyroxene and Ca-amphibole in the residue or their incorporation in fractionating minerals (cf. Fig. 2.12f).

5) The HREE are slightly enriched in modified subvolcanic rocks, plutonic rocks and their altered variants. This is due to the combined effect of fractional crystallization of LREE-rich phases, and build-up and loss of volatiles (F and alkalis), and later hydrothermal alteration.

6) The negative Eu-anomaly is greatest in biotite granite and their altered variants. This anomaly probably is in part due to enhanced Eu mobility with an increase in $f(\text{O}_2)$ associated with degassing and H_2 loss, and during hydrothermal alteration.

2.11 Discussion

The increases in DI, ASI, major and trace elements, and REE documented in this study

are a result of an interplay of processes that include fractional crystallization of a felsic magma, alkali loss, and locally prominent hydrothermal alteration. In this section, the major and trace element variations as well as REE-patterns are discussed in relation to field and petrographic observations so as to distinguish magmatic evolution and postmagmatic alteration.

2.11.1 Magmatic evolution

The increase in ASI from the rhyolite porphyry ($ASI = 0.93$) to hedenbergite fayalite granite porphyry ($0.98 \leq ASI \leq 1.01$) in the Buri ring dyke, despite little or no postmagmatic modifications, is indicative of evolution of the magma in the ring dyke by fractional crystallization. The Jubga-Ziem ring dyke, on the other hand, provides one of the best indications of the development of peraluminous rocks from metaluminous compositions by fractional crystallization combined with alkali loss. Fayalite syenogranite ($ASI = 0.99$) in central parts of the Jubga-Ziem ring dyke crystallized from a metaluminous melt that underwent fractional crystallization to produce marginal, drusy (mildly) peraluminous "hornblende"-biotite granite porphyry ($1.01 \leq ASI \leq 1.05$). Upper sections of the fayalite syenogranite - "hornblende"-biotite granite porphyry are capped by biotite granite porphyry ($ASI = 1.1$) that is more strongly peraluminous, and that developed by crystallization of a volatile-enriched melt. In this example, both fractional crystallization and volatile (alkali) loss seem to have been important, which thus points to the interplay of factors affecting the evolution of the rock suite. In general, the metaluminous subvolcanic rocks of Kwandonkaya complex straddle the metaluminous-peraluminous boundary (Fig. 2.7a), and therefore the mildly peraluminous porphyries could have evolved mainly by fractional crystallization (cf. Zen 1988). Fayalite and hedenbergite crystallization are limited to metaluminous porphyries. Crystallization of hedenbergite and Ca-amphibole depleted the melt in Ca, and drove the metaluminous melts across the metaluminous-

peraluminous boundary, to a mildly peraluminous composition. Fractionation of late magmatic Ca-amphibole that crystallized in metaluminous and mildly peraluminous magma also extended the ASI from metaluminous ($ASI \leq 1$) to mildly peraluminous ($1 \leq ASI \leq 1.2$) compositions (cf. Zen 1986, 1988).

Magmatic evolution by fractional crystallization is also indicated by decreases in TiO_2 , ΣFeO , and Zr with increasing SiO_2 , and increase in Rb (Figs. 2.7c, d, 2.9a, 2.10a). Further indications of evolution by fractionation are provided by the normative plot of quartz - albite - orthoclase (Fig. 2.8d), in which the subvolcanic rocks illustrate an evolution by low pressure fractional crystallization of sanidine_{ss} along water-undersaturated crystallization paths [$0.3 \leq a(H_2O) \leq 0.5$], between 2.5 and 0.5 kbar. Some of the subvolcanic rocks, such as the felsite porphyry in cone sheets of the Jubga-Ziem center, crystallized as near-minimum melts at $P(H_2O)$ of about 0.5 kbar, on the quartz-alkali feldspar cotectic. The least altered hypersolvus biotite granites also appear to have crystallized directly from a relatively water-undersaturated near-minimum rhyolitic melt.

From the major and trace element plots and the REE data, the evolution from the least evolved to the most evolved felsic rock unit involved the fractional crystallization of sanidine_{ss}, fayalite, hedenbergite, Ca-amphibole, Fe-Ti oxides, zircon, chevkinite, and allanite. A quantitative assessment of this model is undertaken in Chapter 8. The normative plot of quartz - albite - orthoclase defines the role of sanidine_{ss}. The fractionation of Fe-rich mafic minerals and Fe-Ti oxides is suggested by the decreasing ΣFeO and TiO_2 with increasing SiO_2 . Zircon fractionation is indicated by the decreasing Zr with the evolution of the rock suite. A Zr content of 120 to 630 ppm in subvolcanic rocks is above the saturation value of 100 ppm in such peraluminous rocks (cf. Watson & Harrison 1984). This accounts for the observation that zircon is invariably early magmatic in the subvolcanic rocks. On the other hand, the biotite granite contains less Zr (90 to

160 ppm), and in this case zircon crystallized later, with (or after) interstitial biotite. It seems that the biotite granites are a product of fractional crystallization of the same parental magma that gave rise to the ring dykes. The role of LREE-rich phases such as chevkinite and allanite is suggested from petrographic observations to be indicative of early to late-magmatic crystallization of these phases, with chevkinite being dominant in metaluminous rock units and allanite in peraluminous rock types. The REE data lend support to inferences made based on petrographic, major and trace element data:

1) The overall predominance of LREE-enrichment is a typical A-type signature, and consistent with the proposal of a comagmatic suite. Zircon, garnet, and orthopyroxene (Fig. 2.12f), as fractionating phases or as residual minerals, could have depleted the initial melt in HREE relative to LREE (Hanson 1978). The early fractionation of HREE-rich phases (or their retention in the residue) is inferred to have been more important than later fractionation of LREE-rich phases (e.g., chevkinite, allanite, alkali feldspar). Both factors contributed to the LREE-rich composition, particularly well developed in the subvolcanic rocks. Assuming retention of such HREE-rich phases in the source rock (cf. Watson & Harrison 1984), it is possible that the HREE were released by later melting (at slightly higher temperature) to result in plutonic rocks having slightly enriched HREE than subvolcanic rocks. The role of garnet here is doubtful, however, because of the mild aluminosity of the resulting HREE-enriched plutonic biotite granites. 2) The more evolved subvolcanic rocks, on the basis of major elements, show a slight depletion in LREE compared to the less evolved rock units. This is explained by the fractional crystallization of allanite and chevkinite.

3) The least evolved subvolcanic rocks have a prominent negative Eu-anomaly, and the magnitude of the negative Eu-anomaly in some individual ring dykes is remarkably constant (e.g., the Buri ring dyke, Fig. 2.12a), whereas the whole subvolcanic suite has a notable range in the

magnitude of negative Eu anomalies (Figs. 2.12a - c). This suggests that the parental magma(s) for the subvolcanic rocks were generated as batches of magma, over a range of melting percentages, hence the range of Eu anomalies. Also, it is likely that these batches of magma underwent some extent of plagioclase fractionation. Another possible explanation is that the $f(\text{O}_2)$ of the source-region was buffered (i.e., remained nearly constant) so as to result in consistent Eu-retention as Eu^{2+} in residual plagioclase (cf. Hanson 1978). The extent of the negative Eu-anomaly is expected to decrease with increasing $f(\text{O})_2$.

4) The middle REE appear to be slightly depleted relative to both the LREE and HREE (Figs. 2.12a - e). The observed depletion in the middle REE could be due to their retention by clinopyroxene and calcic amphibole in the residue or fractionating mineral assemblage (cf. Fig. 2.12f).

Ni is slightly enriched in most units of the plutonic biotite granite (Fig. 2.10a), which suggests contamination either by diffusive exchange between felsic and mafic melts, or interaction of the rocks with a fluid phase in contact with mafic rocks. The role of magma mixing and contamination is explored in Chapter 7.

The petrographic observations that: i) columbite is absent in subvolcanic rocks, whereas it was formed at hydrothermal stages in plutonic rocks, ii) Y-REE-bearing minerals are early magmatic in subvolcanic rocks, whereas they are late-magmatic to hydrothermal in plutonic rocks, iii) zircon is early magmatic in subvolcanic rocks (where thorite is absent), whereas both zircon and thorite are late to postmagmatic in plutonic rocks and their altered variants, all provide indications for the role of postmagmatic processes in enriching and depleting some elements, particularly in the plutonic rocks. In fact, concentrations of Ba, Sr, Nb, and Y decrease with increasing Th and U, and seem to have been perturbed by postmagmatic modifications. Although overall trends suggest evolution by fractional crystallization, signs of postmagmatic overprints are

obvious. The Kwandonkaya suite, in general, displays departures from the typical evolution of pristine, highly evolved A-type granite suite that are known to preserve glass inclusions, such as those investigated by Mungall (1993). This confirms observations that most of the rock units in the NYG have gone at least one step further than the magmatic stages of evolution (Martin & Bowden 1981).

2.11.2 Inferred evolution of biotite granite

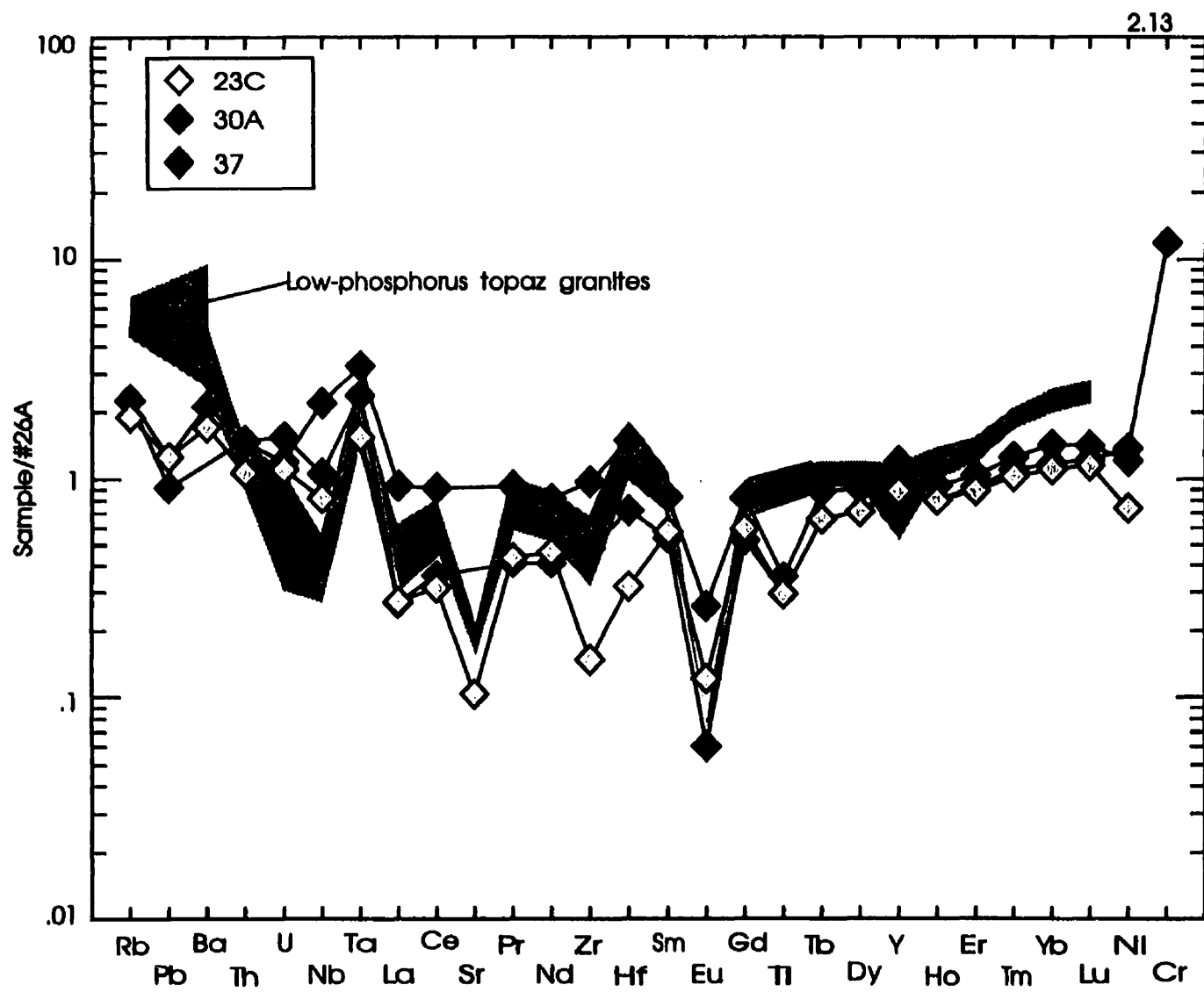
There are several features that suggest a metaluminous parental melt for the biotite granites of the Kwandonkaya complex. First, biotite, the most commonly occurring peraluminous mineral ($1 \leq \text{ASI} \leq 1.5$), is invariably late-magmatic to subsolidus. Furthermore, topaz, which is the most strongly aluminous phase, is clearly subsolidus and of hydrothermal origin. The topaz is invariably later than biotite in Na-metasomatized rocks, greisens, miarolitic pockets, or in drusy cavities within biotite granite. The appearance of biotite and topaz at such late stages of evolution indicates the development of a peraluminous character in residual, interstitial or localized (volatile-rich) melts, and in altered granites as a result of alkali loss and associated hydrothermal alteration below the solidus. The similarities in composition of the mica in hypersolvus orthoclase-bearing biotite granite and in metaluminous hornblende biotite granite (Chapter 4), suggest a comagmatic relationship, and a possible metaluminous parental magma for the biotite granite as well. In fact, the range of ASI (alumina saturation index) for the pristine samples of biotite granite ($1.04 \leq \text{ASI} \leq 1.10$) is similar to that of the subvolcanic rocks ($0.98 \leq \text{ASI} \leq 1.10$) (Table 2. 2). There is a progressive increase in ASI with volatile loss and alteration from topaz-bearing drusy biotite granite ($1.09 \leq \text{ASI} \leq 1.24$), to albitized biotite granite ($1.06 \leq \text{ASI} \leq 1.27$), and the greisens ($1.85 \leq \text{ASI} \leq 6.93$) (Chapter 2). Furthermore, there are similarities in the differentiation index (DI) and agpaitic index (AI) of subvolcanic rocks ($92.6 \leq \text{DI} \leq 95.8$, $0.84 \leq \text{AI} \leq 0.93$) and

pristine biotite granites ($92.8 \leq DI \leq 96.2$, $0.83 \leq AI \leq 0.89$). Thus, the volatile-poor and metaluminous magma that produced the hypersolvus biotite granite may well have produced the peraluminous biotite granite under volatile-rich conditions due to outward and upward alkali loss to the Pan-African country rock (cf. Kalato area of the Tibchi complex: Sakoma 1986).

A comparison of topaz-bearing biotite granites from the Kwandonkaya complex with topaz granites, such as that of Mount Pleasant Ridge in New Brunswick (Taylor 1992), and Spor Mountain in Utah (Christiansen et al. 1986) (Table 2.4 and Fig. 2.13), indicates that the topaz-bearing granites of the Kwandonkaya complex are quite similar to the low-phosphorus subtype of topaz granites of Taylor (1992). The similarities in trace element enrichment patterns in the "type samples" of the low-P topaz granites and topaz-bearing rocks of the Kwandonkaya complex suggest that the former might also have been affected by volatile build-up and loss. In detail, the topaz-bearing granites of the Kwandonkaya complex have slightly lower Rb, Pb, Ba, and HREE compared to the low-P topaz granites of Taylor (1992). Such differences are possibly due to a higher degree of volatile loss and rock-fluid interactions in the biotite granite of the Kwandonkaya complex (comparable to roof zones of Jos-Bukuru complex: Webb 1974, and Kalato area, Tibchi complex: Sakoma 1986).

The drusy topaz-bearing granite on Ziem ridge (sample 37) contains mostly orthoclase, and subsidiary microcline within druses. This sample is hypersolvus and preserves a pristine assemblage of minerals, with a relatively low ASI (1.08), and reflects the effect of early loss of volatiles, which resulted in limited rock-fluid interaction. Another example of drusy topaz-bearing granite (sample 30A) has mainly microcline and albite (a subsolvus assemblage), and is characterized by disseminated cassiterite with topaz and smoky quartz in drusy cavities. This example has relatively high ASI (1.24), and is more enriched in topaz, depicting the effect of late loss of volatiles. The foregoing illustrates well the fact that the observed petrochemical character-

Fig. 2.13: Trace-element abundances in topaz-bearing granites in this study compared to the low-phosphorus type of topaz granite from Mount Pleasant Ridge, New Brunswick (Taylor 1992). The compositions are normalized against that of pristine biotite granite (#26) from the Kwandonkaya complex.



istics of the biotite granites have been modified from what can be considered as magmatic mineral assemblage, even in the most pristine samples. There is also a strong indication that the magma that crystallized to form biotite granite was metaluminous.

2.11.3 Postmagmatic processes

Postmagmatic hydrothermal modifications have been noted in localized areas within subvolcanic rocks, but are more extensive in the plutonic rocks. In the Buri ring dyke, the reddened granite porphyry has a peraluminous character that is attributed to postmagmatic alteration, which led to the destruction of all mafic minerals and the development of pronounced turbidity in the K-feldspar. In the Jubga-Ziem center, alkali loss, accompanied by the replacement of fayalite by late-magmatic annitic biotite + quartz, or postmagmatic magnetite + quartz $\pm$ grunerite, resulted in the development of mild peraluminosity in outer contacts of fayalite syenogranite with the basement. Relict metaluminous mineral assemblages, such as replaced fayalite and hedenbergite in the mildly peraluminous fayalite syenogranite and hornblende biotite granite porphyry, and peraluminous assemblages like biotite and allanite concentrated in the marginal facies, suggest the development of peraluminous compositions by volatile loss.

In a plot of normative quartz - albite - orthoclase, the biotite granites plot away from the $P(H_2O) = 0.5$ kbar minimum, somewhat toward the albite apex (Fig. 2.8d), a feature normally attributed to the buildup of volatiles in the melt in addition to H_2O . Such buildup of volatiles, according to Manning (1981) and Dingwell (1988), depolymerizes the melt, increases the liquidus field of quartz, and shifts the minimum toward the albite apex, as seen in Appendix II.2. The magma that crystallized as the peraluminous biotite granite became saturated; enrichment in halogens facilitated loss of alkalis (as indicated by fluorite- and topaz-bearing drusy facies), and is consistent with the development of the peraluminous composition by volatile loss. In fact, such

degassing of the most evolved magma led to a progressive decrease in $\Sigma(\text{Na}_2\text{O} + \text{K}_2\text{O})$ as the complex crystallized.

The albitized biotite granite immediately around miarolitic cavities mimics the larger-scale processes associated with volatile loss. Prismatic, simply twinned K-feldspar, dogtooth quartz, biotite $\pm$ topaz line the cavities, and exsolved Na- and F-rich fluids pervaded the immediate walls of the cavities and induced metasomatism. The fluids were exsolved due to resurgent boiling, and pervaded the roof and marginal zones of the biotite granite. This resulted in alteration of fine- to medium-grained biotite granite to a saccharoidal albite-enriched granite (cf. Kinnaird 1985), with disseminated columbite and cassiterite. The extent of alteration is found to depend on the fluid/rock ratio and the ASI of the rock before pervasive hydrothermal alteration began. Whereas the above mineral association (which includes Al-Li rich mica and topaz) is produced as a result of alteration of the peraluminous granites, incipient albitization of metaluminous fayalite syenogranite on the other hand is accompanied by coarsening of perthite, replacement of fayalite by chlorite $\pm$ iddingsite $\pm$ annite, with accessory fergusonite, fluocerite, allanite, and ilmenite. Albitized rocks are interpreted as products of alteration by a residual orthomagmatic fluid that evolved during the crystallization of the melt. Such alterations proceeded with a moderate to low fluid/rock ratio, to result in the composition of the altered rocks being buffered by that of the pristine granites. For example, the zirconium content of albitized granites (45 to 130 ppm) suggests that the alteration was buffered by the host rock composition with no significant enrichments.

In the case of greisens, acid metasomatism led to the destruction of feldspars due to replacement of alkalis by the hydrogen ion, and hence the depletion in Na_2O , K_2O , CaO , Sr , Ba , Nb , and Y (Figs. 2.6c, 2.7a, b, 2.9a, b, 2.11a, b). Acid metasomatism also resulted in the build-up of Ti , ΣFe , Mn , Al , Rb , U , Pb and, later, Si (Figs. 2.6b, 2.7c, d, 2.8c, 2.9a, and 2.10b) added

during silification. The greisen zones have an ASI value 2 to 7 times that of their unaltered equivalents. The relative enrichment of Al is expressed by the higher proportion of normative corundum, and secondary minerals that include topaz, Al-rich mica, and allanite. Mass-balance calculations would be required to quantify the effects of postmagmatic alteration. Greisens plot close to the quartz-orthoclase join and are enriched in normative quartz (Fig. 2.8d). The higher concentration of Pb (Fig. 2.10b) in greisens suggests interaction of the acidic fluid with basement rocks. Furthermore, in the greisens there was hydrothermal enrichment of Zr ($85 \leq \text{Zr} \leq 265$) above that contained in the pristine biotite granite. Greisenization is mostly joint-controlled, and therefore proceeded in an open system, accompanying decompression and cooling. According to Kinnaird (1987), from fluid inclusion data, such greisenization occurred between 250 and 380°C.

Postmagmatic modifications can also be inferred from the REE data:

1) The HREE are slightly enriched in albitized and greisenized rocks. This is attributed to the fact that such alterations occurred in F-enriched environments (cf. Kinnaird 1985, 1987). It has been shown by Keppler (1993) that the solubility of REE increases from La to Yb for a given F content (Keppler 1993), hence the enrichment of HREE in highly evolved and relatively F-rich granites.

2) The higher negative Eu-anomaly in altered variants of biotite granite results from Eu mobility due to increases in $f(\text{O}_2)$ associated with degassing and H_2 loss, and during hydrothermal alteration.

2.12 Summary and conclusions

1) The subvolcanic rocks of the Kwandonkaya complex are interpreted as relics of intra-caldera volcanism. Transitional rock-types, between truly volcanic and truly plutonic in character, are preserved as rhyolite porphyry along with granite porphyries (degassed subvolcanic

intrusions).

2) The first (Buri) and second (Jubga-Ziem) centers were controlled by ring faulting and possibly preceded by cauldron subsidence and intra-caldera volcanism, whereas the third center (Panshanu) has no preserved evidence for caldera formation (due to the absence ring faulting and ring-dyke emplacement) (Fig. 2.2). The Kwandonkaya complex evolved in a series of migrating magmatic centers at the volcano-caldera and epizonal levels, although the occurrence of a floored polymagmatic chamber deeper down in the crust (6 - 8 km) (cf. Bonin 1986) is not precluded.

3) Increases in DI, SiO₂, and ASI (hence metaluminous to peraluminous transition) were due to an interplay of factors that included fractional crystallization of sanidine_{ss} at water-undersaturated conditions in subvolcanic rocks [$0.3 \leq a(\text{H}_2\text{O}) \leq 0.5$, at 0.5 to 2.5 kbar]; biotite granites crystallized from near-minimum melts that were volatile-enriched to the point of achieving second boiling and developed a peraluminous character by alkali loss, first at a late magmatic stage; then, the aluminosity (ASI) was enhanced locally by hydrothermal leaching of alkalis.

4) Magmatic evolution by fractional crystallization, from the least evolved to the most evolved felsic rock, is indicated by decreasing ΣFeO , TiO₂, and Zr with increasing SiO₂ and Rb. The fractionation of mafic minerals (fayalite and hedenbergite), Fe-Ti oxides (ilmenite and magnetite), and zircon is inferred from these trends.

5) The REE data corroborate inferences, with respect to magmatic and postmagmatic evolution, made on the basis of major and trace element data: i) the overall similarity in the REE patterns suggests a comagmatic suite; ii) progressive slight depletions in LREE indicates fractionation of LREE-rich phases such as chevkinite and allanite at a magmatic stage, and later enrichments in HREE are related to F and volatile build-up in altered variants, iii) progressive increases in magnitude of the negative Eu anomaly in pristine subvolcanic rocks and biotite

granite indicates fractional crystallization of a ternary feldspar, whereas in altered biotite granite, it indicates selective mobility of Eu associated with loss of H_2 and accompanying oxidation during degassing.

6) Field evidence for mixing between felsic and mafic rocks proves the availability of mafic magma throughout the evolution of the complex. Local contamination of felsic rock units is indicated by enhanced Ni concentrations. The influence of magma mixing, documented in Chapter 7, seems to be highly localized, presumably because of the epizonal nature of the complex.

7) The petrographic and petrochemical features illustrate well the fact that the characteristics of the biotite granites have been modified from what can be considered as magmatic mineral assemblage, even in the most pristine samples. There is also a strong indication that the magma that crystallized to form biotite granite was metaluminous. It has also been indicated that although there are similarities in trace element enrichment patterns of topaz-bearing granites in the Kwandonkaya complex and the low-phosphorus type topaz granites of Taylor (1992), in detail, the topaz-bearing granites in Kwandonkaya have slightly lower Rb, Pb, Ba, and HREE compared to the low-P topaz granites. Such differences are possibly due to a higher degree of volatile loss and rock-fluid interactions in the biotite granite of the Kwandonkaya complex.

8) Postmagmatic alteration is indicated by the apparent mobility of Ca, Sr, Ba, Nb, Y, Th and U in the whole suite, and alkali loss is suggested by the slight decrease in $\Sigma(Na_2O + K_2O)$ with increasing bulk SiO_2 . In the greisens, increased concentrations of Ti, ΣFe , Mn, and Pb points to the mobilization of these elements in the acidic fluids.

Table 2.2: Representative major and trace element data, and CIPW normative compositions.

Legend for Table 2: LOI = loss on ignition. Mineral abbreviations used in listing normative composition: **Qtz** = quartz, **Crn** = corundum, **Zrn** = zircon, **Or** = orthoclase, **Ab** = albite, **An** = anorthite, **Di** = diopside, **Hy** = hypersthene, **Mag** = magnetite, **Hem** = hematite, **Ilm** = ilmenite, **Ap** = apatite, **Rut** = rutile. **D.I.** = differentiation index, **ASI** = alumina saturation index, and **A.I.** = albitic index (see text). The concentrations of major elements, expressed as oxides, are expressed in wt.%, whereas that of the trace elements is expressed in ppm.

Detection limits are: Major elements: 0.01% (oxide); Cr: 15 ppm; V, Ni, and Ba: 10 ppm; Nb, Zr, Sr: 30 ppm; Y, Rb, Pb, Th, U: 5 ppm. Sulphur has a detection limit of 0.1%. Zero values in data imply element concentration is below detection limits. Dashes indicate elements whose concentration was not determined.

	-Rhyolite- porphyry		-----Hedenbergite fayalite granite-----						----Cone sheets----		
Sample	49F	87	41	42	43	44B	65	66	40D	61A	61B
SiO2	73.70	74.10	74.21	73.21	72.69	74.16	71.58	75.58	77.29	77.05	76.82
TiO2	0.36	0.35	0.25	0.25	0.25	0.24	0.31	0.16	0.11	0.09	0.11
Al2O3	12.45	12.28	12.83	12.88	12.98	12.90	13.29	12.38	12.23	12.05	11.67
Fe2O3	3.56	3.32	1.16	1.13	1.33	1.33	3.16	1.97	1.55	1.68	1.08
FeO	-	-	1.76	1.81	1.62	1.41	0.42	0.31	0.31	0.24	0.59
MnO	0.06	0.04	0.08	0.06	0.07	0.08	0.06	0.01	0.01	0.01	0.00
MgO	0.21	0.12	0.00	0.00	0.00	0.00	0.07	0.00	0.00	0.00	0.00
CaO	0.85	0.84	0.62	0.66	0.72	0.53	0.49	0.30	0.24	0.21	0.48
Na2O	3.91	3.90	3.86	3.84	3.86	3.84	3.92	3.36	3.55	3.02	3.01
K2O	4.85	4.83	5.21	5.08	5.19	5.17	5.34	5.08	4.46	4.98	5.15
P2O5	0.16	0.13	0.01	0.02	0.02	0.01	0.03	0.00	0.00	0.00	0.00
LOI	0.46	0.24	0.32	0.30	0.38	0.34	0.74	0.80	0.41	0.55	0.80

Trace elements

Cr	25	25	0	0	0	0	0	0	0	0	0
Ni	0	0	13	13	12	14	17	14	11	13	0
V	13	0	17	17	0	10	12	13	24	11	0
Pb	21	22	9	13	17	11	9	10	8	24	8
S	0	0	0	0	0	0	8720	5860	0	0	800
Rb	166	182	145	145	152	142	121	145	167	154	180
Ba	351	192	208	236	273	283	315	218	81	170	192
Sr	36	26	12	12	19	11	24	22	22	44	21
Ta	9.09	-	-	-	6.01	-	-	-	-	8.13	-
Nb	145	133	138	124	115	133	116	99	128	140	116
Hf	11	-	-	-	11.23	-	-	-	-	6.27	-
Zr	526	524	548	552	506	509	622	292	262	225	249
Y	145	155	160	135	130	202	146	187	80	49	107
Th	24	14	21	19	19	19	21	17	25	31.18	22
U	0	2	14	18	23.31	15	13	14	0	12	14
Rb/Sr	4.61	7.00	12.08	12.08	8.00	12.91	5.04	6.59	7.59	3.50	8.57
Ba/Sr	9.75	7.38	17.33	19.67	14.37	25.73	13.13	9.91	3.68	3.86	9.14
Ba/Rb	2.11	1.05	1.43	1.63	1.82	1.99	2.60	1.50	0.49	1.10	1.07
Th/U	-	7.00	1.54	1.06	0.82	1.27	1.62	1.21	-	2.68	1.57

Normative composition

Qtz	35.87	31.58	29.69	29.42	28.61	30.39	27.57	36.20	39.10	40.19	38.93
Crn	0.44	0.00	0.00	0.00	0.00	0.04	0.24	0.81	1.10	1.31	0.26
Zrn	0.02	0.11	0.11	0.11	0.10	0.10	0.13	0.06	0.05	0.05	0.05
Or	27.97	28.62	30.85	30.40	31.13	30.71	32.03	30.34	26.49	26.69	30.84
Ab	31.47	33.00	32.66	32.84	33.08	32.60	33.61	28.67	30.11	25.72	25.75
An	2.96	1.72	2.27	2.92	2.78	2.58	2.27	1.51	1.20	1.07	2.42
Di	0.00	0.46	0.66	0.24	0.64	0.00	0.00	0.00	0.00	0.00	0.00
Hy	0.19	0.09	1.66	1.99	1.28	1.25	0.18	0.00	0.00	0.00	0.01
Mag	0.93	0.00	1.68	1.66	1.95	1.93	0.67	0.58	0.72	0.55	1.58
Hem	0.00	3.32	0.00	0.00	0.00	0.00	2.74	1.59	0.00	0.00	0.00
Ilm	0.19	0.08	0.47	0.48	0.48	0.46	0.60	0.31	0.21	1.31	0.00
Ap	0.02	0.31	0.02	0.05	0.05	0.02	0.07	0.00	0.00	0.00	0.00
Rut	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00

D.I.	95.31	93.20	93.20	92.66	92.82	93.70	93.21	95.21	95.70	92.60	95.52
ASI	0.94	0.93	0.98	0.99	0.98	1.00	1.01	1.07	1.10	1.12	1.02
A.I.	0.94	0.95	0.93	0.92	0.92	0.92	0.92	0.89	0.87	0.86	0.90

	-----Cone sheets (cont'd)-----							---Fayalite syenogranite---			
	40A	40C	62	63	64	69	71	47A	47B	57	58
SiO2	73.77	73.81	77.44	77.48	77.93	76.79	73.02	74.42	75.47	75.45	77.98
TiO2	0.23	0.20	0.12	0.12	0.12	0.14	0.28	0.12	0.10	0.19	0.12
Al2O3	12.83	13.15	11.44	11.41	11.53	11.84	12.78	13.00	12.80	12.52	11.41
Fe2O3	1.50	2.45	1.15	1.12	0.43	1.02	2.00	1.12	1.02	1.38	0.25
FeO	1.41	0.45	0.85	0.68	0.79	0.64	1.13	0.84	0.57	0.83	1.28
MnO	0.05	0.02	0.07	0.04	0.03	0.01	0.06	0.02	0.01	0.04	0.05
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.81	0.32	0.51	0.48	0.84	0.38	0.49	0.65	0.40	0.43	0.52
Na2O	3.42	3.47	3.06	3.27	3.08	2.98	3.81	3.51	3.37	3.44	3.39
K2O	5.10	5.27	4.71	4.73	4.75	4.96	5.11	5.57	5.61	5.11	4.63
P2O5	0.02	0.02	0.00	0.01	0.00	0.00	0.02	0.00	0.00	0.01	0.00
LOI	0.31	0.62	0.39	0.37	0.40	0.33	0.52	0.36	0.40	0.64	0.58

Trace elements

Cr	0	0	0	0	0	10	0	0	15	0	0
Ni	11	0	18	0	18	20	13	0	61	10	16
V	14	0	13	0	14	16	15	11	16	0	19
Pb	5	0	30	21	21	0	49	33	19	27	14
S	0	0	0	0	0	400	0	3680	0	0	5960
Rb	155	153	190	249	160	161	121	214	197	213	230
Ba	356	368	185	56	191	257	292	120	14	253	56
Sr	25	34	10	8	42	37	11	3	3	13	4
Ta	5.64	-	-	-	8	-	-	-	-	-	8.23
Nb	98	86	97	104	100	140	122	130	127	94	108
Hf	7.06	-	-	-	8.74	-	-	-	-	-	9.20
Zr	529	389	239	238	265	370	595	170	180	330	216
Y	113	59	143	117	129	161	119	143	81	99	158
Th	28	21	20	21	21	21	20	38	30	25	18
U	14	12	14	19	10	13	20	19	16	18	26.69
Rb/Sr	6.20	4.50	19.00	31.13	3.81	4.35	11.00	71.33	65.67	16.38	57.50
Ba/Sr	14.24	10.82	18.50	7.00	4.55	6.95	26.55	40.00	4.67	19.46	14.00
Ba/Rb	2.30	2.41	0.97	0.22	1.19	1.60	2.41	0.56	0.07	1.19	0.24
Th/U	2.00	1.75	1.43	1.11	2.10	1.62	1.00	2.00	1.88	1.39	0.67

Normative composition

Qtz	32.40	33.04	40.48	39.45	39.87	40.03	30.60	31.26	33.67	35.06	38.65
Crn	0.25	1.20	0.37	0.05	0.00	0.88	0.13	0.00	0.45	0.56	0.00
Zrn	0.10	0.08	0.05	0.05	0.05	0.08	0.12	0.03	0.04	0.07	0.04
Or	30.46	31.47	28.09	28.24	28.28	29.75	30.65	33.25	33.45	30.47	27.55
Ab	29.19	29.61	26.06	27.85	26.19	25.53	32.66	29.92	28.70	29.48	28.79
An	3.93	1.48	2.55	2.33	3.60	1.92	2.33	3.25	2.00	2.09	2.22
Di	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.33
Hy	1.07	0.00	0.55	0.20	0.68	0.13	0.08	0.46	0.07	0.15	1.87
Mag	2.19	0.94	1.68	1.63	0.63	1.50	2.94	1.64	1.49	2.01	0.36
Hem	0.00	1.82	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ilm	0.44	0.38	0.23	0.23	0.23	0.27	0.54	0.23	0.19	0.36	0.23
Ap	0.05	0.05	0.00	0.02	0.00	0.00	0.05	0.00	0.00	0.02	0.00
Rut	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00

D.I.	92.05	94.12	94.63	95.54	94.34	95.31	93.01	94.43	95.82	95.01	94.99
ASI	1.02	1.10	1.04	1.00	0.98	1.08	1.01	1.00	1.04	1.05	0.99
A.I.	0.87	0.87	0.89	0.92	0.88	0.88	0.92	0.91	0.91	0.89	0.93

	-"Hornblende" biotite & Biotite granite- porphyry						---Coarse-grained--- biotite granite			Porphyritic Bt granite(JZ)	
	53	55	56	68A	68B	38	26A	26B	45A	18	25A
SiO2	76.49	74.60	73.62	76.01	74.85	76.72	76.59	74.81	76.67	76.83	76.24
TiO2	0.16	0.13	0.19	0.22	0.17	0.11	0.17	0.15	0.07	0.06	0.07
Al2O3	12.67	12.74	13.08	11.93	12.90	12.98	12.77	13.19	12.08	13.08	12.50
Fe2O3	0.67	0.88	0.91	1.39	1.01	0.53	0.06	0.00	0.67	0.67	0.65
FeO	0.39	1.44	1.30	0.95	1.16	0.71	1.55	1.68	0.57	0.44	0.49
MnO	0.00	0.03	0.03	0.03	0.03	0.01	0.02	0.02	0.00	0.01	0.01
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.04	0.00	0.00	0.00	0.00
CaO	0.78	0.59	0.77	0.32	0.54	0.40	0.60	0.64	0.29	0.47	0.55
Na2O	3.19	3.45	3.72	3.13	3.52	3.45	3.43	3.55	3.51	4.05	3.55
K2O	4.98	5.26	5.02	5.19	5.13	5.00	4.98	4.97	4.56	4.41	4.51
P2O5	0.02	0.00	0.02	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.00
LOI	0.67	0.31	0.35	0.35	0.34	0.61	0.41	0.37	0.37	0.44	0.73

Trace elements

Cr	0	0	0	0	17	0	0	0	0	0	12
Ni	16	15	27	27	0	0	19	18	27	33	14
V	12	15	16	14	0	0	11	14	11	17	0
Pb	5	8	10	25	15	14	13	12	29	23	17
S	800	800	800	0	0	0	0	0	0	1200	0
Rb	202	164	162	172	178	249	248	247	282	420	445
Ba	425	125	460	215	354	142	34	148	28	90	14
Sr	70	7	50	8	22	16	29	28	0	4	9
Ta	-	-	5.47	3.41	-	-	7.36	-	-	-	-
Nb	91	149	81.50	70.20	77	85	89	84	102	72	56
Hf	-	-	9.13	10.71	-	-	4.75	-	-	-	-
Zr	263	108	348	374	280	132	170	144	145	100	85
Y	118	127	124	104	97	82	132	90	77	114	102
Th	25	51	22	21.55	19	33	31	29	17	45	50
U	14	15	21.57	15	16	17	32.68	18	23	17	28
Rb/Sr	2.89	23.43	3.24	21.50	8.09	15.56	8.55	8.82	-	105.00	49.44
Ba/Sr	6.07	17.86	9.20	26.89	16.09	8.88	1.17	5.29	-	22.50	1.56
Ba/Rb	2.10	0.76	2.84	1.25	1.99	0.57	0.14	0.60	0.10	0.21	0.03
Th/U	1.79	3.40	1.02	1.47	1.19	1.94	0.95	1.61	0.74	2.65	1.79

Normative composition

Qtz	37.43	32.58	30.90	37.40	33.29	36.31	35.02	32.74	38.57	35.20	37.46
Crn	0.65	0.29	0.16	0.60	0.59	1.15	0.65	0.82	0.86	0.77	0.76
Zrn	0.05	0.02	0.06	0.06	0.06	0.03	0.03	0.03	0.03	0.02	0.02
Or	29.70	31.43	30.14	31.00	30.60	29.67	29.53	29.76	27.49	26.22	27.21
Ab	27.17	29.45	31.90	26.70	29.99	29.22	29.02	30.33	30.17	34.26	30.47
An	3.79	2.96	3.76	1.54	2.64	1.99	2.93	3.15	1.40	2.33	0.00
Di	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.27
Hy	0.00	1.78	1.40	0.30	1.08	0.70	2.66	2.91	0.39	0.18	0.27
Mag	0.80	1.29	1.34	2.03	1.47	0.77	0.09	0.00	0.99	0.97	0.96
Hem	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ilm	0.31	0.25	0.37	0.42	0.33	0.21	0.32	0.29	0.14	0.11	0.13
Ap	0.05	0.00	0.05	0.02	0.02	0.00	0.02	0.02	0.02	0.00	0.00
Rut	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
D.I.	94.30	93.46	92.94	95.10	93.88	95.20	93.57	92.83	96.23	95.68	95.14
ASI	1.05	1.03	1.01	1.05	1.05	1.10	1.05	1.07	1.08	1.07	1.07
A.I.	0.84	0.89	0.88	0.90	0.88	0.86	0.86	0.85	0.89	0.87	0.86

	---Fine-grained biotite--- biotite-granite				---Porphyritic biotite granite--- (Panshanu)				---Albitized granite		
	28B	30A	31	37	1A	1B	2A	25D	36A	15	16
SiO ₂	76.36	75.80	76.56	76.60	77.20	75.79	75.89	76.58	75.17	76.72	76.79
TiO ₂	0.08	0.05	0.08	0.06	0.10	0.13	0.10	0.08	0.09	0.04	0.06
Al ₂ O ₃	12.26	13.16	12.86	12.99	12.80	13.22	12.72	13.73	13.48	13.05	12.96
Fe ₂ O ₃	1.02	0.49	1.04	0.21	0.64	1.40	0.99	0.78	0.00	0.50	0.35
FeO	0.17	0.96	0.29	1.04	0.46	0.00	0.38	0.53	1.36	0.49	0.60
MnO	0.01	0.02	0.01	0.02	0.02	0.01	0.02	0.02	0.01	0.02	0.02
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.36	0.40	0.55	0.44	0.61	0.68	0.59	0.47	0.54	0.30	0.42
Na ₂ O	3.48	3.43	3.73	3.87	3.73	3.78	3.73	3.27	3.78	4.09	4.02
K ₂ O	4.44	3.94	4.32	4.47	4.72	4.43	4.67	4.24	5.02	4.45	4.49
P ₂ O ₅	0.00	0.00	0.00	0.01	0.01	0.02	0.01	0.01	0.01	0.00	0.01
LOI	0.37	0.54	0.76	0.41	0.45	0.56	0.52	0.60	0.40	0.36	0.30

Trace elements

Cr	0	12	0	0	0	50	29	0	22	25	0
Ni	0	26	20	23	31	170	24	12	24	32	28
V	24	0	0	0	0	10	0	0	0	0	0
Pb	11	11	17	15	8	19	16	17	16	25	18
S	800	800	0	0	0	400	400	0	0	400	0
Rb	360	559	359	566	404	342	387	305	347	596	495
Ba	34	0	94	72	94	205	81	0	92	55	94
Sr	8	0	4	0	16	29	17	5	22	0	7
Ta	-	17.54	-	23.91	-	0.18	-	-	7.71	-	-
Nb	85	94	109	195.40	54	61.80	68	92	86	44	80
Hf	-	3.41	-	7.06	-	3.91	-	-	3.74	-	-
Zr	114	81	145	165	103	160	108	90	137	113	79
Y	80	121	179	157	98	86	103	54	94	140	115
Th	61	46	52	45	46	60.02	47	64	39	42	44
U	26	50.27	19	39	28	18	25	0	46.29	46	30
Rb/Sr	45	-	89.75	-	25.25	11.79	22.76	61.00	15.77	-	70.71
Ba/Sr	4.25	-	23.50	-	5.88	7.07	4.76	0.00	4.18	-	13.43
Ba/Rb	0.09	0.00	0.26	0.13	0.23	0.60	0.21	0.00	0.27	0.36	0.19
Th/U	2.35	0.92	2.74	1.15	1.64	3.33	1.88	-	0.84	4.27	1.47

Normative composition

Qtz	30.98	39.88	37.27	35.27	35.87	35.55	35.31	40.29	31.84	35.03	34.99
Crn	1.07	2.54	1.03	0.98	0.44	1.00	0.46	2.92	0.85	0.93	0.72
Zrn	0.02	0.02	0.03	0.03	0.02	0.02	0.02	0.02	0.03	0.02	0.02
Or	26.87	23.92	25.81	26.71	27.97	26.46	28.00	25.25	29.96	26.62	26.80
Ab	29.99	29.54	31.74	32.84	31.47	32.16	31.85	27.75	32.16	34.72	34.11
An	1.82	2.02	2.75	2.12	2.96	3.27	2.89	2.27	2.64	1.49	2.03
Di	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Hy	0.00	1.34	0.00	1.68	0.19	0.00	0.00	0.24	2.39	0.47	0.76
Mag	0.36	0.72	0.75	0.31	0.93	0.00	1.02	1.13	0.00	0.73	0.51
Hem	0.79	0.00	0.53	0.00	0.00	1.41	0.30	0.00	0.00	0.00	0.11
Ilm	0.15	0.10	0.15	0.11	0.19	0.07	0.19	0.15	0.17	0.08	0.02
Ap	0.00	0.00	0.00	0.02	0.02	0.05	0.02	0.02	0.02	0.00	0.00
Rut	0.00	0.00	0.00	0.00	0.00	0.10	0.00	0.00	0.00	0.00	0.00
D.I.	87.84	93.44	94.84	94.82	95.31	94.17	95.16	93.29	93.96	96.37	95.90
ASI	1.10	1.24	1.09	1.08	1.04	1.08	1.04	1.27	1.07	1.08	1.06
A.I.	0.86	0.75	0.84	0.86	0.88	0.83	0.88	0.73	0.86	0.88	0.89

	-----Albitized biotite granite-----								-----Greisens-----		
	17	22C	23C	24	35A	74	75a	75b	35C	36B	36B1
SiO2	77.22	77.42	77.06	77.28	76.41	76.09	78.17	76.82	74.27	78.82	77.32
TiO2	0.06	0.06	0.05	0.06	0.08	0.06	0.10	0.03	0.08	0.21	0.18
Al2O3	12.85	12.17	12.82	12.29	12.63	13.00	12.74	12.77	12.51	11.56	12.74
Fe2O3	1.23	0.60	0.22	0.48	0.08	0.40	0.83	0.98	2.59	4.55	4.21
FeO	0.00	0.63	0.88	0.37	1.09	0.68	0.44	0.00	5.73	1.23	1.16
MnO	0.02	0.03	0.01	0.00	0.02	0.01	0.02	0.00	0.10	0.16	0.14
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.51	0.48	0.52	0.35	0.52	0.54	0.31	0.34	0.00	0.00	0.00
Na2O	3.86	3.88	4.06	3.75	3.73	3.83	4.23	4.72	0.38	0.41	0.38
K2O	4.47	3.85	4.04	4.34	4.36	4.69	2.35	3.11	2.01	1.22	1.13
P2O5	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.02	0.04	0.03
LOI	0.53	0.37	0.51	0.43	0.44	0.39	0.99	0.35	1.33	1.65	1.64

Trace elements

Cr	10	0	0	0	0	0	0	0	0	14	0
Ni	18	35	14	16	15	0	0	16	26	20	11
V	17	11	0	0	0	0	0	10	0	0	236
Pb	13	17	15	11	11	18	31	12	0	375	0
S	0	800	800	400	0	0	0	0	1600	7360	8280
Rb	488	459	472	354	419	440	280	278	725	431	399
Ba	33	71	58	31	61	56	34	100	37	40	54
Sr	5	6	3	3	6	4	10	0	0	0	0
Ta	-	-	11.32	-	12.78	-	-	-	-	6.66	0
Nb	66	72	73	95	132.80	89	68	82	76	78	66
Hf	-	-	1.51	-	4.5	-	-	-	-	9.01	-
Zr	100	74	25	126	123	77	101	159	85	302	228
Y	126	130	116	95	143	111	85	332	11	42	41
Th	51	40	33	73	54	40	58	64	39	48	55
U	22	27	36	28	24	27	19	15	53	54	43
Rb/Sr	97.60	76.50	157.30	118.00	69.83	110.00	28.00	-	-	-	-
Ba/Sr	6.60	11.83	19.33	10.33	10.17	14.00	3.40	-	-	-	-
Ba/Rb	0.07	0.15	0.12	0.09	0.15	0.13	0.12	0.09	0.05	0.09	0.14
Th/U	2.32	1.48	0.92	2.62	2.25	1.48	3.05	0.91	0.74	0.89	1.28

Normative composition

Qtz	36.40	39.04	36.26	38.35	36.39	34.46	44.20	37.12	61.85	73.08	72.74
Crn	0.73	0.73	0.80	0.78	0.84	0.62	2.70	1.02	9.94	9.81	11.24
Zrn	0.02	0.02	0.00	0.03	0.02	0.02	0.02	0.03	0.02	0.06	0.05
Or	26.54	23.13	24.14	26.07	26.12	28.08	14.11	18.72	12.44	7.51	7.02
Ab	32.58	33.12	34.47	32.07	31.90	32.63	36.08	40.43	3.29	3.53	3.30
An	2.46	2.40	2.59	1.76	2.54	2.70	1.49	1.71	0.00	0.00	0.00
Di	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Hy	0.00	0.63	1.38	0.19	1.86	0.84	0.00	0.00	8.64	0.00	0.00
Mag	0.00	0.88	0.32	0.70	0.12	0.84	1.20	0.00	3.84	3.96	3.78
Hem	1.23	0.00	0.00	0.00	0.00	0.00	0.01	0.99	0.00	1.90	1.72
Ilm	0.05	0.11	0.10	0.12	0.15	0.11	0.19	0.00	0.16	0.41	0.35
Ap	0.02	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.05	0.10	0.07
Rut	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.00	0.00

D.I.	95.52	95.29	94.87	96.49	94.50	95.17	94.39	96.27	77.58	84.12	83.01
ASI	1.06	1.09	1.07	1.07	1.07	1.05	1.27	1.09	4.46	5.79	6.93
A.I.	0.87	0.87	0.86	0.88	0.86	0.88	0.75	0.87	0.22	0.17	0.14

	36C
SiO2	79.21
TiO2	0.07
Al2O3	10.59
Fe2O3	1.65
FeO	1.40
MnO	0.12
MgO	0.00
CaO	0.09
Na2O	0.37
K2O	4.56
P2O5	0.00
LOI	1.61

Trace elements

Cr	0
Ni	14
V	12
Pb	96
S	1600
Rb	838
Ba	126
Sr	0
Ta	-
Nb	59
Hf	-
Zr	110
Y	29
Th	53
U	70
Rb/Sr	-
Ba/Sr	-
Ba/Rb	0.15
Th/U	0.76

Normative composition

Qtz	59.78
Crn	4.93
Zrn	0.02
Or	27.81
Ab	3.19
An	0.46
Di	0.00
Hy	1.34
Mag	2.44
Hem	0.00
Ilm	0.14
Ap	0.00
Rut	0.00
D.I.	90.76
ASI	1.85
A.I.	0.52

Table 2.3: Rare-earth-element data (in ppm) for selected rock samples of the Kwandonkaya complex: **Buri ring dyke:** rhyolite porphyry (49A), and hedenbergite granite porphyry (43); **Cone sheets:** felsite porphyry (61A), hastingsite granite porphyry (40A), and hedenbergite fayalite granite porphyry (64); **Jubga-Ziem center:** hypersolvus (fayalite) syenogranite (58), and "hornblende"-biotite granite porphyry (56, and 68A); **Biotite granites:** coarse-grained biotite granite (26A), drusy biotite granite (30A, and 37), and porphyritic biotite granite (1B, and 36A); albitized biotite granite (23C, and 35A), and greisen (36B). **Chon:** chondrite normalization values (after Haskin & Haskin 1968, Nakamura 1974, Rock 1987). These data were obtained by inductively coupled plasma - mass spectrometry (ICP - MS) (solution mode).

#	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	(La/Lu) _N	Eu/Eu*	ΣREE
49F	102.37	221.16	25.66	96.15	24.16	1.03	24.55	3.87	26.52	4.81	13.49	1.88	10.93	1.57	7.73	0.13	558.15
43	161.42	325.95	38.54	140.96	30.94	1.12	30.01	4.19	26.86	4.71	12.28	1.66	10.41	1.47	13.02	0.11	790.52
61A	45.23	80.16	10.06	33.38	8.30	0.24	7.57	1.33	9.63	1.85	5.59	0.91	6.28	0.87	6.16	0.09	211.40
40A	150.95	306.19	35.83	132.11	27.04	1.80	25.50	3.59	23.49	4.15	11.60	1.62	9.94	1.39	12.87	0.21	735.20
64	64.56	139.83	16.64	63.24	16.83	0.69	19.14	3.51	24.91	4.82	14.01	1.92	11.21	1.55	4.94	0.12	382.86
58	70.13	150.34	18.27	69.39	19.57	0.56	23.76	4.41	31.98	6.25	18.42	2.55	14.86	2.00	4.16	0.08	432.49
56	125.77	247.99	29.52	112.24	24.78	1.83	24.12	3.82	25.41	4.72	12.90	1.72	10.39	1.43	10.43	0.23	626.64
68A	73.29	216.09	21.25	82.74	20.78	0.77	19.74	3.10	20.45	3.79	10.33	1.38	8.37	1.11	7.83	0.12	483.19
26A	81.23	155.75	16.81	58.42	13.42	0.66	14.63	2.57	19.37	3.75	11.67	1.76	11.16	1.53	6.29	0.14	392.73
30A	21.62	55.59	6.87	23.76	7.17	0.04	7.65	1.69	13.78	3.01	10.77	1.93	13.15	1.86	1.38	0.02	168.89
37	75.63	141.61	15.66	47.60	11.17	0.17	12.01	2.26	17.53	3.50	12.13	2.20	15.86	2.15	4.17	0.05	359.48
1B	65.52	138.71	14.82	48.58	10.17	0.48	9.31	1.64	12.48	2.77	9.10	1.50	10.05	1.45	5.36	0.15	326.58
36A	40.11	81.06	9.04	30.95	7.39	0.46	8.04	1.58	12.13	2.55	8.48	1.35	8.71	1.21	3.93	0.18	213.06
35A	40.51	89.02	10.80	37.30	10.57	0.17	10.94	2.14	16.87	3.51	11.89	1.96	14.03	1.92	2.50	0.05	251.63
23C	21.94	48.38	7.30	26.83	7.71	0.08	8.66	1.68	13.70	3.00	10.31	1.81	12.35	1.76	1.48	0.03	165.51
36B	65.45	148.50	14.02	44.92	9.26	0.35	7.64	1.28	8.95	1.80	5.71	0.94	6.23	0.90	8.62	0.12	315.95
Chon	0.329	0.865	0.112	0.630	0.203	0.077	0.276	0.047	0.343	0.07	0.225	0.03	0.22	0.039	1.00	1.00	3.466

Table 2.4: A comparison of characteristics of low-phosphorus subtype of topaz granites (cf. Taylor 1992) and topaz-bearing biotite granites in the Kwandonkaya complex.

Feature	Low-phosphorus subtype of topaz granites (Christiansen et al. 1986, Taylor 1992).	Topaz-bearing biotite granites of Kwandonkaya complex
Essential minerals	Qtz, Kspar, Ab($An_{0.5}$), F-rich Toz, Li-mica (Li-Srp, Znw, Lp).	Qtz, Kspar (Or, Mc), Ab($An_{0.4}$), Toz, Li-mica (Fp, Srp, Znw, Trl)
Accessory minerals	Zrn, Nb-Rut, Col, Tan, Cst, Mnz, sulfides, rare phosphates	Cst, Col, Zrn, Ilm, Mon, Fl, Fct
Major elements	Low: TiO_2 (<0.1%), MgO (<0.2%), CaO (<0.5%), ΣFeO (<1.0%), P_2O_5 (<0.10%), $12 \leq Al_2O_3 \leq 14.5\%$, and Cl (<2500ppm). High: Na_2O (>4.0%), SiO_2 (>73%), and F (>0.4%, up to 2.5%). Mildly peraluminous ($1.0 \leq A/CNK \leq 1.2$)	Low: $0.05 \leq TiO_2 \leq 0.08\%$, $MgO = 0\%$, $0.37 \leq CaO \leq 0.52\%$, $0.17 \leq FeO \leq 1.04\%$, $P_2O_5 = 0\%$, $12.86 \leq Al_2O_3 \leq 13.39\%$, inferred Cl (from mica chemistry) Moderate: $3.49 \leq Na_2O \leq 4.07\%$ High: $76.82 \leq SiO_2 \leq 77.78\%$, inferred $Log(f_{HF}/f_{HCl})$ Mildly peraluminous ($1.07 \leq A/CNK \leq 1.1$, up to 1.24)
Minor/Trace elements	High: Li (250 - 2600 ppm), Rb (800 - 2400 ppm), LILE ($20 < Nb < 120$, $10 < Ta < 60$ up to 250 ppm) Very low: V, Cr, Ni, Sr, Ba	High: inferred Li (Ch. 4), $260 \leq Rb \leq 600$, Moderate to high LILE: $73 \leq Nb \leq 195$, $11 \leq Ta \leq 24$ (ppm) low: $0 \leq V \leq 24$, $0 \leq Cr \leq 12$, $0 \leq Ni \leq 26$, $0 \leq Sr \leq 8$, $0 \leq Ba \leq 72$ (ppm)
REE, Y, Th	$150 \leq \Sigma REE \leq 350$ ppm, gullwing-shaped REE pattern, $0.6 \leq (La/Lu)_N \leq 1.5$, Deep negative Eu-anomaly: $(Eu/Eu^*) < 0.1$, high Y ($60 \leq Y \leq 160$ ppm), moderate to high Th ($20 \leq Th \leq 50$ ppm).	$161.5 \leq \Sigma REE \leq 359.5$ (ppm), Gullwing shaped REE pattern (Ch. 2), $1.2 \leq (La/Lu)_N \leq 3.62$, Moderate negative Eu-anomaly: $0.52 \leq Eu/Eu^* \leq 2.21$. $80 \leq Y \leq 157$ (ppm) Moderate to high Th: $33 \leq Th \leq 61$ (ppm)

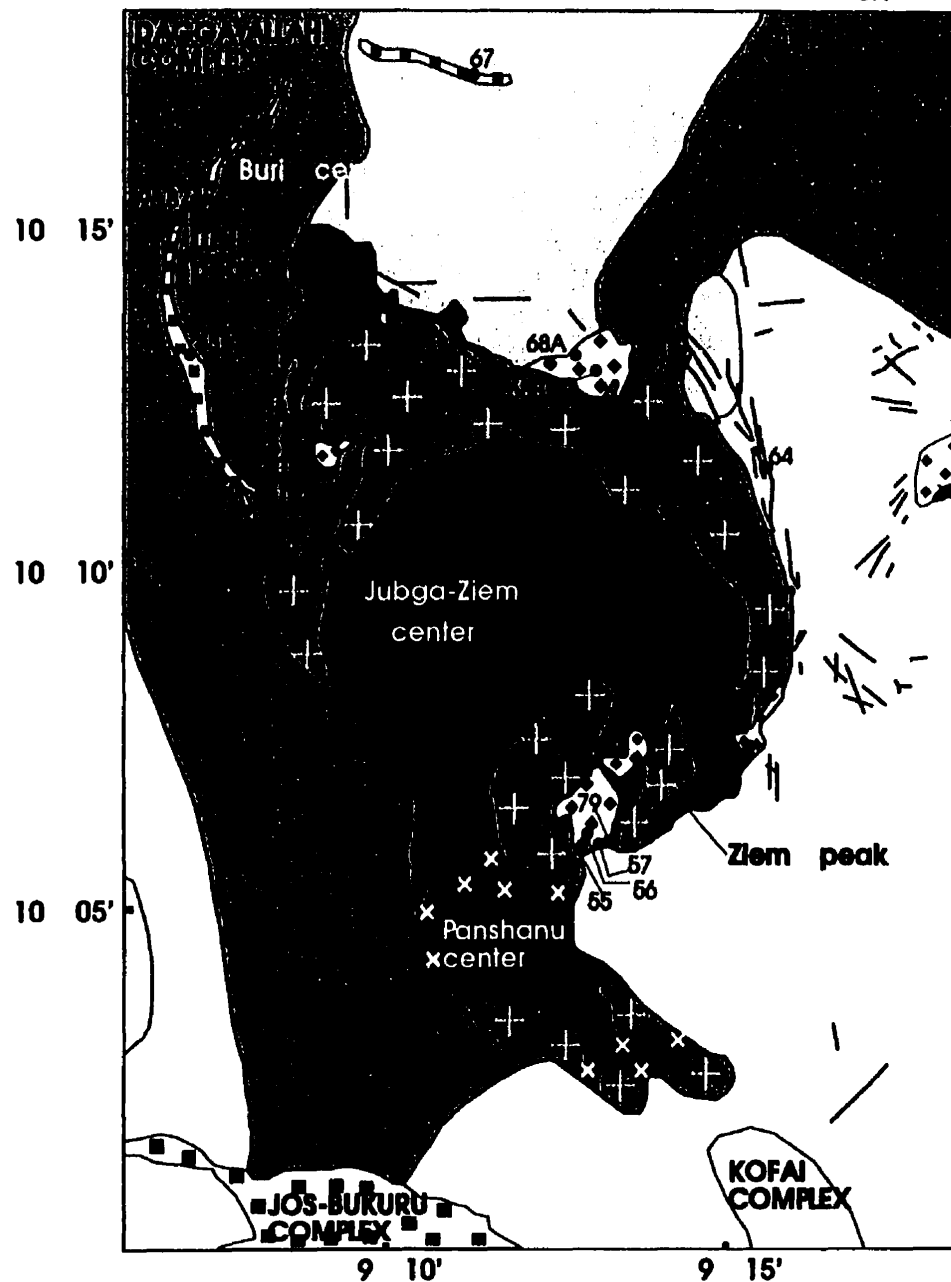
Chapter 3: Olivine, pyroxene, and amphibole chemistry as a key to magmatic to postmagmatic history of the subvolcanic rocks.

3.1 Introduction

The Nigerian Younger Granites (henceforth called NYG; Fig. 1.1) contain a distinctive suite of Fe-rich mafic minerals. Thus, in addition to the structures of the plutons, their tectonic setting, petrochemistry, and patterns of mineralization (Kinnaird 1985, Bowden et al. 1987), the mafic assemblages also serve to distinguish the NYG from the Pan-African (Older) Granites of Nigeria. Subvolcanic rocks in ring dykes and cone sheets of ring complexes are characterized by fayalite and hedenbergite as the principal mafic minerals. Peralkaline complexes in the NYG contain aegirine, arfvedsonite ($\pm$ riebeckite), accessory astrophyllite and aenigmatite as mafic phases, whereas metaluminous to peraluminous granites are characterized by calcic, and calcic-sodic amphiboles, and an extensive range of mica compositions, from siderophyllite-annite to zinnwaldite-lithiophyllite (Bennett et al. 1984, Bowden et al. 1987). Aegirine and arfvedsonite in subvolcanic rocks are restricted to peralkaline ring dykes, such as in the Dago complex (Fig. 1.1; Ike & Sakoma 1985).

In this Chapter, the composition of ferromagnesian minerals in various subvolcanic units of the Kwandonkaya complex (Fig. 3.1) is used as an indicator of degree of evolution and environment of crystallization. These minerals are sensitive indicators of bulk-rock chemistry, fugacity of oxygen $f(\text{O}_2)$, $P(\text{H}_2\text{O})$, temperature (T), and confining pressure (P) during crystallization (i.e., at the magmatic stage) and extent of subsolidus re-equilibration. The order of magmatic crystallization and the patterns of hydrothermal alteration are established in this study using results of petrographic study, scanning electron microscopy (SEM), and electron-microprobe studies of olivine, pyroxene, and amphibole in ring dykes and cone sheets of the Kwandonkaya complex.

Fig. 3.1: Geological map of the Kwandonkaya complex, with location of samples used in this study. This map was modified from Buchanan (1971). Only the samples characterized by electron-microprobe analysis in this study are shown.



• 1 Sample location

MESOZOIC GRANITES

G: Gneiss

Basalt dyke

Fine- to medium-grained biotite granite

Coarse-grained biotite granite

Porphyritic biotite microgranite

"Homblende"-biotite, biotite granite porphyry, +/- fayalite syenogranite

Hedenbergite fayalite granite porphyry

Rhyolite porphyry

Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

Biotite homblende granite

Gneissic granite

Migmatite

1 0 2 4 6 8 km

These results complement previous work on ferromagnesian minerals undertaken in other ring complexes of the NYG. Borley (1963, 1976) undertook the first and only study of ferromagnesian minerals from selected complexes of the NYG (Kwandonkaya not included). Ike et al. (1984, 1985) investigated the mafic mineralogy (olivine, pyroxene, and amphibole) of the Tibchi complex; apart from magmatic trends, these authors provided evidence for the postmagmatic initiation of peralkalinity by the late appearance of alkali amphiboles. Studies of mafic minerals also were part of geological - geochemical investigations of ring complexes (e.g., Bennett et al. 1984, Bowden et al. 1987).

3.2 Distribution of subvolcanic rocks

The Kwandonkaya complex consists predominantly of peraluminous biotite granite, with minor metaluminous rocks preserved in ring dykes. Volcanic rocks, preserved within inner parts of some ring dykes owing to cauldron subsidence, are not preserved in this complex. Detailed field relations, petrography, and textural interpretations of the different rock-units of the Kwandonkaya complex have been previously presented (Chapter 2). In this section, the mode of occurrence of subvolcanic rocks is emphasized.

Rhyolite porphyry is limited in occurrence to the Buri center, exposed only north of Dutsin Lusi, forming an extension, as shown in Fig. 3.1, from the main ring dyke of hedenbergite fayalite granite porphyry. The rhyolite porphyry exposure has a consistent distribution of enclaves, which comprise 15 to 40% of the rock. Overall, the rhyolite porphyry forms a minor part (4%) of a ring dyke possessing sharp contacts with the Pan-African basement, and with hedenbergite fayalite granite porphyry. The hedenbergite fayalite granite porphyry also occurs only in the Buri center. It constitutes 96% of total exposed area of the Buri ring dyke. The intrusion has sharply defined contacts with the Pan-African basement, and irregularly distributed enclaves varying from

nil, in about 90% of the exposures, to 20% in localized areas.

Cone sheets, emplaced into migmatites and gneissic granite, vary from less than one meter across to swarms covering a width of over 100 m, with the highest concentration in the northeastern corner of the Jubga-Ziem center. Hedenbergite hastingsite granite porphyry, hastingsite granite porphyry, and leucocratic granite porphyry are found in minor intrusive bodies in low-lying ridges, together with the swarms of felsite porphyry.

Hypersolvus (fayalite) syenogranite, "hornblende" (quotation marks indicating a provisional name that conforms to previous statements in the literature on NYG) biotite granite porphyry and biotite granite porphyry are preserved as relics of a ring dyke in marginal parts of the Jubga-Ziem center. Coarse granular hypersolvus syenogranite occurs in inner parts of some bodies of "hornblende"-biotite granite, with biotite granite porphyry capping some such bodies. Other occurrences of fayalite syenogranite are drusy and fine-grained. Typically, the samples of "hornblende"-biotite granite porphyry vary from drusy, porphyritic types along contact zones with the Pan-African basement or the main biotite granite, to coarse granular hypersolvus syenogranite, or "hornblende"-biotite granite porphyry inward from the contact. Biotite granite porphyry, on the other hand, occurs in upper sections of hypersolvus syenogranite and "hornblende" biotite granite porphyry, and as "isolated" roof pendants forming the Ziem peak. The hypersolvus syenogranite, "hornblende"-biotite granite porphyry, and biotite granite porphyry occur in direct contact with coarse-grained biotite granite, basement country-rock, and fine- to medium-grained biotite granite.

Coarse-, and medium- to fine-grained biotite granite occur as the predominant rock-types in the Jubga-Ziem and Panshanu centers. These are exposed in their roof zones, within the Jubga-Ziem center, in contact with roof zones of relict subvolcanic rocks in ring dyke. These, along with felsite porphyry cone-sheets, do not contain olivine, pyroxene, and amphibole, and hence are not discussed further in this chapter.

3.3 Petrography of the subvolcanic rocks

A summary of textures and mineralogy of the subvolcanic rocks in the Kwandonkaya complex is presented in Table 3.1, with emphasis on textures and sequence of crystallization of olivine, pyroxene, and amphibole. The rock units are discussed in their relative sequence of intrusion, as in Table 3.1, and as established in earlier studies (Chapter 2). The sequence is 1) Buri ring dyke: rhyolite porphyry and hedenbergite fayalite granite porphyry; 2) Jubga-Ziem center: cone sheets (hedenbergite hastingsite granite porphyry and hastingsite granite porphyry), and 3) Jubga-Ziem ring dyke [hypersolvus (fayalite) syenogranite and "hornblende"-biotite granite porphyry].

Rhyolite porphyry in the Buri ring dyke is composed of microphenocrysts of fayalite (<0.5%), hedenbergite (1 - 5%), orthoclase microperthite (25%), amoeboid quartz (20%), in an aphanitic groundmass composed of quartz and alkali feldspar (30 - 40%), mafic enclaves (15 - 20%), and accessory minerals (Table 3.1). Fayalite forms distinct phenocrysts in the groundmass, and in some sections seems intergrown with feldspar. The fayalite is incipiently altered to iddingsite, chlorite, magnetite, and quartz. Hedenbergite occurs as distinct phenocrysts (ranging from 0.3 x 0.3 mm to 0.8 x 1.0 mm). Some phenocrysts are optically zoned, with a brownish green core and a pale green rim. Hedenbergite is partly replaced by Fe-Ti oxides, chlorite, and Fe-rich mica. Both fayalite and hedenbergite largely crystallized before alkali feldspar and quartz, but after zircon and chevkinite. Amphibole in this rock type occurs principally as a secondary phase after olivine or hedenbergite. Rare amphibole microphenocrysts are altered to chlorite.

The *hedenbergite fayalite granite porphyry*, also found in the Jubga-Ziem center, has the composition: chevkinite (< 0.2%), zircon (0.3%), fayalite (<0.5%), hedenbergite (3-5%), orthoclase microperthite (45%), amoeboid quartz (20%), and a fine-grained groundmass (30 to 50%). Fayalite forms anhedral grains included within microperthitic orthoclase phenocrysts.

Table 3.1: Summary of petrographic features for subvolcanic rocks of Kwandonkaya complex.

Rock Type	Texture	Essential minerals	Mafic Mineral(s)	Accessory minerals
Rhyolite porphyry	Porphyritic	Or, Qtz phenocrysts	Fa, Hd phenocrysts (Cam)	Ckt, Zrn, Ilm, Mag, Ttn, Ap, (Fl, Aln)
Hd Fa granite porphyry	Glomero-porphyritic	Or, Qtz phenocrysts	Hd, Fa phenocrysts (Cam, Bt, Chl)	Zrn, Ilm, Ckt, Ap (Mag, Fl, Aln)
Felsite porphyry	Aphanitic, porphyritic	Or, Qtz phenocrysts	(Bt, Chl)	Aln, Zrn, Ilm, Mag, (Fl, Hem)
Hd Hs granite porphyry	Glomero-porphyritic	Or, Qtz phenocrysts	Hd, Hs	Aln, Mag, Zrn, (Fl, Hem, Ept)
Hs granite porphyry	Glomero-porphyritic	Or, Qtz phenocrysts	Hs (Bt, Chl)	Ckt, Aln, Zrn, Ttn, Ap, Ilm, (Mag, Fl, Brt, Aln)
Fa syenogranite	Drusy, fine-grained, coarse granular	Or, Qtz, (minor Ab)	Fa phenocrysts, (Grn, Bt, Chl, Idn)	Ilm, Mag, Ckt, Aln, Zrn, Fl, (Mag, Mnz, Xnt, Aln, Fct, Fst, Prt, Hem)
Hbl Bt granite porphyry	Drusy, glomero-porphyritic	Or, Qtz phenocrysts	Hbl, Bt phenocrysts	Aln, Ilm, Mag, Zrn
Bt granite porphyry	Glomero-porphyritic	Qtz, Or phenocrysts	Bt	Zrn, Mag

Mineral abbreviations used (mainly from Kretz 1983) are: **Ab**: albite, **Aln**: allanite, **Ap**: apatite, **Brt**: barite, **Bst**: bastnäsite, **Bt**: biotite, **Cam**: calcic amphibole, **Chl**: chlorite, **Ckt**: chevkinite, **Ept**: epidote, **Fa**: fayalite, **Fct**: fluocerite, **Fl**: fluorite, **Fst**: fergusonite, **Grn**: grunerite, **Hbl**: hornblende, **Hd**: hedenbergite, **Hem**: hematite, **Hs**: hastingsite, **Ilm**: ilmenite, **Mag**: magnetite, **Mnz**: monazite, **Prt**: parisite, **Qtz**: quartz, **Rt**: rutile, **Tht**: thorite, **Ttn**: titanite, **Xnt**: xenotime, **Zrn**: zircon. Parentheses enclose secondary minerals.

Hedenbergite occurs as euhedral to subhedral inclusions in feldspar (Fig. 3.2), as well as in distinct phenocrysts. The hedenbergite grains vary in size from 0.3 x 0.3 mm to 0.8 x 1.8 mm, and invariably contain inclusions of chevkinite, zircon, and magnetite close to or along the margin. Hedenbergite grains are replaced to variable extents by amphibole, Fe-rich mica, chlorite, magnetite, and fluorite. Fayalite and hedenbergite in this rock unit, like in the case of rhyolite porphyry, crystallized before alkali feldspar and quartz, but after chevkinite and zircon.

Within cone sheets of the Jubga-Ziem center, the *hedenbergite hastingite granite porphyry* contains hedenbergite intergrown with a subhedral late-magmatic hastingitic hornblende. In thin section, relict inclusions of the hedenbergite (partly replaced by Fe-rich mica, chlorite and epidote) occur within some parts of the amphibole, suggesting that the amphibole formed by reaction of melt with pyroxene. The hedenbergite crystallized before the alkali feldspar. Grains of hastingitic hornblende range from 1 x 4 mm to 1.5 x 6 mm, with an extinction angle ($Z_{\wedge c}$) of 20°; this is intergrown with hedenbergite, which it partly replaces. A ragged rim on hastingitic hornblende is intergrown with the microperthitic feldspar, and is interpreted to have formed after its exsolution. The *hastingite granite porphyry*, also present in the cone sheets, contains clusters of hastingitic hornblende grains, most of which have an irregular, serrated interface against microperthitic feldspar phenocrysts (Fig. 3.3), whereas boundaries with the granophyric groundmass are sharp and planar. Individual grains vary from 0.5 x 2.5 mm to 1.5 x 3 mm, and aggregates, up to 3 x 9 mm. Extinction angles ($Z_{\wedge c}$) range from 20° to 30°. The amphibole grains in this rock type invariably have inclusions of accessory minerals listed in Table 3.1. The principal minerals replacing amphibole are chlorite, Fe-rich mica and magnetite.

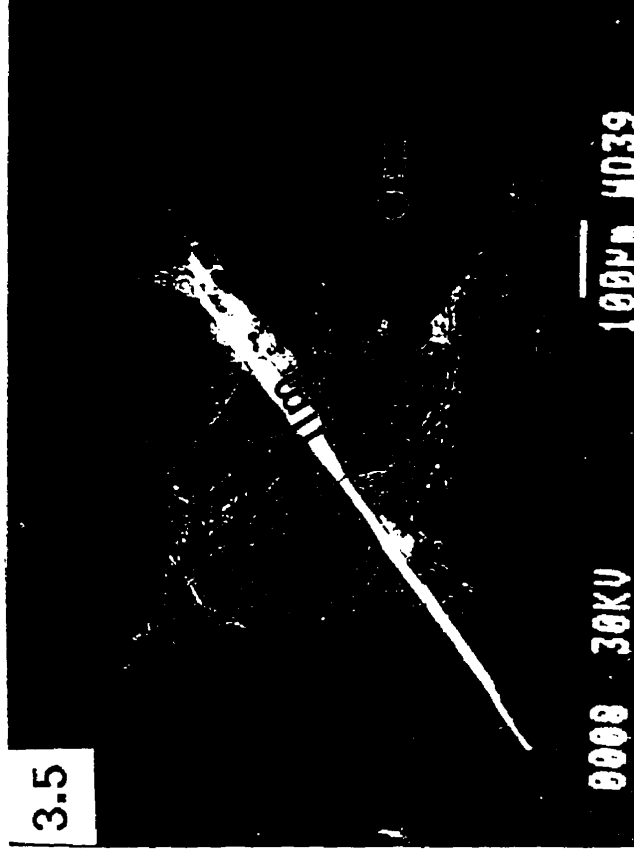
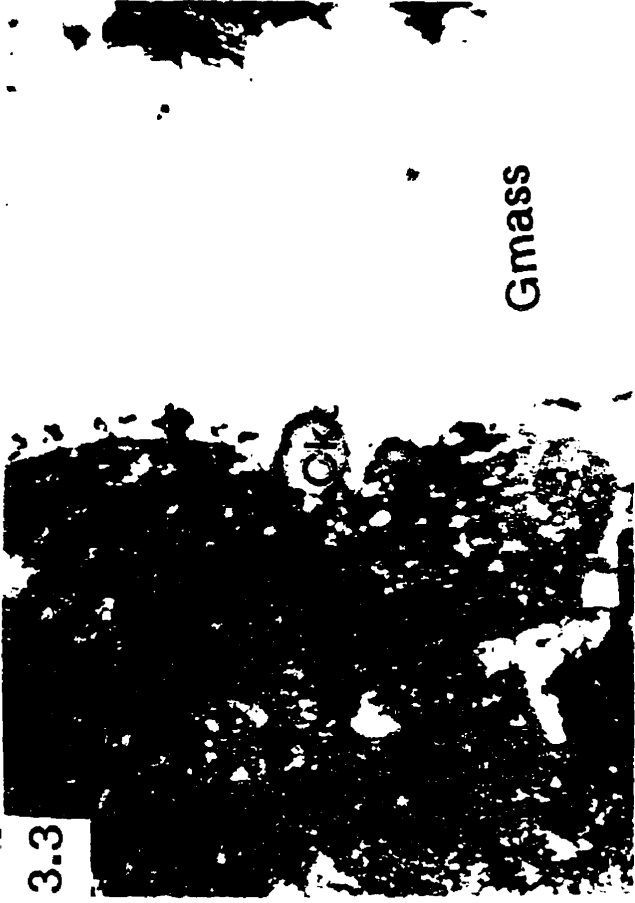
The *hypersolvus (fayalite) syenogranite* in the Jubga-Ziem ring dyke contains the following minerals, listed with their relative proportions and in the sequence of crystallization: fayalite (2%), orthoclase microperthite (70%), quartz (25%), interstitial granophyre (2%), and

Fig. 3.2: Photomicrograph of anhedral (rounded) hedenbergite (Hed) inclusions within simply twinned orthoclase microperthite. The rounding of the hedenbergite phenocrysts could have resulted from the crystallization of alkali feldspar, which terminated the growth of the clinopyroxene. Accessory mineral inclusions in hedenbergite include zircon, chevkinite, and Fe-Ti oxides. The rock is hedenbergite fayalite granite porphyry (sample 43, Fig. 3.1).

Fig. 3.3: Photomicrograph of multigranular aggregates of hastingsitic hornblende (Amp) within hastingsite granite porphyry (sample 64, Fig. 3.1). The amphibole grains typically contain inclusions of chevkinite (Ckt), allanite, and Fe-Ti oxides as well as wavy boundaries with alkali feldspar. The right half of the picture is the microcrystalline groundmass (Gmass) composed of quartz and alkali feldspar.

Fig. 3.4: Photomicrograph fayalite replaced principally by iddingsite. The rock is hypersolvus (fayalite) syenogranite (sample 58, Fig. 3.1). Or = orthoclase, Qtz = quartz.

Fig. 3.5: A SEM-BSE (scanning electron microscope - back-scattered-electron) image of replaced fayalite (Fa). The fayalite is replaced by Fe-rich chlorite + iddingsite, intergrown with acicular ilmenite (Ilm) and both enclosed in multigranular microperthite (Per) and quartz (Qtz). Anastomosing brighter microveins have higher levels of Fe (thuringite) than the dominant darker grey part (stilpnomelane + iddingsite). Also, secondary annitic mica, fergusonite, and fluorite occur along the margin of replaced fayalite. The rock is fayalite syenogranite (sample 58, Fig. 3.1). The width of the field of view in photomicrographs is 3.1 mm.



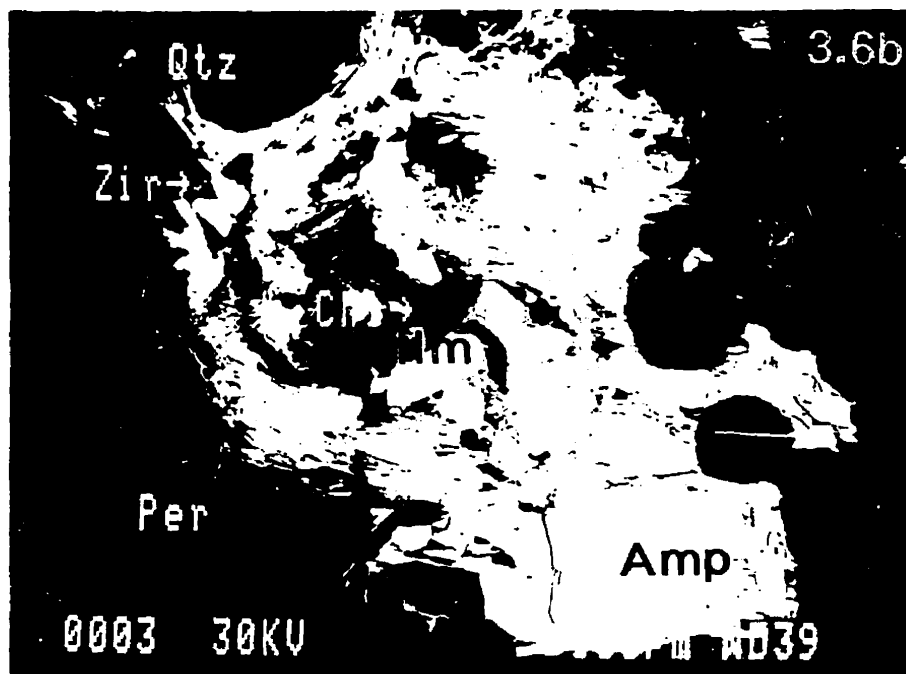
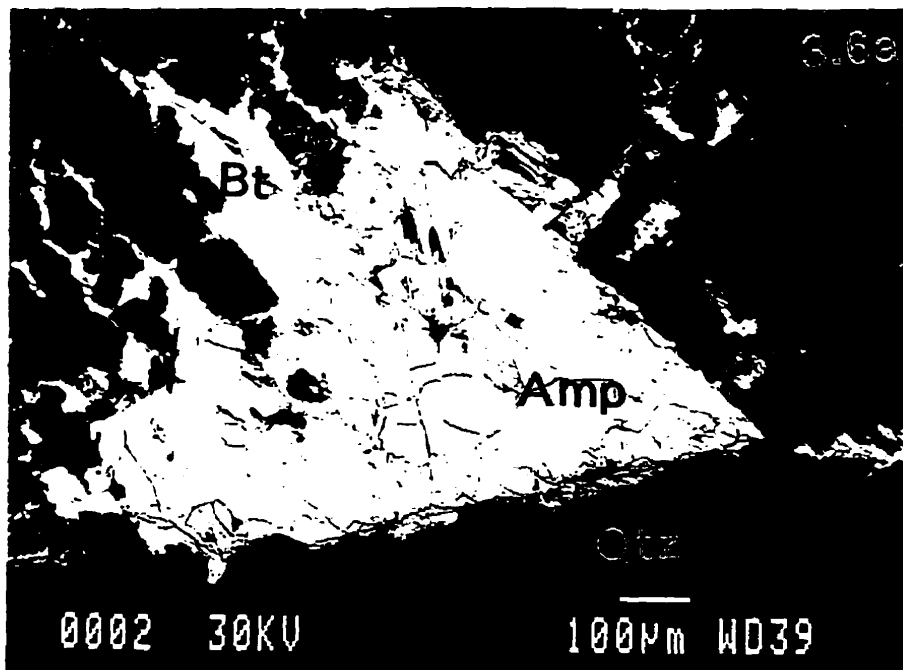
other accessory minerals listed in Table 3.1 (<1%). The fayalite is interstitial, ranging from less than 0.1 x 0.1 mm to 1 x 1.5 mm; it also occurs as an inclusion within feldspar. Generally, fayalite is replaced to different degrees either by iddingsite (Fig. 3.4), or Fe-rich chlorite, annitic mica, quartz $\pm$ magnetite, $\pm$ grunerite, $\pm$ fluorite. Iddingsite pseudomorphs after fayalite are intergrown with ilmenite, zircon, annitic mica, and fergusonite (Fig. 3.5). The grunerite has an extinction angle ($Z\wedge c$) of 11° , from which a tentative composition $\geq \text{Fe}_{97}$ is inferred (after Bowen & Schairer 1935).

"Hornblende"-biotite granite porphyry, also exposed in the Jubga-Ziem center, contains: orthoclase microperthite (30-60%), quartz (25-35%), hornblende (2-4%), allanite (<0.5%), biotite (<1%), Fe-Ti oxides (<0.5%), and interstitial fine-grained groundmass (20 to 50%). Grains of amphibole range from less than 0.1 x 0.1 mm to 2 x 5 mm, with aggregates reaching over 3 x 7 mm. In some instances, polycrystalline clots of euhedral to subhedral amphibole are enclosed in mica (Fig. 3.6a). The enclosure of these polycrystalline clots by mica suggests that the amphibole crystallized at late magmatic stages, and postdates the alkali feldspar but predates the mica. In a few examples, grains of amphibole enclose hedenbergite (completely replaced by magnetite, chlorite, quartz $\pm$ fluorite) (Fig. 3.6b). This distribution also suggests a late magmatic crystallization by reaction of clinopyroxene with residual melt to produce amphibole in the Jubga-Ziem ring dyke. Because the amphibole formed by replacement of hedenbergite, and that formed by direct crystallization from volatile-rich melt, are not associated, their relative sequence of crystallization is not discernible.

3.4 Mineral Chemistry

The analytical methods employed are briefly outlined. This is followed by a presentation of mineral chemical variations of fayalite, hedenbergite, and calcic amphibole.

Fig. 3.6: Petrographic features of amphibole in the Jubga-Ziem ring dyke. a) A SEM-BSE image of polycrystalline, euhedral to subhedral grains of amphibole (Amp), which is brighter, enclosed by biotite (Bt), which is darker, with prominent cleavage. One half of the biotite grain is subhedral, whereas the other is ragged and pervades fractures within microperthite. This suggests late-magmatic crystallization in a drusy cavity, with both amphibole and biotite postdating the formation of alkali feldspar in subvolcanic rocks. The section is from "hornblende"-biotite granite porphyry (sample 68B, Fig. 3.1). **b)** A BSE-SEM image of amphibole (Amp) enclosing hedenbergite (completely replaced by Fe-rich chlorite). Euhedral quartz (black) along the margins is incorporated in amphibole. Associated minerals include ilmenite (Ilm), fluorite (Fl), and zircon (Zir). The rock type is "hornblende"-biotite granite porphyry (sample 68B, Fig. 3.1).



3.4.1 Analytical methods

The olivine, pyroxene, and amphibole in these rocks were analyzed by wavelength-dispersion spectrometry (WDS) using a Cameca CAMEBAX electron microprobe at McGill University. An accelerating voltage of 15 kV and beam current of 20 nA were used, with counting times of 100 seconds for F, 50 seconds for Cl, and 25 seconds for other elements. The following standards were used: andradite (Fe), MnTi (Mn, Ti), albite (Na), diopside (Si, Mg, and Ca), orthoclase (K, Al), fluorite (F), VANA (Cl), NiO (Ni), Cr₂O₃ (Cr), and willemite (Zn). Analyses were run using the ONQANT software package. Mineral chemical data presented in this study are average results (except where otherwise indicated) of spot analyses, typically of 4 to 12 grains (Tables 3.2 to 3.4). Reconnaissance scanning electron microscopy and energy-dispersion spectroscopy (SEM-EDS) were undertaken to identify inclusions, and products of alteration, or to establish the pattern of zoning using the JEOL SEM at McGill University, and in some cases with the EDS system attached to the electron microprobe.

Total iron, determined as FeO, was recalculated to FeO and Fe₂O₃ (for fayalite and hedenbergite) after the method of Droop (1987). This method utilizes the general principle of normalization to a fixed number of anions and to a fixed number of cations (based on the ideal mineral formula), and then adjusting Fe³⁺/Fe²⁺ ratio to obtain electroneutrality. Results of the pyroxene analyses were evaluated using the "criteria for acceptable pyroxene analyses" of Cameron & Papike (1980), which include specifications for charge balance, T-, M1-, and M2-occupancy, and application of the criteria by Robinson (1980). Site occupancies were calculated using the scheme proposed by Cameron & Papike (1980). The nomenclature of pyroxenes follows the usage proposed by the IMA Subcommittee on pyroxenes (Morimoto 1989). For amphiboles, the formula was normalized to 24(O, OH, F, Cl), and Fe (total) is reported as Fe²⁺. Other approaches of recalculation based on normalization of the formula to a fixed number of cations

(e.g., 13, 15, and 16) all have flaws that result from underlying assumptions. These assumptions tend to, according to Hawthorne (1983), obscure certain schemes of substitution, and make results of certain analyses look more acceptable. The nomenclature of amphiboles follows the recommendations of the IMA Subcommittee on amphiboles (Leake 1978), and of Hawthorne (1981, 1983).

3.4.2 *Fayalite*

Averaged compositions of fayalite from the Kwandonkaya complex are given in Table 3.2. The metaluminous rhyolite porphyry in the Buri ring dyke contains fayalite with an average composition: $\text{Fa}_{94.2}\text{Fo}_{3.2}\text{Tp}_{2.6}$, $\text{F/FM} [\text{Fe}/(\text{Fe}+\text{Mg})] = 0.97$. Only a few of the analyses from this rock revealed detectable traces of Cr (0.03% Cr_2O_3) and Ni (0.01% NiO). Calculations show that the fayalite has no laihunite (i.e., Fe^{3+}) component. In the fayalite syenogranite from the Jubga-Ziem center, the fayalite has an even more evolved average composition: $\text{Fa}_{97.9}\text{Fo}_{0.3}\text{Tp}_{1.8}$, $\text{F/FM} = 0.99$. In this case, the fayalite has no traces of Cr or Ni, but does seem to contain between 0.2 to 1.7% of the laihunite component, which occurs by the substitution: $3\text{Fe}^{2+} = 2\text{Fe}^{3+} + \square$, where $\square$ is a vacancy. The increase in F/FM ratio of fayalite, from 0.97 in the rhyolite porphyry to 0.99 in the syenogranite, is accompanied by a decrease in the level of tephroite ($\text{Tp}_{2.6}$ to $\text{Tp}_{1.8}$) and calcium (from 0.33 to 0.09% CaO). Similar Fe-rich compositions of fayalite also have been described from subvolcanic rocks in other ring complexes in Nigeria, with quoted Fa contents including Mn (Tibchi: $\text{Fa}_{91.97}$, Ike et al. 1984; Dutse: $\text{Fa}_{94.95}$, and Shira: Fa_{99} , Bennett 1981, Bennett et al. 1984).

3.4.3 *Pyroxene*

The results of pyroxene analyses, principally from the Buri ring dyke, are presented in

Table 3.3. Compositions of core and rim of grains within the same thin section were averaged separately. The pyroxene compositions fall in the field of quadrilateral ("Quad") pyroxenes in the Q-J plot (not shown) of Morimoto (1989), where $Q = \Sigma(\text{Ca}+\text{Mg}+\text{Fe})$, the quadrilateral components, and $J = 2\text{Na}$. In the pyroxene quadrilateral (Fig. 3.7), the compositions plot close to or at the hedenbergite corner. Generally, there is a low degree of Na-for-Ca substitution in hedenbergite from the Kwadonkaya complex (Fig. 3.8), with the compositions lying entirely within the field typical of metaluminous rock-types, as defined by Bonin & Giret (1985a).

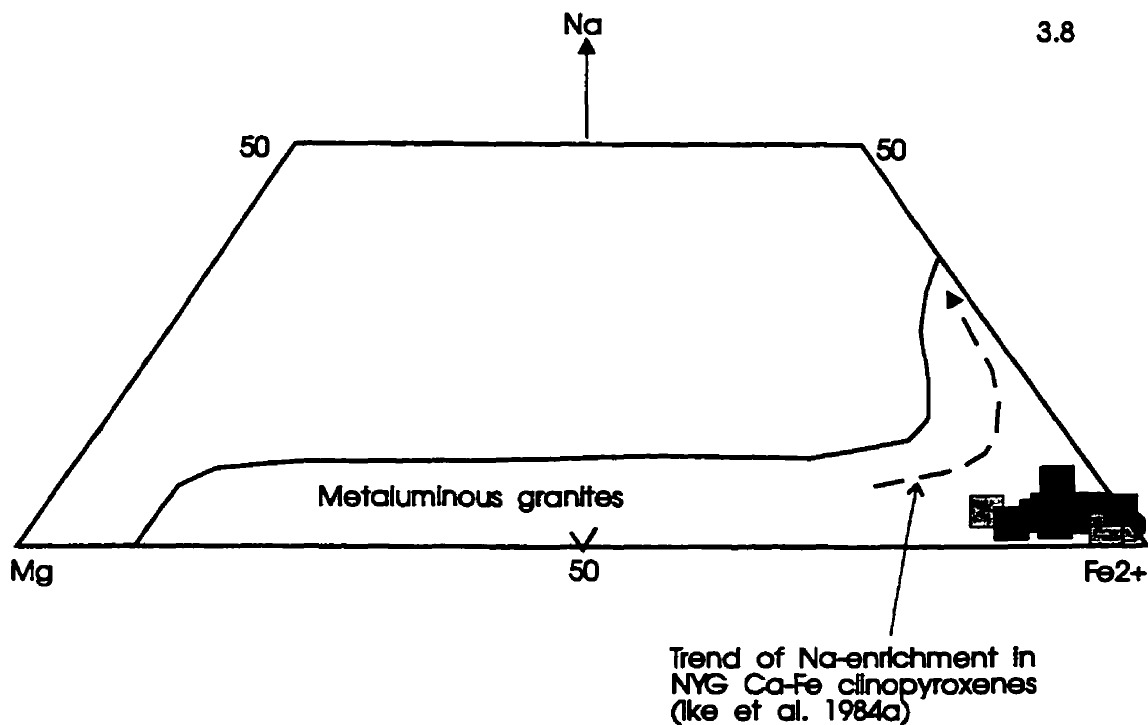
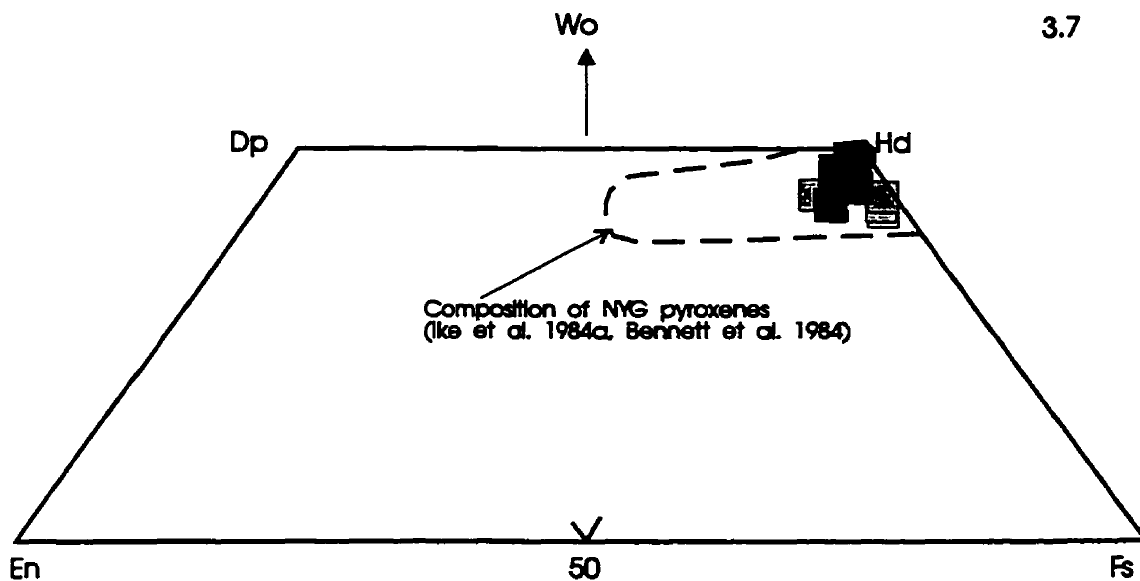
Rhyolite porphyry samples that do not show evidence of hybridization of mafic and felsic magmas contain a pyroxene of average composition: $\text{Fs}_{54.4}\text{Wo}_{44.6}\text{En}_1$, $\text{F/FM} = 0.99$. Variations in zoned grains are subtle: the composition of the core in sample 87 is: $\text{Fs}_{55.7}\text{Wo}_{42.7}\text{En}_{1.6}$, $\text{F/FM} = 0.97$, and that of the rim (87r), $\text{Fs}_{55.7}\text{Wo}_{43.6}\text{En}_{0.8}$, $\text{F/FM} = 0.99$. Zoning in this case is normal. The pyroxene compositions quoted above are more Fe-rich (i.e., more evolved) than in those samples of rhyolite porphyry with disaggregated enclaves, in which the pyroxene has an average composition $\text{Fs}_{48.7}\text{Wo}_{44.4}\text{En}_7$, $\text{F/FM} = 0.88$. The lower F/FM value is close to that of the enclaves ($\text{F/FM} = 0.86$), and suggests that the pyroxene is most likely xenocrystic, originating by disaggregation of mafic enclaves, as the rhyolitic host is significantly more evolved (bulk $\text{F/FM} = 0.94$), but the pyroxene could also have crystallized from contaminated rhyolitic melt.

Samples of hedenbergite fayalite granite porphyry contain unzoned hedenbergite having a composition in the range: $\text{Fs}_{49.1-51.8}\text{Wo}_{47.7-45.9}\text{En}_{3.8-2.3}$, $0.93 \leq \text{F/FM} \leq 0.96$. The least evolved composition ($\text{Fs}_{50.5}\text{Wo}_{43.2}\text{En}_{6.3}$) is found in samples without obvious enclaves; it may be a product of a contaminated magma. The hedenbergite in some samples with obvious enclaves, from the northeastern part of the Buri ring dyke, is unzoned and has an average composition $\text{Fs}_{50.2}\text{Wo}_{46.3}\text{En}_{3.5}$, $\text{F/FM} = 0.95$, a value similar to that in samples without obvious enclaves. Thus, this pyroxene crystallized in equilibrium with the felsic melt. Two patterns of zoning are observed

Fig. 3.7: Plot of pyroxene data in the pyroxene quadrilateral En-Dp-Hd-Fs (En = enstatite, Dp = diopside, Hd = hedenbergite, and Fs = ferrosilite). Dashed line outlines the field of pyroxene compositions for the NYG (after Bennett et al. 1984, Ike et al. 1984). Symbols used in this and other diagrams are: filled squares (pale: rhyolite porphyry, dark: hedenbergite fayalite granite porphyry), filled circles (hastingsite $\pm$ hedenbergite granite porphyry), and filled diamonds (hornblende biotite granite porphyry).

Fig. 3.8: A plot of the lower part of the Mg-Fe-Na diagram for pyroxene compositions. The solid line outlines the composition of clinopyroxene in metaluminous rocks (after Bonin & Giret 1985a, b). Arrow indicates the postmagmatic Na-enrichment trend documented in the Tibchi complex (Ike et al. 1984).

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in some of the samples: a) normal zoning in samples without enclaves: core (43c): $\text{Fs}_{50.9}\text{Wo}_{47.3}\text{En}_{1.8}$, and rim (43r): $\text{Fs}_{49.8}\text{Wo}_{48.6}\text{En}_{1.6}$; and b) reverse zoning in grains relatively close to microsyenodiorite enclaves: core (67c): $\text{Fs}_{51.6}\text{Wo}_{45.4}\text{En}_{3.0}$, $\text{F/FM} = 0.95$; intermediate zone (67i): $\text{Fs}_{50}\text{Wo}_{46.3}\text{En}_{3.7}$, $\text{F/FM} = 0.93$, and rim: with $\text{Fs}_{49.9}\text{Wo}_{46.3}\text{En}_{3.8}$, $\text{F/FM} \approx 0.93$. In addition, a relatively sodium-rich hedenbergite (#43.3, Table 3.3) is more Fe-rich ($\text{Fs}_{53.2}\text{Wo}_{44.3}\text{En}_{2.0}$) than the average composition of hedenbergite in the Buri ring dyke; it shows the highest extent of Fe^{3+} incorporation in M1 (8.5%) recorded in this suite, and also the highest incorporation of Na (9%) in M2, possibly reflecting late-magmatic to subsolidus oxidation.

In the metaluminous hedenbergite hastingsite granite porphyry of the cone sheets, hedenbergite grains are relatively homogeneous, having an average composition $\text{Fs}_{50.9}\text{Wo}_{49.0}\text{En}_{0.1}$, and $\text{F/FM} = 1.0$. Here the F/FM of hedenbergite is the same as that of the host rock, consistent with lack of xenocrystic material.

In summary, there are significant overlaps among pyroxene compositions in the Buri ring dyke: rhyolite porphyry ($\text{Fs}_{48.7-55.4}\text{Wo}_{42.9-44.9}\text{En}_{0.78-7.0}$, $0.88 \leq \text{F/FM} \leq 0.99$), and hedenbergite fayalite granite porphyry ($\text{Fs}_{49-51.8}\text{Wo}_{43.2-48.6}\text{En}_{1.7-6.3}$, $0.89 \leq \text{F/FM} \leq 0.97$). The same magma probably is responsible for the two rock types. Normal zoning documents *in situ* fractional crystallization, whereas reverse zoning reflects contamination by mixing of felsic and trachyandesitic melts.

3.4.4 Amphibole

Analytical results for amphiboles are given in Table 3.4. The site occupancies are calculated after the scheme proposed by Leake (1978), Robinson et al. (1982), and Hawthorne (1983). In this discussion, I will refer to the sites M4 as B, and M1, M2, M3 as C. The amphiboles in the subvolcanic rocks of Kwandonkaya Complex are of the calcic variety, implying that in all cases $(\text{Ca} + \text{Na})_{\text{B}}$ exceeds 1.34 apfu (atoms per formula unit), and Na_{B} is less than 0.67.

The order of discussion of amphibole compositions follows the relative age of geological units.

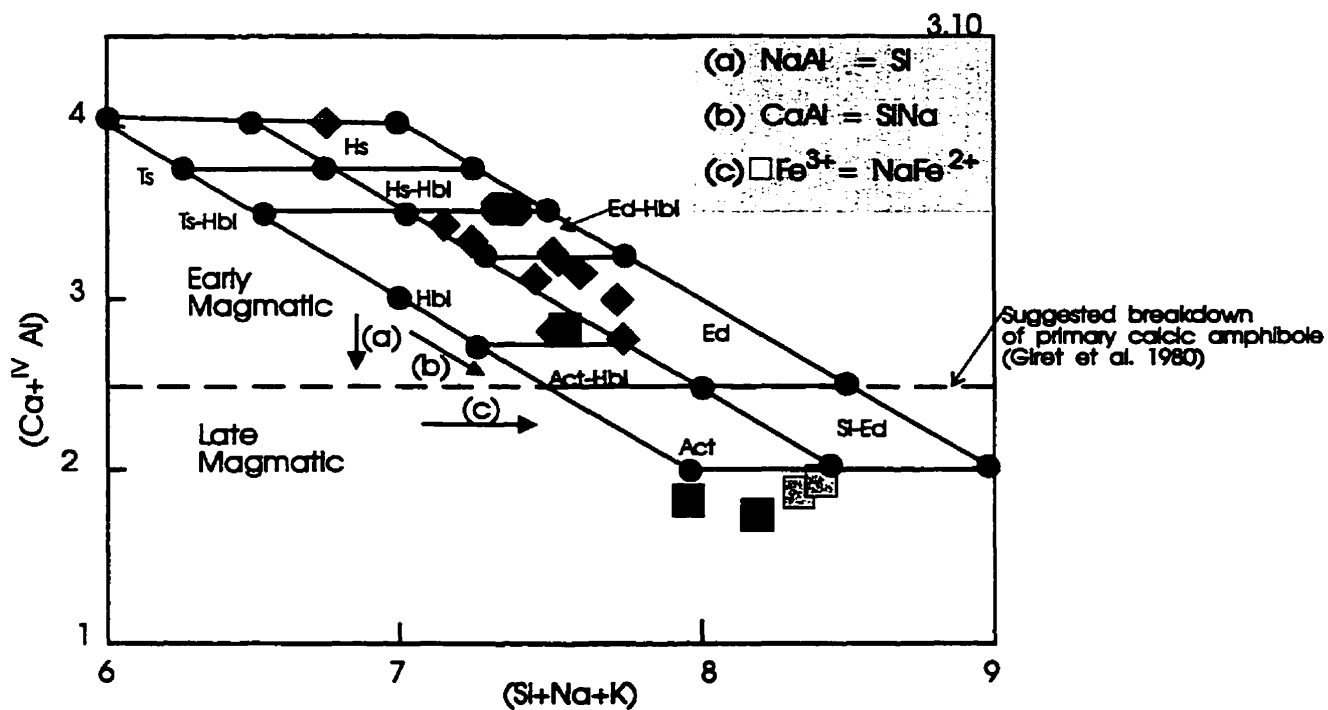
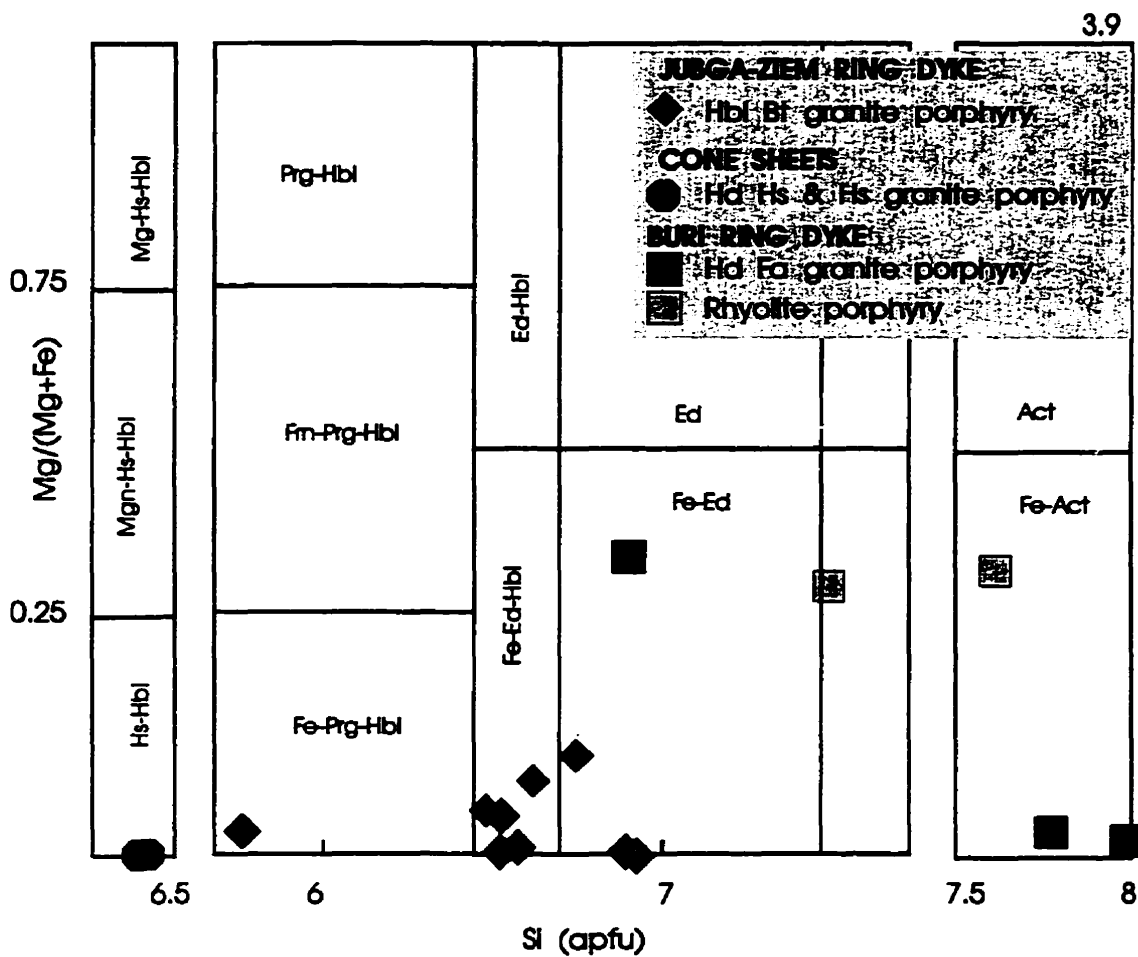
Some units in the Buri ring dyke characterized by obvious signs of disaggregation of trachyandesitic material contain relatively magnesian amphibole. For example, samples of metaluminous rhyolite porphyry with such enclaves contains ferroedenite with $F/FM = 0.74$ and ferroactinolite with $F/FM = 0.65$ (Fig. 3.9). The ferroedenite has the lowest proportion of Al, and the highest proportion of Ti. The F/FM of the ferroedenite is much lower than that of the felsic host rock (bulk $F/FM = 0.97$), and also clearly more mafic than that of the mafic enclaves (bulk $F/FM = 0.86$). Thus, ferroedenite in this case can be considered a xenocryst, formed from the disaggregation of mafic enclaves, themselves hybrid judging from petrographic and mineral chemical features studied further in Chapter 7. The ferroactinolite was most likely produced by the subsolidus replacement of clinopyroxene xenocrysts derived from the enclaves. Along with their more magnesian composition, the amphiboles in the rhyolite porphyry have a relatively high fluorine content [ferroedenite ($X_F: 0.37$, $X_{Cl}: 0.03$), ferroactinolite ($X_F: 0.25$, $X_{Cl}: 0.02$)].

Within the metaluminous hedenbergite fayalite granite porphyry (Buri ring dyke), some samples contain ferroedenite with a lower F/FM , 0.63, suggestive of an origin by replacement of xenocrysts of clinopyroxene. In other samples, also bearing enclaves, the ferroactinolite ($F/FM = 0.95$) is similar to that of the hedenbergite in this rock type, and more comparable with that of the felsic host. From petrographic studies, amphiboles in this rock type are considered secondary; one common characteristic of these amphiboles is their relatively low halogen content [ferroedenite ($X_F: 0.1$, $X_{Cl}: 0.03$), ferroactinolite ($X_F: 0.02$, $X_{Cl}: 0.01$)]. The ferroactinolite in this case was produced by metasomatic replacement of pyroxene from the felsic host, and is similar in composition and origin to that in the Mahé granitic complex, Seychelles (Mitchell 1990), where this mode of origin is considered well established.

In the cone sheets, the amphibole in the metaluminous hedenbergite-bearing granite

Fig. 3.9: Plots of $\text{Mg}/(\text{Mg}+\text{Fe})$ vs. Si (apfu) for amphiboles. The classification is after Leake (1978).

Fig 3.10: Plot of $(\text{Ca} + {}^{\text{IV}}\text{Al})$ vs. $(\text{Si} + \text{Na} + \text{K})$ for amphiboles. The fields for "early magmatic" and "late-magmatic" amphibole (after Giret et al. 1980) are indicated, as well as major substitution trends.



porphyry is hastingsitic hornblende (Fig. 3.9), with $F/FM = 1.0$. The amphibole in this rock type is the only one, out of all the ones analyzed from the Kwadonkaya complex, without vacancies ($\square$) in the A site, and also has the highest proportion of halogens in the OH site ($X_F = 0.13$, $X_{Cl} = 0.32$). Mildly peraluminous hastingsite granite porphyry also contains a hastingsitic hornblende (Fig. 3.9), with a F/FM of 0.99, and is as aluminous as in the hedenbergite hastingsite granite porphyry. The halogen content is again high ($X_F: 0.07$, $X_{Cl}: 0.36$), with the mol fraction of chlorine being the highest of the suite. Also, these are the only amphiboles having X_{Cl} distinctly greater than X_F . The amphibole in these cone sheets is considered primary on the basis of petrographic observations.

In the Jubga-Ziem ring dyke, mildly peraluminous "hornblende"-biotite granite porphyry contains ferropargasitic hornblende ($F/FM = 0.94$), ferroedenitic hornblende ($0.9 \leq F/FM \leq 1.0$) or ferroedenite ($F/FM = 1.0$) (Fig. 3.9). Their halogen contents vary [ferropargasitic hornblende ($X_F = 0.16$, $X_{Cl} = 0.12$), ferroedenitic hornblende (range in X_F 0.03-0.23, average = 0.13; range in X_{Cl} : 0.04-0.12, average = 0.09), and ferroedenite ($X_F = 0.12$, $X_{Cl} = 0.06$)]. The increase in F/Cl ratio with F/FM , in spite of Fe-F avoidance, indicates amphibole crystallization in progressively evolved and relatively F-enriched residual melt. Drusy marginal parts of the "hornblende"-biotite granite porphyry in the southern part of Jubga-Ziem center have amphibole compositions that are more evolved (ferroedenite to ferroedenitic hornblende) than samples further in from the contact (ferropargasitic hornblende). In this example, the X_F of amphibole decreases inward from the outer contact, whereas X_{Cl} increases. It is likely that loss of Cl occurred due to degassing near the contact.

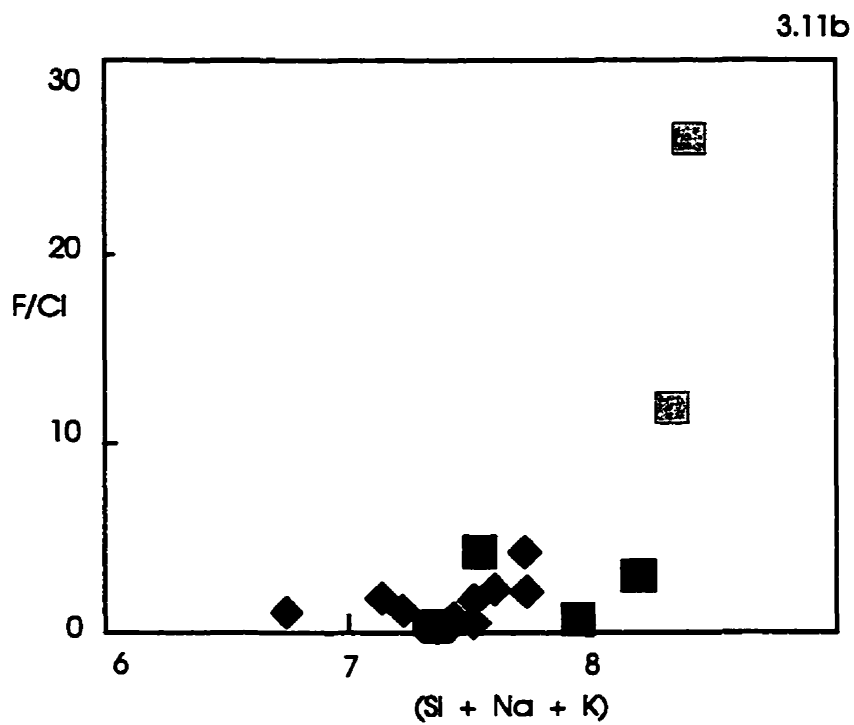
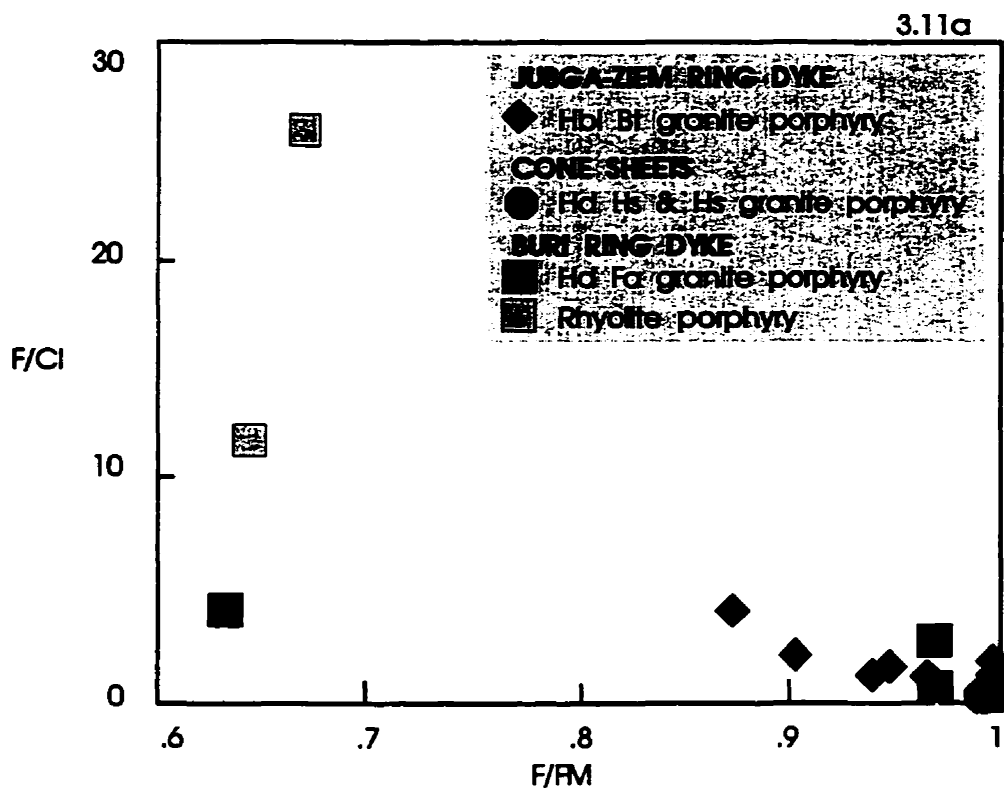
In a plot of $(Ca+^{IV}Al)$ vs $(Si+Na+K)$, the hastingsitic hornblende, ferroedenite, ferropargasitic hornblende, and ferroedenitic hornblende compositions described above all plot in the field of "early magmatic" amphiboles, above the line at $(Ca+^{IV}Al) = 2.5$, proposed by Giret

et al. (1980) as the boundary between magmatic and postmagmatic calcic amphiboles (Fig. 3.10). The ferroactinolite falls in their "late-magmatic" field. From petrographic studies, amphiboles plotting in the "early magmatic field" of Giret et al. (1980) are actually late-magmatic to subsolidus, whereas those plotting in the "late-magmatic field" are secondary. Thus, there is no correlation in this suite between the above distinctions based on Figure 3.10 and the sequence of crystallization. Overall, the dominant scheme of substitution in the late-magmatic to subsolidus amphiboles is edenitic $[(\text{Na}+\text{K})_{\text{A}} + \text{VI Al} \rightleftharpoons \square_{\text{A}} + \text{Si}]$, whereas in the secondary amphiboles, $\text{Ca} + \text{VI Al} \rightleftharpoons \text{Na} + \text{Si}$, and $\square + \text{Fe}^{3+} \rightleftharpoons \text{Na} + \text{Fe}^{2+}$ substitutions are important (Fig. 3.10). Plots of $X_{\text{F}}/X_{\text{Cl}}$ against $(\text{Si}+\text{Na}+\text{K})$ and F/FM (Figs. 3.11a, b) yield overlapping trends, and thus indicate that the respective compositions most likely crystallized from comagmatic batches of magma possibly related by fractionation from a single parent. However, variable extent of Cl loss through degassing and postmagmatic modification of the halogen content also could explain the nonlinear trend. With the exception of the hastingsitic hornblende, which has X_{Cl} greater than X_{F} , all other amphiboles from Kwadonkaya complex have X_{F} greater or similar to X_{Cl} . The ferroactinolite (subsolidus) has relatively low total halogen contents.

3.5 Discussion

To apply published experimental data on the stability of olivine, pyroxene, and amphibole analyzed, the structural model proposed earlier (Chapter 2), and those of Bonin (1986) and Kinnaird (1987) offer a good frame of reference. In a comparative study of ring complexes, Bonin (1986) proposed that felsic magma rises from deeper chambers in the crust, at 7-32 km, with maximum clustering at about 15 km, to a subvolcanic reservoir at a depth of 1-4 km. In this model, a volcano-caldera complex forms the highest expression of the plume of magma on the surface. In a study of fluid inclusions in quartz phenocrysts within subvolcanic rocks of the Saiya

Fig. 3.11: Plots of a) F/Cl vs. F/FM [Fe/(Fe+Mg)], and b) F/Cl vs. (Si + Na + K) for amphiboles.



Shokobo Complex, Kinnaird (1987) found evidence for crystallization of β quartz at a depth of about 6 km, at about 750°C. The phenocrysts (including fayalite, hedenbergite, and sanidine_{ss}) were entrained in the melt and carried from a deeper-level magma chamber, to the subvolcanic level. Crystal-rich melts with short residence-times in the upper chamber were extruded as crystal-rich ignimbrites and other volcanic units (now removed by erosion in this complex but still present in other NYG complexes) or emplaced in cone sheets and ring dykes. Experimental data collected over a pressure range of 0.5 to 3 kbar thus seem appropriate for an appraisal of conditions of formation of the minerals. This range is also in agreement with conditions of crystallization deduced from the plot of normative albite-orthoclase-quartz (Fig. 3.12a).

3.5.1 Magmatic evolution

Significant overlaps in the DI (differentiation index) of subvolcanic rocks have previously been documented (Chapter 2): the Buri ring dyke ($92.2 \leq DI \leq 94.4$), cone sheets ($91.2 \leq DI \leq 94.9$), and the Jubga-Ziem ring dyke ($93.1 \leq DI \leq 95.1$). This led to the preliminary conclusion that rock units in the Kwandonkaya complex crystallized from comagmatic batches of melt. A triangular plot of normative albite-orthoclase-quartz backed with textural evidence (Fig. 3.12a), suggests that the subvolcanic rocks crystallized along water-undersaturated paths as defined by Holtz et al. (1989, 1991a, b) and Ebadi & Johannes (1991), as also discussed in Chapter 2. Furthermore, the late-magmatic appearance of amphibole (Figs. 3.10) supports the hypothesis that the original magma was relatively reduced and H₂O-undersaturated (Wones & Gilbert 1982).

In most cases, the **fayalite** is significantly altered, presumably owing to subsolidus increase in $f(\text{O}_2)$, and could not be analyzed. Therefore, the contrasts in composition between the rhyolite porphyry and fayalite syenogranite may not be meaningful. However, they do suggest that the rhyolite porphyry is the more primitive. Also, variations in Fa and Tp contents of fayalite in

Fig. 3.12: a) A triangular plot of normative albite-orthoclase-quartz for the subvolcanic rocks of Kwandonkaya complex. The inset (enlarged in Appendix II.2) indicates various minima at different pressures and volatile contents. Symbols are the same as in Fig. 3.9.

b) P-T diagram displaying the stability of ferromagnesian phases and the granite solidus at different water activities. Stability curves: a: ferroactinolite (at QFM), b: α - $\rightleftharpoons$ β -quartz transition, c: ferropargasite (QFM), d: hastingsite (QFM), e: fayalite + sanidine $\rightleftharpoons$ annite + quartz, f: fayalite + magnetite + quartz + fluids $\rightleftharpoons$ grunerite, g: ferropargasite (WM), and h: hastingsite (WM). Here, where QFM refers to the fayalite-magnetite-quartz buffer, and WM refers to the wüstite-magnetite buffer. Minimum compositions at different water activities from Ebadi & Johannes (1991), whereas stability data for ferromagnesian minerals are taken from various sources.

these rocks (Mn decreases as Fe increases) suggest that the batches of magma that gave these two units were not related by fractionation from a single parental composition. Mn should increase as F/FM increases with fractionation (Brown 1982), which is not the case here. Furthermore, the coexistence of fayalite and ilmenite in these rocks indicates that crystallization probably was initiated below the QFM buffer.

The hedenbergite in the subvolcanic rocks, like fayalite, occurs as a phenocryst phase. There is no significant build-up in Na (Fig. 3.8), and most compositions lie at the Fe-rich end of the compositional field of clinopyroxene found in metaluminous silica-saturated rocks (Bonin & Giret 1985a). The postmagmatic initiation of Na-enrichment noted in the hedenbergite in the Tibchi complex (Ike et al. 1984) is absent, perhaps because of the greater extent of degassing here. Overlaps documented in pyroxene compositions of the various units (Figs. 3.7, 8) suggest a comagmatic suite and further stresses the suggestion (made on the basis of fayalite compositions) that the subvolcanic rocks in the Buri ring dyke and Jubga-Ziem center formed from different batches of magma. The normal zoning in some hedenbergite grains, from both rhyolite porphyry and hedenbergite granite porphyry, documents fractional crystallization without influx of mafic melt. Reverse zoning documented in enclave-bearing hedenbergite granite porphyry reflects the effect of local contamination by relatively more mafic material (trachyandesitic enclaves) dispersed in the rock. The effects of hybridization are further examined in a later chapter.

The effect of addition of the orthoclase component on the phase relations in the plane $\text{Mg}_2\text{SiO}_4\text{-Fe}_2\text{SiO}_4\text{-CaMgSi}_2\text{O}_6\text{-CaFeSi}_2\text{O}_6$, in both the Mg-free and the Fe-free bounding joins, is to suppress the liquidus field of silica and to expand that of the mafic silicates (Hoover & Irvine 1978). This observation explains the early crystallization of fayalite and hedenbergite in subvolcanic rocks of Kwandonkaya complex. The coexistence of quartz and fayalite instead of

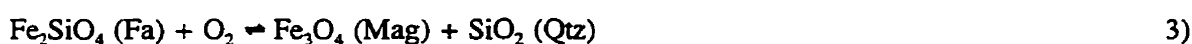
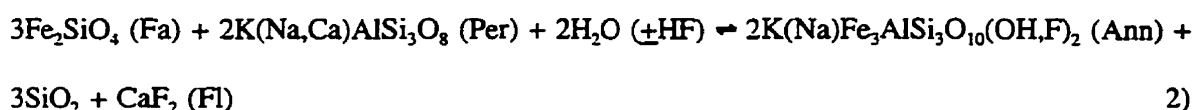
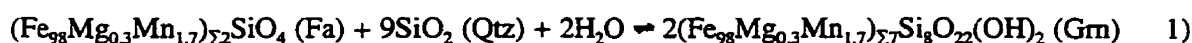
ferrosilite is expected at the inferred low pressure of crystallization (cf. Ferguson 1978, Lindsley 1980, Robinson 1980).

The **amphibole** compositions cover a range of F/FM values (Figs. 3.9, 3.11a, b). The noncorrelation of the F/FM (0.63 to 1.00) of the amphibole with the degree of evolution of their host rocks (i.e., DI) is comparable to the case in the Baie-des-Moutons complex, Quebec (Lalonde & Martin 1983), whose amphiboles (ferroedenitic hornblende to ferroedenite) also have a limited range of F/FM (0.61-0.83). Increases in $f(\text{O}_2)$ usually leads to the co-precipitation of Fe-Ti oxides and accompanying decrease in the Fe/Mg ratio of ferromagnesian minerals that crystallize at late-magmatic stages. In assemblages consisting only of ferromagnesian silicate(s) plus liquid, Fe/Mg in hornblende decreases with temperature and Fe/Mg of the liquid, at constant redox conditions (Mitchell & Platt 1978, Lindsley 1980, Stephenson & Upton 1982, Schneiderman 1991, Rowins et al. 1991). The trend is usually reversed (i.e., Fe/Mg increases) during cooling by solid-solid reactions involving Fe- and Ti-rich accessory phases (Gilbert et al. 1982), e.g., ilmenite, titanite, magnetite, fayalite, and allanite. The hastingsitic hornblende has X_{Cl} greater than X_{F} . This may explain the observation that both hedenbergite hastingsite- and hastingsite-granite porphyry have the most prominent occurrence of chevkinite and allanite, of all the rocks in the Kwadonkaya complex. All other amphibole compositions have more fluorine than chlorine, and invariably have allanite associated with them, although they cannot be described as mineralized.

The presence of primary amphibole in some subvolcanic rock units reflects the fact that the melt composition shifted from the feldspar compositional plane into corundum-normative "space", as the hornblende must be "diopside-normative" (Cawthorn 1976, Helz 1979, Gilbert et al. 1982). Thus the appearance of amphibole in the subvolcanic rocks was controlled by the aluminum saturation index, and volatile enrichment in the residual melt.

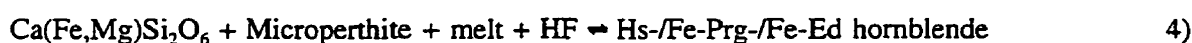
3.5.2 Subsolidus re-equilibration

In the fayalite syenogranite, fayalite is replaced by quartz, magnetite, and grunerite ($\pm$ iddingsite, chlorite); upon further cooling, the grunerite reacted with perthite to form annite. Such a subsolidus assemblage implies that the fayalite syenogranite re-equilibrated primarily at a subsolidus temperature above the stability field of annite + quartz (cf. Gilbert et al. 1982), and below the alkali feldspar solvus (ca. $< 500^\circ\text{C}$). Some of the possible reactions during the late-magmatic to subsolidus alteration of fayalite include:



The alteration of fayalite to magnetite + quartz, grunerite, or iddingsite is consistent with increasing $f(\text{O}_2)$ and loss of H_2 and H_2O (Janeczek 1989).

Grains of pyroxene are replaced mainly by amphiboles, discussed below, or annitic mica, chlorite, $\pm$ magnetite. Schematic reactions that could have occurred include:



Changes associated with the breakdown of primary ferroedenitic amphibole involve decreases in alkalis and $^{\text{VI}}\text{Al}$ (at late-magmatic stages) due to the substitution: $(\text{Na}+\text{K})_{\text{A}} + ^{\text{VI}}\text{Al} \rightleftharpoons \square_{\text{A}} + \text{Si}$ (Fig. 3.10). Other substitutions that account for the observed changes include: $^{\text{VI}}\text{Ca} + \text{Al} \rightleftharpoons \text{Na} + \text{Si}$, and $\square + \text{Fe}^{3+} \rightleftharpoons \text{Na} + \text{Fe}^{2+}$ (at subsolidus stages). Mitchell (1990) characterized the latter as the "ferroactinolytic subtrend", which is also known as the "oxidation trend". It differs from the "peralkaline" trend, which produces ferrichterite, arfvedsonite or riebeckite, at a late- or post-magmatic stage in the Tibchi Complex in Nigeria (Ike et al. 1985), the Tarraouadji Complex in

Niger (Fabriès & Rocci 1965, Perez 1985), and the Ras ed Dom Complex in Sudan (O'Halloran 1985). Mitchell (1990) explained the "ferroactinolite subtrend" as due to trapping of "unevolved" (i.e., nonperalkaline) fluids that reacted and altered the earlier ferroedenite. The limited extent of incorporation of Na in amphiboles in the subvolcanic rocks indicates that the ferroedenite in rhyolite porphyry and hedenbergite porphyry probably were produced by late-magmatic reaction of hedenbergite with the residual melt, whereas ferroactinolite was produced by reaction of either ferroedenite or hedenbergite with an "unevolved" residual fluid.

3.5.3 Inferred P-T conditions of formation

There is very little experimental work on bulk compositions that are as extreme in their degree of Fe-enrichment as those encountered in this study. The evolution of ferromagnesian minerals toward Fe-rich compositions requires low $f(\text{O}_2)$, since oxidizing conditions will precipitate ilmenite and magnetite, with consequent depletion of Fe^{2+} from the melt so as to inhibit crystallization of Fe-rich mafic minerals.

Experimental studies at one kilobar and above the WM buffer indicate that fayalite is stable down to about 800°C, but becomes unstable where $f(\text{O}_2)$ approaches the QFM buffer (Barker et al. 1975). From these results, one can infer that the crystallization of fayalite-bearing rocks was initiated above 800°C.

In the system $\text{Mg}_2\text{SiO}_4\text{-Fe}_2\text{SiO}_4\text{-CaMgSi}_2\text{O}_6\text{-CaFeSi}_2\text{O}_6\text{-KAlSi}_3\text{O}_8\text{-SiO}_2$, at less than 1 kbar pressure, $0.8 \leq \text{F/FM} \leq 1.0$, and a temperature below 1050°C, the stable assemblage is fayalite + clinopyroxene + tridymite + K-feldspar for bulk compositions with less than 5% ferromagnesian minerals (Huebner 1980). This assemblage is similar to that found in rhyolite porphyry and hedenbergite fayalite granite porphyry in the Buri ring dyke. The occurrence of hedenbergite-rich pyroxene and ilmenite in some of the subvolcanic rocks indicates that $f(\text{O}_2)$ was close to the QFM

buffer, whereas the assemblage magnetite + quartz + titanite associated with hedenbergite, or amphibole (or products of their subsolidus replacement) indicates relatively high $f(\text{O}_2)$ (Wones 1989).

The initially water-undersaturated conditions precluded the crystallization of amphiboles at an early magmatic stage, since their stability curves at QFM oxygen buffer all fall in the subsolidus range for such water-undersaturated granite solidi [$0.3 \leq a(\text{H}_2\text{O}) \leq 0.5$] (Fig. 3.12b). Generally, lower $f(\text{O}_2)$ (e.g., along the WM buffer shown in Fig. 3.12b), and solid solutions expand the stability field of the amphibole to intersect the granite solidi at higher temperatures, such that the amphibole becomes primary. Also, crystallization-induced buildup of volatiles (fluorine + water) could have led to the stabilization of amphibole at P-T conditions in the presence of melt. In any case, the amphibole would be late relative to the bulk crystallization of the magma similar to the petrographic observation in the Kwandonkaya subvolcanic rocks. At $P_{\text{fluids}} = 3$ kbar, and along the QFM buffer, hastingsite, of comparable degree of Fe enrichment, is known to decompose at 710°C to an assemblage of hedenbergite + plagioclase (+alkali feldspar) + fayalite + quartz + fluid (the magmatic assemblage in rhyolite porphyry and hedenbergite fayalite granite porphyry), and the stability curve intersects the granite solidus at 2.3 kbar, at about 680°C (Thomas 1982). The stability curve for ferropargasite (at QFM) is subparallel to that of hastingsite (Gilbert 1966), as shown in Figure 3.12b, indicating an increase in stability limit with Al-for-Fe³⁺ substitution. The decrease in both A-site occupancy and ^{IV}Al content from the hastingsitic hornblende via ferropargasitic hornblende, ferroedenitic hornblende, and ferroedenite to ferroactinolite should, according to Gilbert et al. (1982), correlate with decreasing temperature at constant P, $a_{\text{H}_2\text{O}}$, $f(\text{O}_2)$, and bulk composition. The occurrence of annitic mica along with primary amphibole suggests that the stability field of magmatic amphibole overlapped with that of annite + quartz (Fig. 3.12b).

The stability of ferroactinolite, as shown in Fig. 3.12b, does not intersect the water-saturated granite solidus at QFM conditions of the buffer. This is consistent with its predominantly subsolidus and secondary origin. The inferred difference between grunerite stability and that of annite + quartz is less than 20°C, and hence explains their co-occurrence as a replacement of fayalite.

In summary, the crystallization of fayalite and hedenbergite was initiated above or close to the alkali feldspar solvus, since the appearance of orthoclase microperthite (sanidine_{ss}) apparently terminated their growth. Such crystallization was also above the temperature of crystallization of β -quartz (i.e. $T > 750^\circ\text{C}$). However, because evidence exists for continued crystallization of fayalite and hedenbergite until the end stages of alkali feldspar crystallization, the final temperature of crystallization was close to the alkali feldspar solvus. Amphibole crystallization postdates alkali feldspar, and occurs in the stability field of hastingsite and pargasite, but above that of ferroactinolite, possibly in the range of 600-700°C, close to QFM buffer.

3.6 Summary and conclusion

1) The presence of Fe²⁺-rich ferromagnesian phases at the initial and final stages of crystallization indicates relatively reduced melts throughout.

2) The crystallization of fayalite and hedenbergite was initiated above or close to the alkali feldspar solvus, since the appearance of orthoclase microperthite (sanidine_{ss}) apparently terminated their growth, whereas amphibole crystallization invariably spanned the late-magmatic to subsolidus range (i.e., postdates alkali feldspar crystallization).

3) The pyroxene and amphibole data indicate relatively low Na-content, as is typical of metaluminous rocks. The limited incorporation of Na in hedenbergite and calcic amphibole

indicates lack of substantial alkali enrichment in melts at both early and final stages of crystallization of the subvolcanic rocks; Kwandonkaya thus does not show the late- to post-magmatic peralkaline trend documented in some metaluminous-peraluminous complexes.

4) Variations of Fe/Mg of fayalite, hedenbergite, and amphibole are not sequential with respect to the relative age of emplacement and degree of evolution of their host rocks. Factors other than fractionation, namely, different extents of volatile build-up and loss, and different degrees of subsolidus equilibration, thus also influenced the composition of the ferromagnesian minerals in the subvolcanic rocks of the complex.

5) In the subvolcanic rocks of Kwandonkaya complex, the fayalite, hedenbergite, and amphibole generally have F/FM greater than that of the bulk (whole) rock and are distinguished from the case of some mixed rocks of the Buri ring dyke, which have the F/FM values of the mafic phases influenced by that of the mafic enclaves. The noncorrelation of F/FM of a mineral and that of the host rock, along with reverse zoning, provide evidence of interaction of the metaluminous magma with more mafic material.

6) Subsolidus assemblages in place of the magmatic ferromagnesian phases consistently indicate a shift of $f(\text{O}_2)$ conditions to above QFM, and an increase in $a(\text{H}_2\text{O})$, and F/Cl with decreasing temperature.

Table 3.2: Average composition of fayalite: Rhyolite porphyry:- 49B; and fayalite syenogranite:- 57.

	Buri	Jubga-
	49B	Ziem
	57	
n	4	7
SiO ₂	30.64	28.98
TiO ₂	0.02	0.02
Al ₂ O ₃	0.01	0.01
Cr ₂ O ₃	0.01	-
Fe ₂ O ₃	-	0.63
FeO	66.07	67.92
MnO	1.79	1.25
NiO	0.00	-
MgO	1.26	0.12
CaO	0.33	0.08
Na ₂ O	0.01	0.02
K ₂ O	0.00	0.00
Total	100.15	99.03

Formula on basis of 4 oxygens

Si ⁴⁺	1.02	0.99
Ti ⁴⁺	0.00	0.00
Al ³⁺	0.00	0.00
Cr ³⁺	0.00	-
Fe ³⁺	-	0.02
Fe ²⁺	1.85	1.94
Mn ²⁺	0.05	0.04
Ni ²⁺	0.00	-
Mg ²⁺	0.06	0.01
Ca ²⁺	0.01	0.00
Na ⁺	0.00	0.00
K ⁺	0.00	0.00
Fo	3.21	0.31
Fa	94.20	97.88
Te	2.59	1.81
F/FM	0.97	0.99

Fe³⁺ is calculated after (Droop et al. 1987) for fayalite (Table 3.2), and clinopyroxene data (following Table 3.3): $F = 2X(1-T/S)$, where F = Fe³⁺ ions per X atoms of oxygen in a mineral formula, T = ideal number of cations per formula unit, and S = observed cation total per X oxygens calculated assuming all iron to be Fe²⁺.

Table 3.3: Average compositions of clinopyroxene: Hedenbergite hastingsite granite porphyry (cone sheet):- 64: hedenbergite partly replaced by hastingsitic hornblende; **rhyolite porphyry:-** 87.1 (unzoned grain), 87.2c, 87.2i, 87.2r: core, intermediate zone and rim of zoned hedenbergite, 49B: hedenbergite microphenocrysts; **hedenbergite fayalite granite porphyry:-** 43.1: hedenbergite phenocrysts, 43.2: relatively Mg-rich hedenbergite, 43.3: relatively sodic hedenbergite, 43.4c and 43.4r: core and rim of zoned hedenbergite grain, 67.1: hedenbergite phenocrysts, 67.2c, 67.2i, 67.3r: core, intermediate zone, and rim of hedenbergite.

Buri ring dyke

Cone
sheet
64

	87.1	87.2c	87.2i	87.2r	49B	43.1	43.2	43.3	43.4c	43.4r	67.1	67.2c	67.2i	67.2r	
n	2	2	1	2	2	15	2	2	1	1	2	2	1	2	5
SiO ₂	46.97	47.61	47.69	47.58	49.07	49.05	49.30	48.19	47.67	47.25	47.94	48.41	48.17	48.36	48.25
TiO ₂	0.46	0.33	0.38	0.22	0.13	0.11	0.38	0.20	0.14	0.11	0.08	0.10	0.16	0.09	0.07
Al ₂ O ₃	0.57	0.44	0.39	0.35	0.38	0.19	0.50	0.19	0.18	0.23	0.12	0.17	0.26	0.15	0.53
Cr ₂ O ₃	-	-	-	-	-	0.01	0.02	-	-	-	-	-	-	-	0.00
Fe ₂ O ₃	1.03	0.84	1.65	1.19	0.44	-	3.27	2.79	3.14	0.84	1.45	1.78	1.10	0.96	1.15
FeO	29.48	29.58	30.49	29.56	26.81	28.28	28.03	26.67	27.45	26.69	27.52	26.85	26.99	27.06	27.54
MnO	0.81	0.66	0.79	0.65	0.69	0.64	0.72	0.69	0.72	0.63	0.85	0.77	0.66	0.74	0.83
NiO	-	-	-	-	0.04	0.02	0.02	-	-	-	-	-	-	-	0.03
MgO	0.26	0.24	0.52	0.25	2.16	0.71	1.96	0.74	0.55	0.50	0.66	1.18	1.31	1.18	0.05
CaO	18.91	19.08	18.40	18.93	19.08	19.55	18.73	18.83	19.92	20.33	20.42	20.07	19.79	20.33	21.33
Na ₂ O	0.34	0.45	0.34	0.46	0.62	0.52	0.41	1.08	0.52	0.50	0.28	0.54	0.52	0.41	0.41
K ₂ O	0.00	0.00	-	0.01	0.02	0.01	0.05	0.00	0.01	0.01	0.05	0.00	-	0.01	0.01
Total	98.95	99.43	99.85	99.66	100.20	99.53	100.10	99.86	99.96	99.39	98.77	99.55	99.64	99.44	100.03
FeO(t)	30.51	30.51	31.25	31.05	27.89	28.68	28.03	29.60	29.96	29.51	28.27	28.16	28.59	28.05	28.41
Structural formula on the basis of 6 oxygens															
IVSi	1.97	1.99	1.98	1.99	2.00	2.02	2.00	2.00	1.98	1.98	2.00	2.00	1.99	2.00	1.99
IVAl	0.03	0.02	0.02	0.01	0.00	0.00	-	0.01	0.01	0.01	0.00	0.00	0.01	0.00	0.01
T site	2.00	2.00	2.00	2.00	2.00	2.02	2.00	2.00	1.99	1.99	2.00	2.00	2.00	2.00	2.00
VIAI	0.00	0.01	0.00	0.00	0.02	0.01	0.02	0.00	-	-	0.00	0.01	0.00	0.01	0.02
Ti	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr	-	-	-	-	-	0.00	0.00	-	-	-	-	-	-	-	0.00
Fe ³⁺	0.04	0.03	0.03	0.05	0.04	0.01	-	0.10	0.09	0.10	0.03	0.04	0.06	0.03	0.03
Fe ²⁺	1.03	1.03	1.06	1.03	0.91	0.97	0.95	0.92	0.95	0.93	0.96	0.92	0.93	0.93	0.95
Mn ²⁺	0.03	0.02	0.03	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.03	0.03	0.02	0.03	0.03
Ni	-	-	-	-	0.00	0.00	0.00	-	-	-	-	-	-	-	0.00
Mg	0.02	0.01	0.03	0.02	0.13	0.04	0.12	0.05	0.03	0.03	0.04	0.07	0.08	0.07	0.00
Ca	0.85	0.85	0.82	0.84	0.83	0.86	0.82	0.83	0.88	0.90	0.91	0.89	0.87	0.90	0.94
Na	0.03	0.04	0.03	0.04	0.05	0.04	0.03	0.09	0.04	0.04	0.02	0.04	0.04	0.03	0.03
K	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
M1,M2	2.01	2.00	2.00	2.01	2.01	1.97	1.98	2.01	2.02	2.03	2.00	2.00	2.01	2.01	2.00
Fs	54.41	54.32	55.44	54.47	48.66	51.80	50.49	51.16	50.87	49.76	50.16	49.11	49.36	49.01	50.11
Wo	44.72	44.90	42.87	44.70	44.35	45.88	43.23	46.32	47.30	48.58	47.69	47.04	46.37	47.16	49.72
En	0.87	0.78	1.69	0.83	6.98	2.32	6.28	2.51	1.83	1.65	2.15	3.86	4.27	3.82	0.17
F/FM	0.99	0.99	0.97	0.99	0.88	0.96	0.89	0.96	0.97	0.97	0.96	0.93	0.92	0.93	1.00

Table 3.4: Average composition of amphibole. **Rhyolite porphyry:-** 49A.1: ferroactinolite, and 49A.2: silicic ferroedenite; **hedenbergite granite porphyry (Buri ring dyke):-** 41: ferroedenite replacing clinopyroxene, 67: ferroactinolite replacing clinopyroxene; **hedenbergite hastingite granite porphyry (cone sheet):-** 64: hastingitic hornblende replacing hedenbergite; **Hastingsite granite porphyry (cone sheet):-** hastingitic hornblende; **"hornblende"-biotite granite porphyry:-** 51: ferroedenite, 55: ferroedenitic hornblende, 56.1: ferroedenitic hornblende, 56.2: ferropargasitic hornblende, 68A: ferroedenitic hornblende, 68B.1: ferropargasitic hornblende, 68B.2: ferroedenitic hornblende, 68B.3: ferroedenite, 79.1: ferroedenitic hornblende, and 79.2: ferroedenite.

	Buri ring dyke				Cone sheets		Jubga-Ziem ring dyke									
	49A.1	49A.2	41	67	64	40A	51	55	56.1	56.2	68A	68B.1	68B.2	68B.3	79.1	79.2
n	2	1	3	3	7	10	6	6	10	1	3	1	1	2	10	1
SiO ₂	49.59	48.72	45.96	47.46	37.37	37.62	41.45	39.39	38.63	34.14	39.32	38.14	39.48	41.34	39.97	40.81
TiO ₂	0.37	0.39	0.59	0.13	0.73	0.46	0.96	1.20	1.62	2.19	1.83	0.27	0.28	0.78	1.62	1.61
Al ₂ O ₃	3.38	3.44	7.85	1.45	8.57	8.72	5.67	8.05	7.69	11.49	7.34	10.32	9.22	6.49	7.49	7.15
FeO	24.23	25.48	23.71	34.01	34.18	34.12	34.87	32.66	32.10	33.24	34.32	33.42	33.93	34.35	30.55	28.82
MnO	0.57	0.60	0.69	0.70	0.57	0.78	0.73	0.64	0.56	0.78	0.72	0.91	0.86	1.02	0.55	0.52
MgO	7.54	7.09	7.77	1.09	0.03	0.18	0.04	0.96	1.12	0.64	0.07	0.04	0.04	0.08	1.79	2.34
ZnO	0.20	0.15	0.08	-	-	0.03	-	-	-	-	-	-	-	-	-	-
CaO	9.02	8.85	10.85	10.65	10.46	10.57	9.65	10.12	10.04	10.01	9.64	10.19	10.63	9.63	10.09	9.99
Na ₂ O	1.93	2.38	1.43	0.36	1.72	1.39	1.70	1.98	2.05	2.03	1.65	1.23	1.22	1.13	2.01	2.07
K ₂ O	0.76	0.80	0.91	0.25	1.84	2.12	1.02	1.45	1.41	1.13	1.31	1.48	1.19	0.93	1.39	1.25
H ₂ O	1.54	1.26	1.74	1.78	0.98	1.01	1.48	1.29	1.26	1.36	1.46	1.61	1.71	1.68	1.21	1.16
F	0.80	1.39	0.42	0.64	0.47	0.24	0.43	0.68	0.61	0.45	0.29	0.24	0.11	0.07	0.85	1.09
Cl	0.13	0.10	0.18	0.09	2.17	2.50	0.38	0.74	0.87	0.70	0.73	0.23	0.17	0.28	0.71	0.50
O=F+Cl	-0.40	-0.61	-0.22	-0.04	-0.66	-0.69	-0.27	-0.45	-0.45	-0.35	-0.29	-0.15	-0.08	-0.09	-0.52	-0.57
Total	99.70	100.03	101.96	98.56	98.42	99.06	98.12	98.69	97.52	97.82	98.38	97.92	98.75	97.69	97.71	96.72
Structural formula on the basis of 24(O,OH,F,Cl)																
IVSi	7.61	7.53	6.94	7.79	6.41	6.42	6.97	6.58	6.53	5.83	6.62	6.42	6.57	6.94	6.66	6.79
IVAl	0.39	0.47	1.06	0.21	1.59	1.58	1.03	1.42	1.45	2.17	1.38	1.58	1.43	1.06	1.34	1.21
IVTi	-	-	-	-	-	-	-	-	0.02	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.22	0.15	0.34	0.06	0.14	0.17	0.09	0.16	0.09	0.14	0.08	0.46	0.38	0.22	0.14	0.19
Ti	0.04	0.04	0.07	0.02	0.09	0.06	0.12	0.15	0.18	0.28	0.23	0.03	0.03	0.10	0.20	0.20
Mg	1.73	1.63	1.75	0.27	0.01	0.04	0.01	0.24	0.28	0.16	0.02	0.01	0.01	0.02	0.45	0.58
Fe ²⁺	3.01	3.17	2.85	4.65	4.76	4.73	4.78	4.45	4.45	4.41	4.67	4.50	4.58	4.66	4.21	4.01
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	-	-	-	-	0.00	-	-	-	-	-	0.00	0.01
M_{1,2,3}	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Fe ²⁺	0.10	0.12	0.14	0.01	0.14	0.14	0.12	0.11	0.10	0.33	0.17	0.20	0.15	0.16	0.05	-
Zn	0.02	0.02	0.01	-	-	0.00	-	-	-	-	-	-	-	-	-	-
Mn	0.07	0.08	0.09	0.10	0.08	0.11	0.10	0.09	0.08	0.11	0.10	0.13	0.12	0.14	0.08	0.06
Ca	1.48	1.46	1.74	1.87	1.78	1.74	1.74	1.78	1.78	1.56	1.72	1.67	1.73	1.69	1.80	1.78
Na	0.32	0.32	0.02	0.03	-	-	0.04	0.02	0.04	-	0.01	-	-	0.01	0.08	0.16
M₄ site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	0.01	0.00	0.15	0.19	-	0.03	0.04	0.27	0.02	0.17	0.17	0.04	0.00	-
Na	0.25	0.39	0.40	0.09	0.57	0.46	0.52	0.62	0.64	0.67	0.52	0.40	0.39	0.36	0.57	0.51
K	0.15	0.16	0.18	0.05	0.40	0.46	0.22	0.31	0.31	0.25	0.28	0.32	0.25	0.20	0.30	0.26
A site	0.40	0.55	0.59	0.14	1.12	1.11	0.74	0.96	0.99	1.19	0.82	0.89	0.81	0.61	0.87	0.77
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.58	1.29	1.75	1.74	1.11	1.15	1.66	1.43	1.42	1.55	1.64	1.81	1.90	1.89	1.35	1.29
F	0.39	0.68	0.20	0.12	0.25	0.13	0.23	0.36	0.33	0.24	0.15	0.13	0.06	0.04	0.45	0.57
Cl	0.03	0.03	0.05	0.14	0.63	0.72	0.11	0.21	0.25	0.20	0.21	0.06	0.05	0.08	0.20	0.14
F/FM	0.64	0.67	0.63	0.97	1.00	0.99	1.00	0.95	0.94	0.97	1.00	1.00	1.00	1.00	0.91	0.87

Chapter 4: Mica chemistry as an indicator of magmatic to postmagmatic evolution, and halogen fugacities.

4.1 Introduction

The Kwandonkaya igneous complex is located on the eastern edge of Jos Plateau (Fig. 1.1) in the NYG (Nigerian Younger Granite) province. Biotite-bearing peraluminous granites constitute about 50% of the total exposed area of the NYG, and occur either as minor components in predominantly peralkaline complexes in the northern part of the province, or as the main rock types, constituting 95-100% of complexes, in the southern part of the province (Kinnaird 1985). The NYG were the principal source of tin, niobium and tungsten in Nigeria for most of this century. The country was the world's fifth largest producer of tin, and before 1965, 95% of the world's columbite supplies came from the NYG. Both cassiterite and columbite were produced from eluvial and alluvial reserves originating from decomposition and erosion of biotite granite, and wolframite was produced from primary mineralization also associated with the biotite granites. Although mining has declined since the mid-seventies, the reserves are still extensive (Kogbe & Obialo 1976, Kinnaird 1987). It is only within biotite granites, or in alterations of country rock related to their emplacement, that post-joint greisens, quartz-wolframite, disseminated-, and lode-type Sn-Nb-W $\pm$ (Pb-Zn) mineralization have been documented in the NYG. This chapter focuses on crystal chemistry of the biotite and its relation to the evolution of biotite-bearing granites in the Kwandonkaya Complex, and relates these findings to results on other subvolcanic-plutonic complexes. A study of chemistry of the biotite is important in order to characterize the late-magmatic, subsolidus, and hydrothermal stages of biotite-bearing granites in the NYG, and to account for the interplay of alteration and mineralization. The nomenclature used follows the usage of Hawthorne and Černý (1982), Bailey (1984a, b), and Černý and Burt (1984); the classification of Li-Fe and Li-Al micas is after Foster (1960) and Tindle and Webb (1990).

Previous studies of mica chemistry have been undertaken as part of province-wide studies of ferromagnesian minerals in the NYG. Borley (1976) reported that relatively magnesian biotite occurs in earlier intrusions in the Sara-Fier Complex, and that annitic biotite occurs in riebeckite biotite granite and hornblende biotite granite in the Amo Complex. In a study of the Afu complex in the NYG, Imeokparia (1982) concluded, from studies of Sn content in biotite, that the Sn was mobilized during alteration of biotite to sericite and chlorite (during greisenization), and deposited in mineralized zones. In their studies of complexes in the northern part of NYG province, Bennett et al. (1984) concluded that 1) Fe-rich mica crystallized at a magmatic stage, whereas Al-Li-rich micas are postmagmatic, and 2) postmagmatic alteration and greisen formation occurred in a F-rich environment. Bowden et al. (1987) also documented a range of mica compositions associated with a variety of types of alteration: protolithionite-zinnwaldite-trilithionite in Na-metasomatized rocks, annite - ferroan siderophyllite in K-metasomatized zones, and lithian siderophyllite in greisens. Kinnaird et al. (1992) identified annitic mica as the magmatic ferromagnesian mineral typical of peraluminous biotite granites of the NYG; she documented its replacement to varying extent by chlorite and sericite, whereas the Li-Fe and Li-Al micas were found to develop at the expense of chlorite and sericite.

Geological mapping of Kwandonkaya Complex was initiated by Bain & Russ (1923), and the complex was re-surveyed during wartime efforts to locate tungsten resources (Jacobson & Jaques 1944). Buchanan (1971) published an updated geological map from a field survey initiated R. Black in 1956. Imeokparia (1984) analyzed biotite in "hornblende" (quotation marks indicating a provisional name that conforms to previous statements in the literature on NYG) biotite granite porphyry, coarse-grained (Panshanu) biotite granite, medium-grained (Guskai) biotite granite, and fine-grained (Jubga) biotite granite as part of a geochemical study of the complex. In his study, all the mica analyzed was found to be annitic biotite, with the proportion of F substantially greater

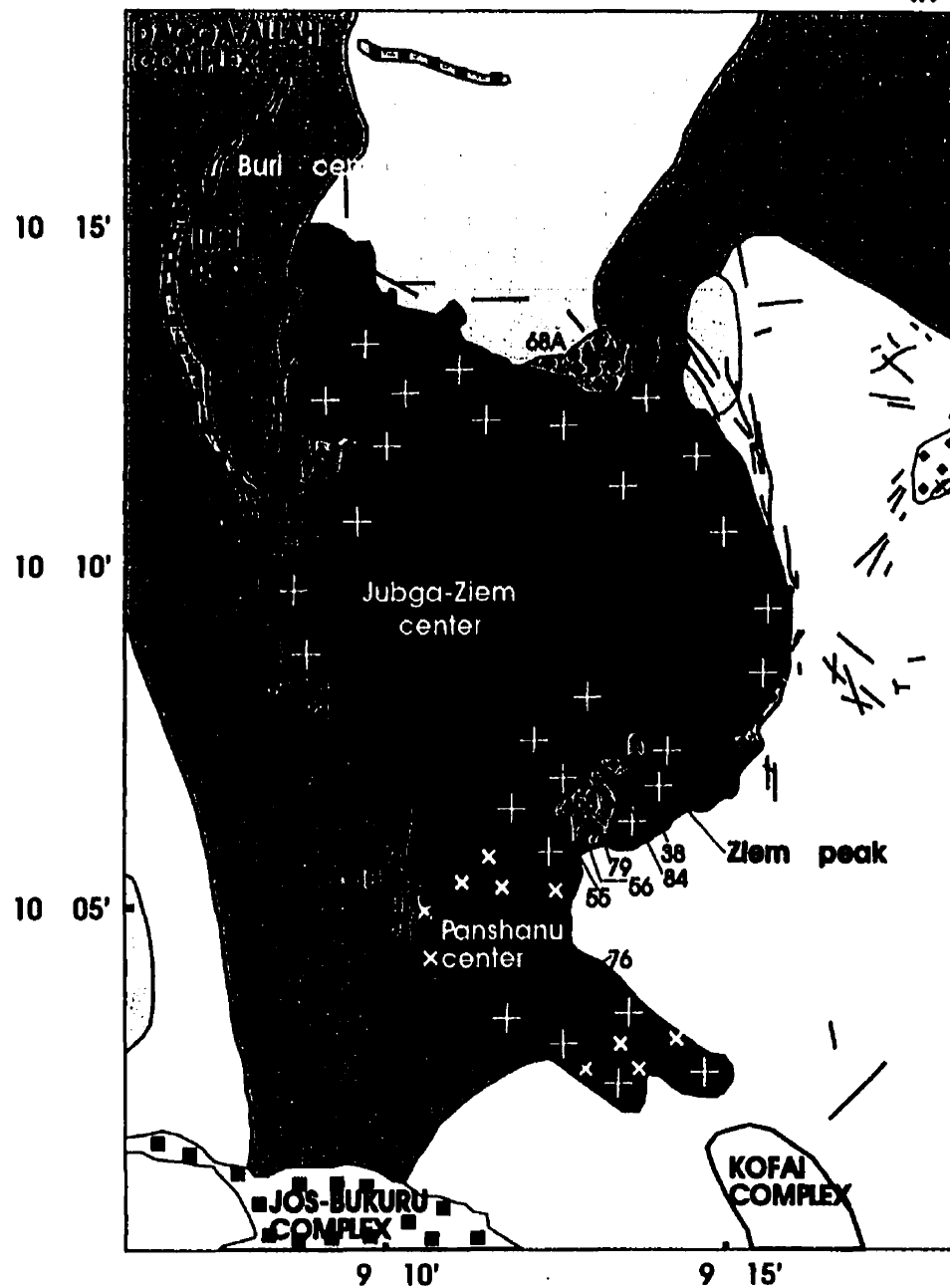
than Cl; he concluded that mica in the Kwandonkaya complex crystallized at $f(\text{O}_2)$ conditions slightly above the QFM buffer. The spectrum of rocks sampled in this study is broader, and includes micas from the Buri ring dyke, biotite granite porphyry, columbite-bearing albitized biotite granite, and greisen zones not investigated in previous studies of the complex. In this study, magmatic and postmagmatic mica are distinguished, and this information, along with relevant chemical data, are used to indicate the nature of the residual melt at late stages of its crystallization, and of the hydrothermal fluids that caused alteration and mineralization. This study of micas complements the investigation of fayalite, hedenbergite, and amphiboles in subvolcanic rocks of the Kwandonkaya Complex (Chapter 3).

4.2 Distribution of biotite-bearing rock units

The Kwandonkaya complex is composed of three overlapping intrusive centers that have a NNW-SSE alignment: Buri, Jubga-Ziem, and Panshanu (Fig. 4.1). Generally, cross-cutting relationships, established in chapter 2, indicate a migration of intrusive centers: the Buri center is truncated by the Jubga-Ziem center, and the latter is cut by the Panshanu center (Fig. 4.1). The relationships between the different units have been previously documented (Chapter 2), and only biotite-bearing rocks are discussed here.

Rhyolite porphyry is limited in occurrence to the Buri center, exposed only north of Lusi peak, forming an extension from the main granite porphyry ring-dyke (Fig. 4.1). Fayalite syenogranite, "hornblende"-biotite granite porphyry and biotite granite porphyry are preserved in the marginal parts of the Jubga-Ziem center. Fayalite syenogranite occurs in the inner parts of some bodies of "hornblende"-biotite granite porphyry, with some exposures of fayalite syenogranite varying transitionally from fine-grained to coarse-grained types. Typically, the samples of "hornblende"-biotite and biotite granite porphyries vary from drusy and porphyritic to

Fig. 4.1: Geological map of Kwandonkaya complex with location of samples whose mica was analyzed.



• Sample location

Basalt dyke

MESOZOIC GRANITES

Plutonic rocks and altered variants

G: Gneiss

■ Fine- to medium-grained biotite granite

+ Coarse-grained biotite granite

x Porphyritic biotite microgranite

Subvolcanic rocks

■ "Homblende"-biotite, biotite granite porphyry, +/- fayalite syenogranite

■ Hedenbergite fayalite granite porphyry

■ Rhyolite porphyry

■ Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

■ Biotite homblende granite

■ Gneissic granite

■ Migmatite

1 0 2 4 6 8 km

miarolitic variants along contacts with the basement complex or earlier biotite granite, to normal "hornblende"-biotite granite porphyry inward from the contacts. The emplacement of the Jubga-Ziem ring dyke represents the next event because it cross-cuts the southern part of the Buri ring dyke. The Jubga-Ziem ring dyke was later intruded by several textural variants of biotite granite.

Coarse-grained biotite granite occurs as arcuate units along margins in the Jubga-Ziem and Panshanu centers. In the Jubga-Ziem center, the continuity of the arcs is breached by cross-cutting, later, fine- to medium-grained biotite granite. The coarse-grained biotite granite cross-cuts the Jubga-Ziem ring dyke and is intruded by later fine- to medium-grained biotite granite. In the Panshanu center, the coarse-grained biotite granite has a similar relationship to the fine- to medium-grained biotite granite.

Medium- to fine-grained biotite granite occupies the central parts of the Jubga-Ziem and Panshanu centers, in contact with the coarse-grained biotite granite and the Pan-African basement. This rock type includes drusy, equigranular, and porphyritic types. Albitized variants occur in marginal parts. Post-joint greisens are limited in occurrence to porphyritic and fine-grained biotite granites within the Jubga-Ziem center, with alteration zones that include: a central quartz vein, quartz-rich greenish to grey greisen zone, and an outer reddened granite zone.

Table 4.1 is a summary of petrographic features of mica-bearing rock units.

4.3 Petrography of the subvolcanic rocks

Within the Buri ring dyke, *rhyolite porphyry* invariably contain phenocrysts of fayalite, hedenbergite, $\pm$ amphibole. This rock type does not contain primary biotite. Biotite and chlorite both replace amphiboles and hedenbergite. Polygranular micrometric, ragged flakes of biotite on Fe-Ti oxides are attributed to subsolidus re-equilibration of the opaque phases.

In the Jubga-Ziem ring dyke, the marginal parts of *fayalite syenogranite* have biotite

Table 4.1: Petrographic features of mica-bearing rock types in the Kwandonkaya complex.

Rock Type (texture)	Minerals	Main mica feature(s)	Mica types/pleochroic scheme
Rhyolite porphyry (porphyritic)	Or, Qtz, Hd, Fa, Fe-Ed, minor [Fe-Act, Ann-Bt, Chl]; {Ckt, Zrn, Ilm, Mag, Ttn, Ap}	Secondary, replaces Amp and Cpx.	Ann-Bt: $\alpha = \text{gr/brn}$, $\beta \equiv \gamma = \text{rds-brn}$
Fa syenogranite (coarse-, fine-grained, drusy)	Or, Qtz, Fa, Ann-Bt, [Idn, Grn, Chl]; {Zrn, Fl, Fst, Flc, Aln, Ckt, Ilm, Mnz}	Lm/Sb; μm -sized to 0.7x0.5 mm, rarely 1x1.5 mm.	Ann-Bt: $\alpha = \text{gr/brn}$, $\beta \equiv \gamma = \text{dark gr/rds-brn}$
Hbl Bt granite porphyry (drusy, glomeroporphyritic)	Or, Qtz, Ab, Fe-Ed, Hbl, Ann-Bt, [Src]; {Zrn, Aln, Fl, Xnt, Fct}	Lm/Sb; μm -sized to 1.2x1.5 mm.	Ann-Bt: $\alpha = \text{gr/brn}$, $\beta \equiv \gamma = \text{dark grn/brn}$
Bt granite porphyry (glomeroporphyritic)	Or, Qtz, Ab, Bt, minor [Chl, Src]; [Ilm, Zrn, Fl]	Lm/Sb; μm -sized to 0.2x1 mm.	Ann-Fp: $\alpha = \text{gr/brn}$, $\beta \equiv \gamma = \text{dark gr/brn}$
Coarse-grained Bt granite (coarse granular)	Or, Mc, Ab, Qtz, Bt, [Chl, Src]; [Ilm, Zrn, Flc, Aln, Mnz, Hem, Ap, Mag]	Lm/Sb; 0.1x0.2 mm to 1.5x3 mm, with 3x4 mm aggregates.	Ann-Fp-Znw: $\alpha = \text{gr/brn}$, $\beta \equiv \gamma = \text{rds-brn}$; and T: pl-clr (weakly pleochroic)
Fine- to medium-grained Bt granite (equigranular, porphyritic, drusy)	Or, Mc, Ab, Qtz, Bt, [Chl, Src]; [Ilm, Zrn, Tht, Aln, Mnz, Xnt, Hem, Ap, Fl, Mag]	Lm/Sb; some zoned grains in drusy granite; μm -sized to 1.1x2.5 mm.	Fp-Znw & Srp: $\alpha = \text{pl-gr/brn/bl/ylw}$, $\beta \equiv \gamma = \text{dark gr/brn/bls-gr/ylw-brn}$; and T: pl-clr
Albitized Bt granite (Fine granular or sugary)	[Mc, Ab, Qtz, Bt, Chl]; [Col, Cst, Toz, Zrn, Mag, Mnz, Fl, Aln, Tht, Hem]	M/H; μm -sized to 0.3x2.5 mm.	Fp-Znw: $\alpha = \text{pl-gr/brn}$, $\beta \equiv \gamma = \text{dark gr/brn}$, and T: pl-clr
Post-joint greisens (even-grained, drusy)	[Qtz, Mc, Ab, Bt, Chl, Src]; {Cst, Toz, Col, Zrn, Tht, Rt, Aln, Hem}	M/H; μm -sized to 0.6x0.7 mm, with 2x5 mm aggregates.	Fp-Srp: $\alpha = \text{pl-gr}$, $\beta \equiv \gamma = \text{dark grn/brn}$, and T: pl-cl

Mineral abbreviations used (mostly after Kretz 1983) are: **Ab**: albite, **Aln**: allanite, **Ann-Bt**: annitic biotite, **Ap**: apatite, **Bt**: biotite, **Cam**: calcic amphibole, **Chl**: chlorite, **Ckt**: chevkinite, **Col**: columbite, **Cst**: cassiterite, **Fa**: fayalite, **Flc**: fluocerite, **Fe-Act**: ferroactinolite, **Fe-Ed**: ferroedenite, **Fe-Prg-Hbl**: ferropargasitic hornblende, **Fp-Znw**: ferroprotolithionite-zinnwaldite, **Fl**: fluorite, **Fst**: fergusonite, **Grn**: grunerite, **Hbl**: hornblende, **Hd**: hedenbergite, **Hem**: hematite, **Hs**: hastingsite, **Idn**: iddingsite, **Ilm**: ilmenite, **Mag**: magnetite, **Mnz**: monazite, **Mc**: microcline, **Qtz**: quartz, **Rt**: rutile, **Src**: sericite, **Srp**: siderophyllite, **Tht**: thorite, **Toz**: topaz, **T**: trilitionite, **Ttn**: titanite, **Xnt**: xenotime, **Znw**: zinnwaldite, and **Zrn**: zircon. Accessory minerals in curly brackets, and secondary minerals are in square brackets. *Colour abbreviations*: **gr**: green, **brn**: brown, **rds-brn**: reddish brown, **pl-clr**: pale coloured, **pl-gr**: pale green, **bl**: blue, **ylw**: yellow, **bls-gr**: bluish green, and **ylw-brn**: yellowish brown.

Other abbreviations: **Lm/Sb**: late magmatic/subsolidus; **M/H**: magmatic/hydrothermal.

predominating over fayalite, whereas in the more voluminous inner zone, fayalite is predominant. Some biotite grains are interstitial and possess a serrate margin, suggesting late-magmatic crystallization. However, the majority of the biotite grains occur mainly within microperthite or albite. Some samples show the boundary between altered fayalite and orthoclase microperthite to be characterized by annite and quartz along with ilmenite and fergusonite. This most likely results from the subsolidus reaction: fayalite + sanidine_{ss} + volatiles = annite + quartz (cf. Eugster & Wones 1962).

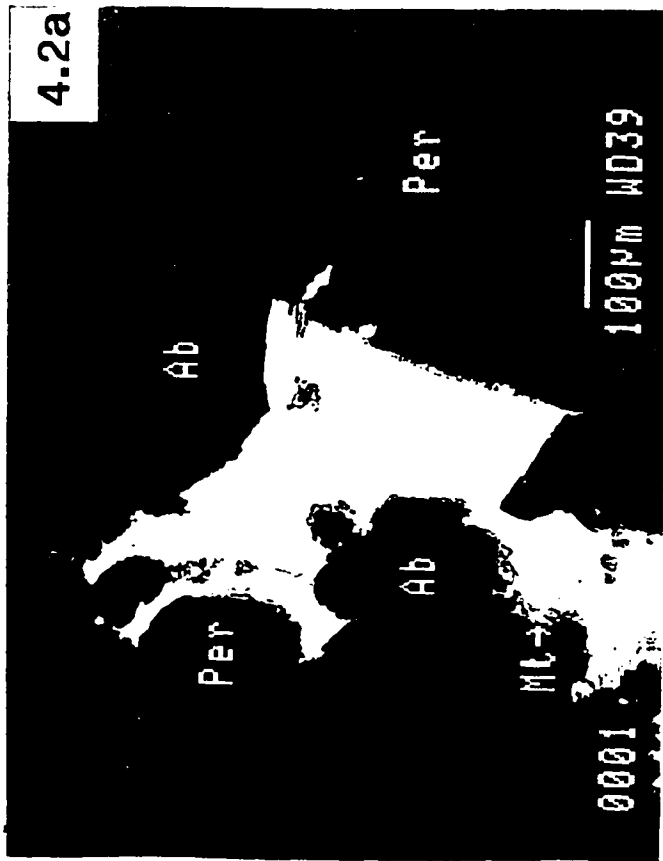
In "*hornblende*"-biotite and biotite granite porphyry, also in the Jubga-Ziem ring dyke, biotite varies from a subhedral and interstitial type that is intergrown with microperthite, both possessing a serrate or ragged boundary (Fig. 4.2a, Appendix IV.1a), to biotite intergrown with "*hornblende*" (Fig. 3.6a, chapter 3) or partly enclosing allanite (Fig. 4.2b). The serrate or ragged boundary with microperthite is more consistently developed at boundaries between mica and albite (Fig. 4.2a), and, in fact seems to postdate exsolution; it could arise from subsolidus exchange reaction of the type $K \rightleftharpoons Na$ between albite and mica to produce the swapped boundaries in an overgrowth on the "primary" biotite. It is inferred that the interstitial and ragged appearance of biotite grains (Appendix IV.1a) indicates that the crystallization of mica spanned the late-magmatic to subsolidus stages, with the latter postdating exsolution in alkali feldspar. The intergrowth of biotite and "*hornblende*" in some sections (Fig. 3.6a) suggests that the stability field of amphibole overlapped with that of annite + quartz. Textural evidence suggests that biotite also postdates the appearance of zircon, allanite, and ilmenite.

4.4 Petrography of the plutonic and altered rocks

These rock types, predominant in the Jubga-Ziem and Panshanu centers, are composed of different textural variants of biotite granite, as exposed in the two composite intrusive centers

Fig. 4.2: SEM-BSE (Scanning electron microscope - back-scattered electron) images of mica in "hornblende"-biotite granite porphyry (sample 68B, Fig. 4.1). **a)** Interstitial biotite (Bt) with serrate boundary and vermicular extensions into albite (Ab) domains of microperthite (Per); Mt: magnetite. The serrate or vermicular intergrain boundary could arise from an exchange reaction involving albite and mica. **b)** Biotite (Bt) grains partly enclosing a zoned crystal of allanite (Aln) with low BSE or darker zones that have lower ΣREE than the brighter parts. Some grains of biotite have inclusions of microperthite. Qtz: quartz

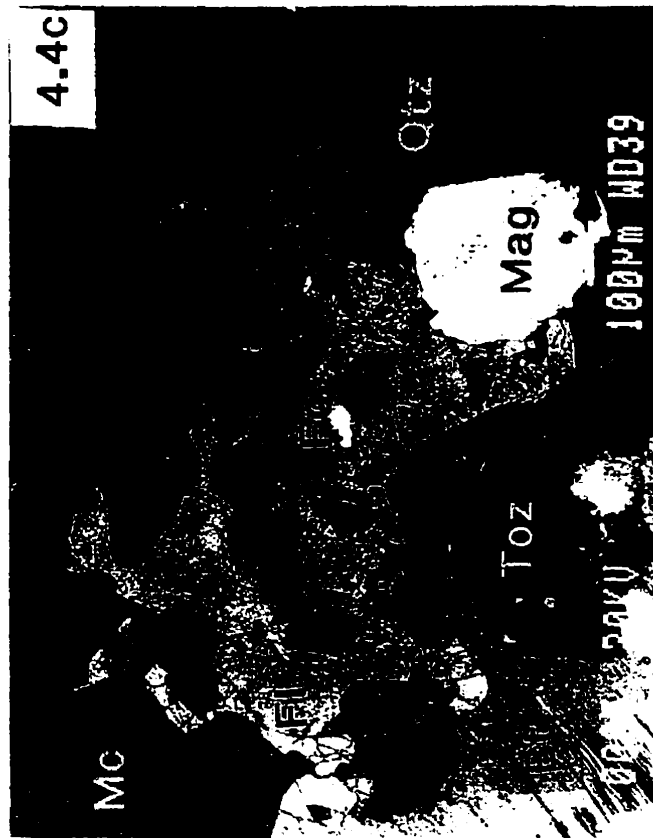
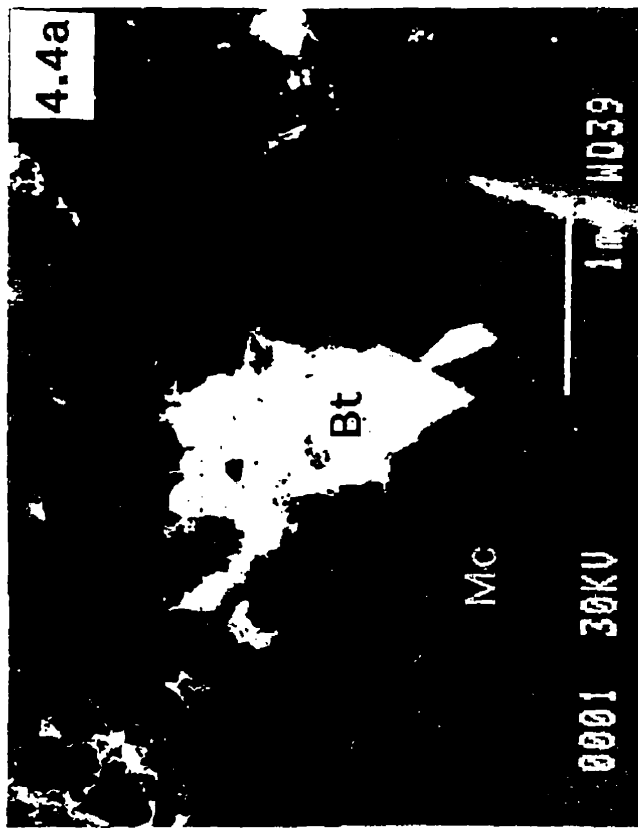
Fig. 4.3: SEM-BSE images of mica in biotite granite: **a)** Biotite (Bt) in a coarse-grained biotite granite (sample 26A, Fig. 4.1) with a core composed of microcrystalline (graphic-like) intergrowths of quartz (dark grey) and biotite (medium grey). Inclusions of light grey ilmenite (Ilm), and specks of monazite (very light) occur along with subsidiary fluorite (Fl) and zircon. The centre-right part of the picture has inclusions of fluocerite (Fc) that seem primary and are replaced principally by parisite and minor allanite. **b)** Biotite (Bt) in hypersolvus fine-grained biotite granite (sample 83, Fig. 4.1) with an inclusion of magnetite bearing exsolved ilmenite lamellae (Ilm-Mag). Pale mica (the parts that seem darker) occur as patches that are irregularly distributed and appear to be produced by "droplets" of acidic fluid interacting with the dark mica.



(Fig. 4.1). In the *coarse-grained biotite granite*, the biotite forms subhedral interstitial grains, with most of the grains possessing a serrate margin where in contact with albite domains in the perthite. Some of the interstitial biotite has inclusions of accessory minerals, with such inclusions being more abundant in poikilitic cores, whereas marginal parts are relatively uniform (Fig. 4.3a). Some other sections of the rock show biotite to be replaced by pale-coloured mica, sericite, or chlorite. Overall, the interstitial nature of the biotite indicates late-magmatic crystallization, whereas the ragged appearance of the biotite indicates an origin by local replacement in the subsolidus stages, postdating exsolution in the alkali feldspar.

The *medium- to fine-grained biotite granites* varies from equigranular, porphyritic, to drusy types. Three distinct types of mica are distinguished. a) First is the more typical (dark) mica, which constitutes the dominant ferromagnesian mineral in over 95% of the volume of these rock units. The hypersolvus variety of fine-grained biotite granite contains biotite with occluded magnetite that has exsolved lamellae of ilmenite (Fig. 4.3b). In most sections, both interstitial biotite, possessing sharp boundaries, and ragged biotite replacing microperthite and albite are found. The interstitial biotite is interpreted to be late-magmatic, whereas the ragged biotite is subsolidus. In some sections, both types of biotite are replaced to different extents by fluorite, Fe-Ti oxides, chlorite, sericite, or muscovite. b) The second type of mica is pale green to bluish green, and is typical of the drusy "biotite" granite exposed on the crest of Ziem ridge (sample 37). The mica is biaxial positive and is tentatively identified as lithian siderophyllite (cf. Foster 1960, Wilcox 1984). The relatively larger grains (0.5 x 1 to 0.7 x 1.5 mm) are interstitial and subhedral, and also are zoned (Fig. 4.4a), displaying a transitional variation in colour, with pale yellowish green, colourless, and relatively darker (lilac) bluish green zones. Some of the zoned mica seems to have undergone postcrystallization homogenization so as to yield a nearly uniform grain with relics as evidence of the zoning (Appendix IV.1b). The subhedral and interstitial character, serrate

Fig. 4.4: SEM-BSE images of mica in drusy biotite granites (a, b), and altered rocks (c, d). a) Zoned mica (Bt) in drusy fine-grained biotite granite (sample 37, Fig. 4.1): darker part (zinnwaldite-siderophyllite) has slightly higher Si, Al, K, F, and Li and lower Ti, Fe, Na, and Cl than the lighter part (trilithionite). Some of the zoned mica sections exhibit different extents of subsolidus homogenization of the zoning. b) Biotite (Bt) in topaz-cassiterite-bearing fine-grained biotite granite (sample 30A, Fig. 4.1). The biotite is partly replaced by patches of pale-coloured mica (trilithionite). c) Biotite (Bt) associated with fluorite (Fl), topaz (Toz), magnetite (Mag), and fluocerite (Fc) within albitized biotite granite (sample 23C, Fig. 4.1). d) Pale-coloured mica in post-joint quartz-mica greisen (sample 36C, Fig. 4.1). The mica bears inclusions of thorite (Tht) as well as Hf- and Th-rich zircon (Zrn).



margin of some grains, along with the presence of zoning, suggest crystallization at late-magmatic to subsolidus stages during loss of volatiles to produce zoned mica in drusy granite. c) The third type of mica is pale coloured, and biaxial negative. This is tentatively identified as zinnwaldite, based on its similarity in characteristics to zinnwalditic mica documented by Kinnaid et al. (1992) and Wilcox (1984). In most sections examined, the zinnwaldite occurs as distinct grains partly replaced by pale-coloured mica (Fig. 4.4b), or intergrown with the dark mica. Overall, the zinnwaldite constitutes between 10 to 20% of the mica present, but in some marginal fine-grained granite bearing disseminated cassiterite and topaz, the zinnwaldite constitutes over 95% of the mica in the rock. The pale-coloured mica spans the subsolidus range of crystallization.

The altered rocks include *albitized biotite granite*, which is fine-grained, with equigranular and porphyritic types. Unlike biotite in the fresher biotite granite, the biotite here is very fine-grained and predominantly anhedral (Fig. 4.4c). Another distinctive feature is the occurrence of microscopic lamellae and specks of hematite (in mica and feldspar) that produce hematite speckling in hand specimen. The biotite is replaced to various extents by chlorite, fluorite, $\pm$ Fe-Ti oxides. In some of the samples, a pale-coloured mica is prominent, which, along with Fe-Ti oxides, replaces the pre-existing primary greenish brown biotite. Overall, the biotite in this rock type crystallized during hydrothermal alteration of marginal parts of fine- to medium-grained biotite granite.

In the *greisens*, the mica fraction includes siderophyllite, annitic mica, and a pale to colourless variant. In the less altered wall zones adjacent to the greisen, ragged microcrystalline aggregates of mica occur. In the inner zones of greisens, two types of hydrothermally deposited micas are distinguished: i) relatively coarse (up to 1 x 2 mm), pale green to dark brown biotite. Such grains are distinctly interstitial in quartz-rich domains, or partly enclose quartz. This type of mica is minor and appears to have crystallized in vugs, after silicification. ii) Fine-grained (μm -

sized to 0.3 x 0.5 mm) colourless or pale green to dark bluish green mica, which is tentatively identified as siderophyllite and forms the dominant type of mica in greisens, occurring as felted anhedral grains (Fig. 4.4d), or multigranular aggregates intergrown with microcrystalline quartz. In some cases, the mica is colourless and pleochroic to pale yellowish brown, and thus probably tends to zinnwaldite. Some of the mica aggregates, which replace turbid K-feldspar, are chloritized.

4.5 Mica Chemistry

The sequence of presentation is: analytical methods, followed by chemical variations in the mica.

4.5.1 Analytical methods

Chemical data were collected by wavelength-dispersion spectrometry (WDS) using the Cameca CAMEBAX electron microprobe at McGill University. An accelerating voltage of 15 kV, a beam current of 20 nA, spot size of 4 μm , and counting time of 25 seconds were used, except for Cl and F, for which 50 and 100 seconds, respectively, were used. The following natural and synthetic standards were used: andradite (Fe), MnTi-oxide (Mn, Ti), albite (Na), diopside (Si, Mg, Ca), orthoclase (K,Al), fluorite (F), and vanadinite (Cl). Some of the mineral standards as well as a biotite standard were analyzed intermittently to check for drift in standardization. The raw data were ZAF-corrected and reduced using online ONQANT software package. SEM studies were undertaken to characterize inclusions of accessory minerals in mica, and to establish the pattern of zoning observed optically.

Tindle & Webb (1990) derived a relationship between mica composition (SiO_2 content) and Li_2O from a database of over 400 analyzed samples, which builds on the observed

relationships of Foster (1960), Monier & Robert (1986) and Stone et al. (1988). The relationship (given in Table 4.2) provides an inferred Li_2O value within ± 1 wt.% of wet-chemical determinations. The method is ineffective for low-Li, Fe-rich micas (with less than 1.5% Li_2O) because the error of ± 1 wt.% is relatively significant in such cases.

The structural formula was calculated on the basis of 24 (O,OH,F,Cl): $\text{X}_2\text{Y}_6\text{Z}_8\text{O}_{20}(\text{OH},\text{F},\text{Cl})_4$ (Table 4.2). The assumption that all Fe is present as FeO is made, as it was by Bennett et al. (1984) and Tindle & Webb (1990). The actual amount of Fe^{3+} present in the micas is dependent on the $f(\text{O}_2)$ and composition of the system (Eugster & Wones 1962, Wones 1972). The proportion of three mica components (annite, siderophyllite, and phlogopite) was calculated after relationships derived to describe phlogopite-annite-siderophyllite solid solutions (Gunow et al. 1980), and given in footnotes in Table 4.2. This method of calculation is based on an assumption that the proportion of other cations (e.g., Ti, Mn, Li, Fe^{3+}) in the Y site is very low. The method of Munoz (1984, 1992) was adopted to calculate halogen fugacities of the fluids that re-equilibrated with the micas. Table 4.2 presents averaged compositions of mica for individual samples, and the location of samples analyzed are given in Figure 4.1. In cases of zoning or where more than one type of mica was analyzed in a sample, separate averages are reported for the micas. The chemical data are presented in the sequence: subvolcanic rocks in ring dykes, plutonic rocks, and albitized and greisenized plutonic rocks.

4.5.2 Variations in mica chemistry

Mica composition, expressed in terms of the components annite (Ann), siderophyllite (Srp), and phlogopite (Phl), in individual rock units is presented in Table 4.3, and the detailed data (results of spot analyses) are given in Appendix IV.2 - 4. This presentation emphasizes variations of mica compositions among the various rock units, as well as variation of mica compositions

between subvolcanic and plutonic units. The variations in mica chemistry are presented in the order late-magmatic to subsolidus mica, followed by secondary mica.

Late-magmatic to subsolidus mica

The late-magmatic to subsolidus micas include annitic biotite, and ferroprotolithionite - zinnwaldite - siderophyllite.

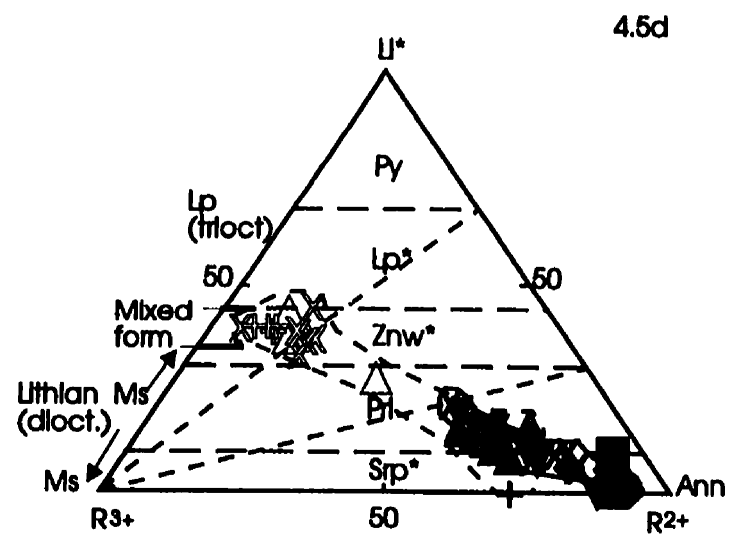
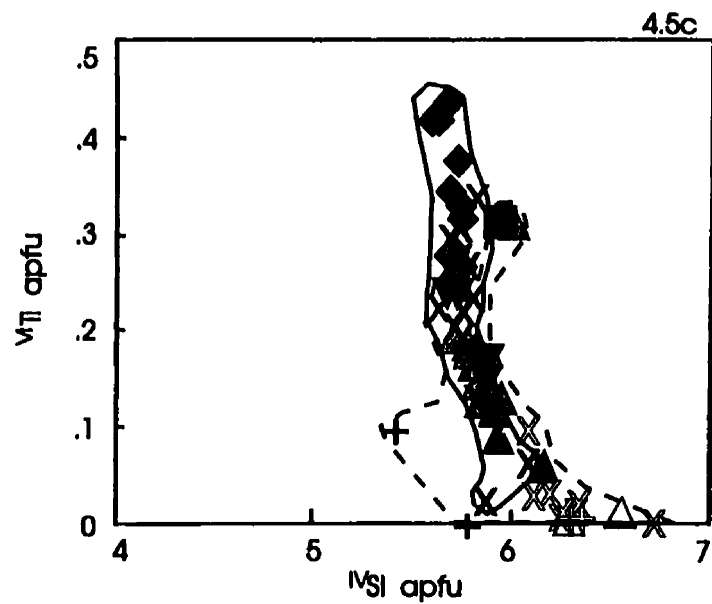
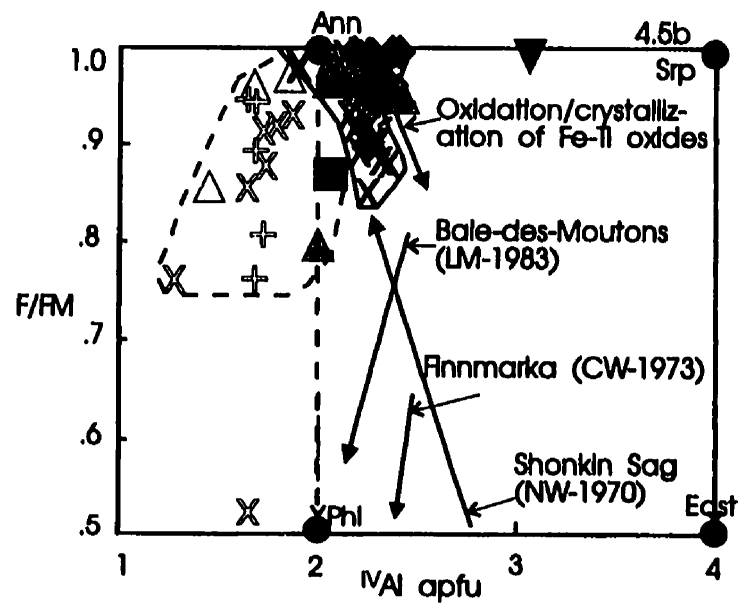
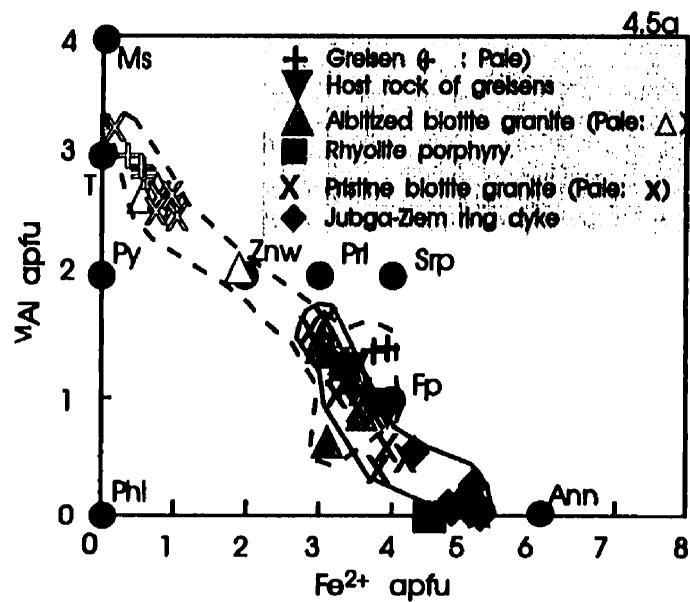
Annitic biotite occurs in the fayalite syenogranite and "hornblende"-biotite granite porphyry of the Jubga-Ziem center, and hypersolvus biotite granite (Fig. 4.5a). There are no clear chemical distinctions between the late-magmatic and replacement-type annitic biotite (Fig. 4.5a). The annitic biotite has ^{VI}Al content (expressed by Srp or siderophyllite content) which increases from the fayalite syenogranite (Srp₂₇₋₃₉) to "hornblende"-biotite granite porphyry (Srp₂₆₋₅₂) and biotite granite porphyry (Srp₄₉₋₅₈). In the biotite granite porphyry, the mica's composition has thus the lowest annite and highest siderophyllite content of the dark mica in the subvolcanic rocks; it is transitional in this sense to that in the plutonic rocks (Fig. 4.5a, Table 4.3). The F/FM (0.95) of the dark mica in the biotite granite porphyry is lower than that of the whole rock (1.0). This disparity presumably reflects the fact that magnetite cocrystallized with biotite, because of locally higher $f(O_2)$. This condition is known to result in the formation of biotite apparently enriched in Mg relative to the bulk rock (Eugster & Wones 1962, Anderson et al. 1980).

The annitic mica in the hypersolvus variant of biotite granite is also Fe-enriched (F/FM = 1.00), and has the highest Ti content ($^{VI}Ti = 0.34$ apfu) in the plutonic biotite granites. This may be due to the relatively higher temperature of crystallization of hypersolvus biotite granite (cf. De Albuquerque 1973), or to relatively low $f(O_2)$, and build-up of Ti in the melt, with ilmenite postdating the crystallization of biotite.

The dark mica in the series ferroprotolithionite - zinnwaldite - siderophyllite documented

Fig. 4.5: Plots of ^{VI}Al vs. Fe^{2+} (a), F/FM vs. ^{IV}Al (b), ^{VI}Ti vs. Si (c), and Triangular plot of $R^{3+} - R^{2+} - Li^+$ (d). a) ^{VI}Al apfu vs. Fe^{2+} apfu. Abbreviations used in this and other Figures: **Ann** = annite, **Phl** = phlogopite, **Fp** = ferroprotolithionite, **Prl** = protolithionite, **Py** = polyolithionite, **T** = trilithionite, **Znw** = zinnwaldite, **Lp** = lepidolite, **Ms** = muscovite, **East** = eastonite, **diact.** = dioctahedral, **triact.** = trioctahedral. b) F/FM $[Fe/(Fe+Mg)]$ vs. ^{IV}Al apfu. The data envelope enclosing dark mica (solid line) shows that the transition from subvolcanic to plutonic rocks involved increase in ^{IV}Al with a decrease in F/FM, whereas the data envelope enclosing open symbols (dashed line) indicates that the formation of pale mica involved decreases in ^{IV}Al and F/FM. Trends of dark mica composition from Baie-des-Moutons (LM-1983: Lalonde & Martin 1983a, b) and Finnmarka complex (CW-1973: Czamanske & Wones 1973) define an oxidation trend, whereas the Shonkin Sag (NW-1970: Nash & Wilkinson 1970) illustrates a reducing trend. c) ^{VI}Ti vs. Si apfu. Alteration trends associated with depletion of Ti include sodic, potassic and acid metasomatism, and chloritization. d) Triangular plot of $R^{3+} - ^{VI}(Al^{3+}+Ti^{4+}) - R^{2+} - ^{VI}(Fe^{2+}+Mn^{2+}+Mg) - Li^+$ (calculated Li). Mixed form indicate range for mixed dioctahedral-trioctahedral mica, along the muscovite - lepidolite join. Horizontal boundaries are after Foster (1960); asterisk indicates mica compositions defined by the short dashed (diagonal) lines after Stone et al. (1988). The pale mica is intermediate between dioctahedral lithian muscovite and trioctahedral lepidolite (=mixed form).

In all the Figures, the two envelopes of data show dark mica (solid line), and pale mica (dashed line).



in the coarse-grained biotite granite and pristine biotite granite is inferred to be primary. The composition of this mica in coarse-grained biotite granite is similar to that in the biotite granite porphyry (Table 4.3). The drusy biotite granite, on the other hand, contains a dark mica of ferroprotolithionite-zinnwaldite composition ($F/FM = 0.98$, $^{VI}Ti = 0.08$ apfu, $F/Cl = 213$), and pale zones of trilithionite. The trilithionite ($F/FM = 0.97$, $^{VI}Ti = 0.04$ apfu, and $F/Cl = 883$) in oscillatory zones (within predominant siderophyllite) has slightly higher Si, Al, K, Mg and F, and lower Ti, Fe, Na and Cl than the dark mica (siderophyllite). In this instance, the trilithionite is considered primary, having crystallized from volatile-rich melt. The F content of some pale zones reach 5.18 wt.% ($X_F = 0.62$), which correlates with the inferred degree of Li enrichment. The host rock to the greisens is also characterized by a dark mica ($F/FM = 0.97$) in the compositional range of ferroprotolithionite-zinnwaldite, and is generally unaltered.

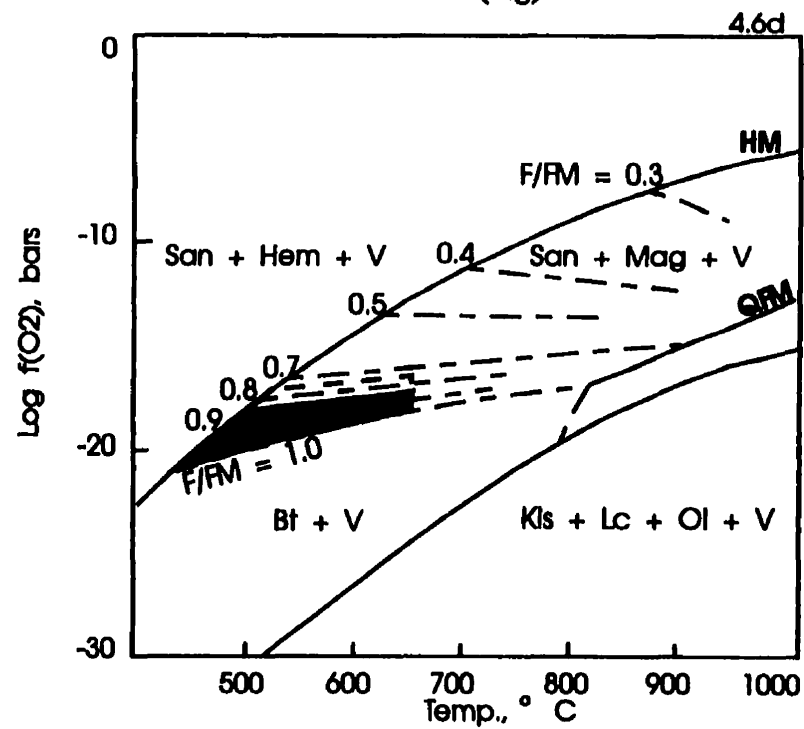
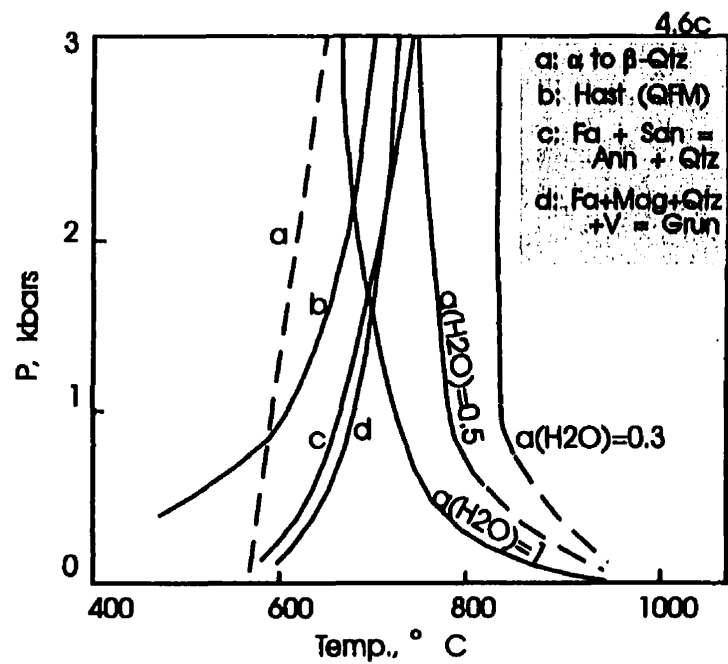
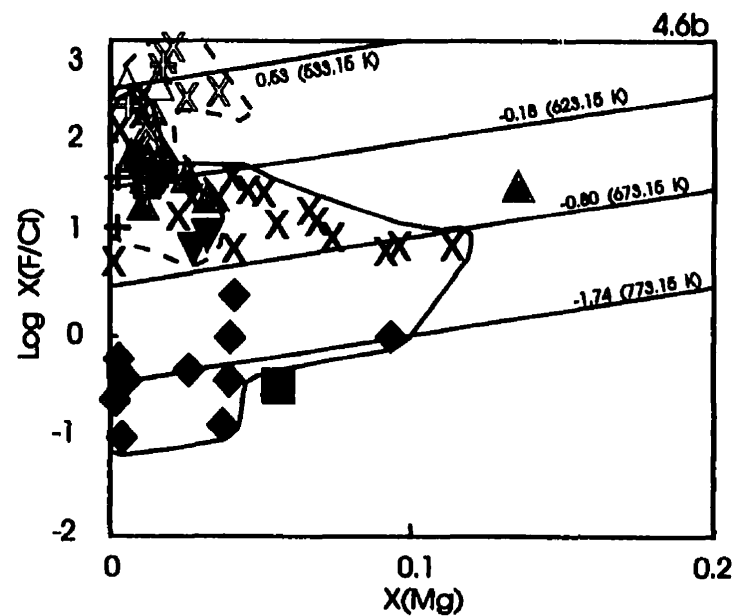
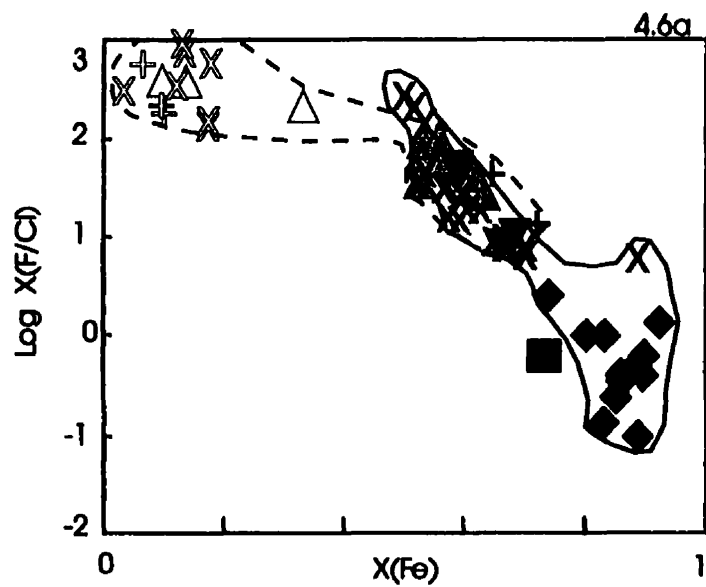
The decrease in F/FM of mica with evolution from subvolcanic to plutonic rocks in this study (Fig. 4.5b) reflect higher levels of $f(O_2)$ in the plutonic rocks, generalized by the reaction (after Wones & Eugster 1965): $2K(Fe_2Mg)_{\Sigma 3}AlSi_3O_{10}(OH)_2 + O_2 \rightleftharpoons K(Mg_2Fe)_{\Sigma 3}AlSi_3O_{10}(OH)_2 + KAlSi_3O_8 + Fe_3O_4 + 2H_2O$. Magmatic oxidation has been documented in rock suites characterized by decreasing of F/FM as observed in the mica from Kwandonkaya subvolcanic rocks (Fig. 4.5b). This is similar to observations in the Finnmarka complex (Czamanske & Wones 1973), Baie-des-Moutons complex (Lalonde & Martin 1983a, b), and Murdock Creek intrusion (Rowins et al. 1991).

In summary, the principal variations in mica composition (annitic biotite $\rightarrow$ ferroprotolithionite-siderophyllite-zinnwaldite) include:

1) The increase in the ^{VI}Al (and ΣAl), Li, as well as Si with decreasing ^{IV}Al , ΣTi , and Fe^{2+} of the dark mica from subvolcanic to plutonic rocks (Figs. 4.5a-d, 4.6a, b). These increases indicate substitutions of the type: $^{VI}(Al, Li)^{IV}Si^{VI}R^{2+}_{.1}{}^{IV}Al_{.1}$ (where $R^{2+} = Fe, Mg, \text{ and } Ti$)

Fig 4.6: Log $X_{\text{F/Cl}}$ vs. $X(\text{Fe})$ (a), Log $X_{\text{F/Cl}}$ vs. X_{Mg} (b), P-T diagram (c), and Log $f(\text{O}_2)$ - T diagram (d): a) Log $X_{\text{F/Cl}}$ vs. X_{Fe} . The negative correlation between the F/Cl ratio and Fe is attributed to Fe-F avoidance (see text). b) Plot of Log $X(\text{F/Cl})$ vs. X_{Mg} . These are contoured for Log $f_{(\text{HF})}/f_{(\text{HCl})}$ at different temperatures of equilibration (see Table 4.4). The isopleths of equal fluorine enrichment have a positive slope. c) P-T diagram for the stability curves of end-member mica and associated minerals: a = alpha to beta quartz transition, b = hastingsite stability (at QFM), c = annite + quartz stability, and d = grunerite stability (QFM). The water-saturated granite solidus is after Tuttle & Bowen (1958), and water-undersaturated granite solidi after Ebadi & Johannes (1991). The annite + quartz stability curve intersects the water-saturated granite solidus at about 700°C and 1.5 kbar. d) Plot of Log $f(\text{O}_2)$ in bars *versus* temperature (in °C), with the effect of varying Fe/(Fe + Mg) or F/FM shown. The range of stability of the Kwandonkaya micas, based on their range of F/FM (magmatic mica: $0.85 \leq \text{F/FM} \leq 1.0$, and secondary mica: $0.76 \leq \text{F/FM} \leq 1.0$) at 1 kbar is shown to be less than 650°C. Bt = biotite, Hem = hematite, Kfs = kalsilite, Lc = leucite, Mag = magnetite, Ol = olivine, and V = vapour/gas. The experimental database pertains to end-member mica in the series annite-phlogopite.

Symbols are the same as in Fig. 4.5a.



accompanying fractional crystallization. Furthermore, the F/FM of the dark mica decreases slightly with increasing ^{IV}Al (Fig. 4.5b). In this diagram, the dark mica in albitized biotite granite overlaps with that of pristine samples of biotite granite. The significance of this overlap is addressed below.

2) The observed increase in F/Cl content of the dark mica from subvolcanic to plutonic rocks resulted from substitutions of the type $F(Cl,OH)_{-1}$. This accompanied the build-up of F relative to Cl in the volatiles during crystallization.

3) The trilithionite, in pale zones of mica in drusy biotite granite, typically has enhanced Si, ΣAl (and ^{IV}Al), Li, and F/Cl, and lower Fe, Ti (especially ^{VI}Ti), ^{IV}Al , and Cl compared to the darker zones of dark mica. This mica formed by pseudomorphic replacement of the dark mica.

Secondary mica

The secondary micas include annitic biotite in rhyolite porphyry, ferroprotolithionite - zinnwaldite - siderophyllite in albitized biotite granite and greisens, as well as trilithionite in both pristine and altered biotite granite.

The secondary mica in the rhyolite porphyry ($Srp_{0.6}$) is the most strongly annitic ($0.87 \leq F/FM \leq 0.93$) of all the subvolcanic rocks (Fig. 4.5a). On the basis of petrographic observations, this mica replaces hedenbergite.

In the albitized biotite granite, the secondary mica is ferroprotolithionite-zinnwaldite ($0.95 \leq F/FM \leq 0.99$), which is similar to that of the less altered biotite granite (Table 4.3). This similarity may be due to the fact that the major constituents of the secondary mica were not strongly modified from the mica in primary assemblages. Relatively Mg-rich mica ($0.80 \leq F/FM \leq 0.93$) also occurs in a drusy albitized biotite granite in the Jubga-Ziem center (Table 4.3). This mica is relatively Al-poor and Mg-rich compared to other occurrences of the dark mica in this unit.

Within greisens, the pale brown to greenish ferroprotolithionite-siderophyllite type ($F/FM = 1$) is more aluminous than the wallrock mica. The F/Cl ratio (43) is relatively low, possibly because F partitioned preferentially into the coexisting topaz. The composition of this type of mica seems to have been only slightly modified from that in the host rock (Fig. 4.5a), on the basis of the slight differences in their major constituents. This similarity is possibly due to a low fluid/rock ratio.

Trilithionite occurs as a subsolidus replacement in textural variants of biotite granite (as blotches or patches in dark mica: Fig. 4.3b, 4.4b), in pale zones of mica in drusy biotite granite (Fig. 4.4a), and as a hydrothermal mica in albitized biotite granite and greisens (Fig. 4.4d). The trilithionite typically has a higher proportion of Si, ^{VI}Al (and ΣAl), Na, and F, and a lower proportion of Fe, Ti, Mn, Mg, ^{IV}Al and Cl relative to the dark mica (Figs. 4.5 - 4.6). This type of mica contains the highest inferred Li (2 to 4.1 wt.% Li_2O), and plots in the field of zinnwaldite after Foster (1960, Fig. 4.6b), close to the "mixed form" (i.e., mixed dioctahedral and trioctahedral) mica. Since the Li content is only inferred, the term trilithionite is used on the basis of the classification scheme in Figure 4.5a.

The trilithionite in the albitized biotite granite ($0.86 \leq F/FM \leq 0.98$) is similar to that in the drusy biotite granite (Fig. 4.5b). The lowest F/FM of the trilithionite occurs in a rock type bearing only very pale mica with hematite lining cleavages in the mica. The hematite presumably resulted from the oxidation of Fe^{2+} in the sheet silicate.

There are three types of pale mica (trilithionite) in the greisens. The first type is relatively Fe-rich (Ann_{17-24} and $F/FM = 1$), the most Ti-rich of pale micas in greisen ($^{VI}Ti = 1.0$ apfu), relatively Al-poor (Srp_{75-82}), and has intermediate F/Cl ratio ($30 \leq F/Cl \leq 65$). Such a Ti-rich pale mica has not been previously documented in the Kwadonkaya complex. This mica may have formed by hydrothermal deposition rather than replacement. The second type of pale mica is

relatively Fe-poor (Ann_{4-10} , $0.74 \leq \text{F/FM} \leq 0.97$), Al-rich (Srp_{87-94}) compared to the first type, and has the least Ti content ($0 \leq {}^{\text{VI}}\text{Ti} \leq 0.02$ apfu), and the highest F/Cl ratio ($115 \leq \text{F/Cl} \leq 1487$) of the greisen micas. This second type of mica is significantly different from the unaltered (dark) mica in the host rock and might have been formed by the replacement of feldspars under higher fluid/rock ratio. The third type of pale mica in greisens, trilithionite, is similar to the preceding one in terms of its Fe content (Ann_{3-10}), but is slightly more Al-rich (Srp_{90-97}), and has a wider range of Ti ($0 \leq {}^{\text{VI}}\text{Ti} \leq 0.27$ apfu) and the lowest F/Cl ratio of the greisen trilithionite ($8 \leq \text{F/Cl} \leq 15.9$). This type of mica is secondary, and possibly formed by replacement of the second type. Whereas the second and third types of mica commonly coexist, the first type is rare and anomalous. The rarity of this mica might be because the greisenizing fluids invariably underwent oxidation, accompanying cassiterite deposition, to result in the formation of predominantly low F/FM mica. Thus, the preservation of this sort of relatively Fe- and Ti-rich pale mica is not commonly expected.

In summary, during postmagmatic equilibration and hydrothermal alteration, the trilithionite formed from the dark mica by pseudomorphic replacement (annite $\rightarrow$ siderophyllite $\rightarrow$ trilithionite) via substitutions of the type i) ${}^{\text{VI}}(\text{Al, Li, } \square)({}^{\text{VI}}\text{R}^{2+})_{.1}$ (where $\square$ = vacancies in the octahedral site, and $\text{R}^{2+} = \text{Fe}^{2+}, \text{Ti}$), and ii) $\text{F}(\text{OH, Cl})_{.1}$.

4.5.3 Halogen content and Fe-F avoidance

Halogen contents of mica in the subvolcanic rocks are such that the proportion of Cl is invariably greater than F, except for mica in biotite granite porphyry, a few examples of fayalite syenogranite and "hornblende"-biotite granite porphyry, in which the proportion of F is slightly greater than that of Cl (Fig. 4.6a). In the fayalite syenogranite, the low F/Cl (0.13 to 1.35) may be due to formation of the mica from a Cl-rich and F-poor fluid phase, in agreement with its

characterization as being of subsolidus-hydrothermal origin (i.e., replacing fayalite + micropertthite). In the "hornblende"-biotite granite porphyry, the $(F/Cl)^{Bt}$ (between 0 and 1.1) and the $(F/Cl)^{Hbl}$ (between 0.7 and 2.4) yield a $K_D^{(F-Cl)}$ (biotite-hornblende) [i.e., $(X_F^{Bt}/X_{Cl}^{Bt})/(X_F^{Hbl}/X_{Cl}^{Hbl})$] in the range 0.03 to 0.41. Such a low value supports the inference made earlier that biotite crystallized later than the amphibole, because, according to Anfigolov et al. (1977), the $K_D^{(F-Cl)}$ (biotite-hornblende) should be close to 1.2 for an equilibrium pair. In general, the mica in the biotite granite porphyry is the only case in subvolcanic rocks with F greater than Cl, and marks the crossover of compositions from subvolcanic rocks ($Cl > F$).

In the plutonic rocks, F is enriched over Cl in both dark and pale micas. The drusy biotite granite on the Ziem ridge and west of Jubga village contains the highest F content (Fig. 4.6a). These drusy varieties are characterized by topaz, and in some instances "disseminated" topaz-cassiterite mineralization. Although fluorite is ubiquitous, in close association with mica of different F-content, it is absent in some rock units with high-F mica. Thus, the occurrence of fluorite is not an index of fluorine enrichment. The pale mica is most depleted in chlorine, and consequently has a very high F/Cl ratio (Fig. 4.6a), although its fluorine content is similar to that of the coexisting dark mica.

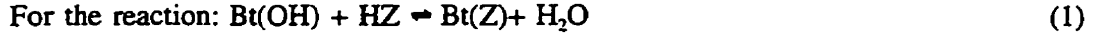
The role of **Fe-F avoidance** is clearly demonstrated in Figure 4.6a by the decreasing F/Cl ratio with increasing X_{Fe} of the dark mica (cf. Ramberg 1952, Munoz & Ludington 1974, Munoz & Swensen 1981, Munoz 1984, Mason 1992). The Fe-F avoidance (Fig. 4.6a) has been explained by thermodynamic factors (Ramberg 1952), and crystal-field theory (Rosenberg & Foit 1977). According to Ramberg (1952), bonds between Mg and F are stronger than those between Fe and F because the reaction: $Mg(OH)_2 + FeF_2 = MgF_2 + Fe(OH)_2$ is exothermic ($\Delta H = 116.7$ kcal/mol). Furthermore, according to Wagman et al. (1982), the equilibrium constant for the same reaction ($\log K = 9.57$ at 298 K) is relatively large. Therefore, Mg-F bonds are more likely to form than

bonds between Fe and F. It has also been shown that the value of crystal-field splitting parameter (Δ_o) of $[\text{Fe}(\text{OH})_6]^{4+}$ is 1000 cm^{-1} , whereas that of $[\text{FeF}_6]^{4+}$ is 7600 cm^{-1} (Rosenberg & Foit 1977). Thus, octahedrally coordinated $[\text{FeF}_6]^{4+}$ will have a smaller crystal-field stabilization energy (CFSE) than the OH-equivalent, implying that the presence of Fe will not favour incorporation of F in Fe-rich micas.

Mason (1992) has provided a detailed examination of mechanisms by which structural adjustments are made so as to accommodate F in Fe-rich micas (Appendix IV.5a, b). These mechanisms lead to violations of the ideal-mixing models of Fe and Mg in the octahedral site as well as F and OH in the OH-site. According to Mason (1992), if the octahedrally coordinated cations are randomly distributed, the maximum level of F is given by $(X_F)^{\text{max}} = (X_{\text{Mg}})^3$ (Appendix IV.5b). In order for the mica to have a higher amount of F, as is the case here, it must contain domains enriched in Fe-OH and Mg-F rich, i.e., short-range order (SRO), by forming Mg_3 groups in order to comply with Fe-F avoidance (Appendix IV.5a, cf. Mason 1992). SRO controls the maximum amount of F-substitution in mica, and permits the intake of higher amounts of F by clustering, with $X_F^{\text{Max}} = X_{\text{Mg}}$ (Appendix IV.5b). Therefore, ordered biotite contains a higher level of F than disordered biotite (Mason 1992). Less than 40% of the Kwadonkaya mica compositions plot in "Fe-F avoidance violated" field in a plot of X_F against $X_{(\text{Mg}+\text{Ti}+\text{Li})}$ (Appendix IV.5b), with dark mica in the rock suite plotting in the field of "clusters". The relatively low proportion of mica (in the Kwadonkaya data-set) violating Fe-F avoidance arises not only because of formation of Mg-F "clusters", but also because of the role of Li and Ti, which also can enhance the ability of mica to accept F in its structure (Appendix IV.5b). Pale mica show a positive correlation between X_F and X_{Fe} (Fig. 4.6a). This may be due to the fact that such mica has a relatively low Fe content, and higher inferred Li than the dark mica, which has a negative correlation of Fe with F content.

4.5.4 Inferred halogen fugacities of fluids

The derivation of fugacities of halogen in fluids that equilibrated with mica is discussed by Munoz (1992). He used thermodynamic and experimental data for components within the reciprocal system OH-annite - (F, Cl)-annite $\rightleftharpoons$ OH-phlogopite - (F,Cl)-phlogopite to calculate equilibrium constants for (F,Cl) $\rightleftharpoons$ OH exchange between biotite and fluid.



where Bt(OH) and Bt(Z) are hydroxyl-bearing, and halogen-bearing micas, respectively, the equilibrium constant is:

$$K = [a_Z^{(\text{Bt})} \cdot a_{\text{H}_2\text{O}}^{(\text{fluid})}] / [a_{\text{OH}}^{(\text{Bt})} \cdot a_{\text{HZ}}^{(\text{fluid})}] \quad (2)$$

$$= [\gamma X_Z^{(\text{Bt})} \cdot a_{\text{H}_2\text{O}}^{(\text{fluid})}] / [\gamma X_{\text{OH}}^{(\text{Bt})} \cdot a_{\text{HZ}}^{(\text{fluid})}] \quad (3)$$

Where γ = activity coefficient, a is the activity, and X the mole fraction of the subscripted component within the phase given in the superscript. Choosing the standard state of H_2O and HZ , and assuming ideal mixing of F, Cl, and OH in mica, then $\gamma Z(\text{Bt})/\gamma \text{OH}(\text{Bt}) = 1$ (i.e., $a_Z^{\text{Bt}} = X_Z^{\text{Bt}}$, and $a_{\text{OH}}^{\text{Bt}} = X_{\text{OH}}^{\text{Bt}}$), and the above relation, expressed in logarithmic terms, becomes:

$$\text{Log} K = \text{Log}[X_Z^{(\text{Bt})}/X_{\text{OH}}^{(\text{Bt})}]^{(\text{Bt,ss})} + \text{Log}(f_{\text{H}_2\text{O}}/f_{\text{HZ}})^{(\text{fluid})} \quad (4)$$

The equilibrium constant expressions for OH $\rightleftharpoons$ F and OH $\rightleftharpoons$ Cl exchange after Munoz (1992) are:

$$\text{Log} K_{(\text{F-OH})} = 1000/T(2.37 + 1.1X_{\text{Mg}}) + 0.43 \quad (5)$$

$$= \text{Log}(f_{\text{H}_2\text{O}}/f_{\text{HF}}) + \text{Log}(X_{\text{F}}/X_{\text{OH}}) \quad (6)$$

$$\therefore, \text{Log}(f_{\text{H}_2\text{O}}/f_{\text{HF}}) = 1000/T(2.37 + 1.1X_{\text{Mg}}) + 0.43 - \text{Log}(X_{\text{F}}/X_{\text{OH}}) \quad (7)$$

$$\text{Also, } \text{Log} K_{(\text{Cl-OH})} = 1000/T(1.15 - 0.55X_{\text{Mg}}) + 0.68 \quad (8)$$

$$= \text{Log}(f_{\text{H}_2\text{O}}/f_{\text{HCl}}) + \text{Log}(X_{\text{Cl}}/X_{\text{OH}}) \quad (9)$$

$$\therefore, \text{Log}(f_{\text{H}_2\text{O}}/f_{\text{HCl}}) = 1000/T(1.15 - 0.55X_{\text{Mg}}) + 0.68 - \text{Log}(X_{\text{Cl}}/X_{\text{OH}}) \quad (10)$$

Combining equations (7) and (10):

$$\text{Log}(f_{\text{HF}}/f_{\text{HCl}})^{\text{fluid}} = -1000/T(1.22 + 1.65X_{\text{Mg}}) + 0.25 + \text{Log}(X_{\text{F}}/X_{\text{Cl}}) \quad (11)$$

where T in the above equations is temperature in Kelvin. The use of equation (11) avoids the problems inherent in the calculation of mole fractions of mica components (Munoz & Ludington 1974, Gunow et al. 1980, Munoz & Swensen 1981, Munoz 1984, Mora & Valley 1989, Parsons et al. 1991).

The following temperatures were chosen for the calculation of $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$. For the subvolcanic rocks, an upper value of 500°C represents the approximate temperature for catastrophic coarsening of perthite, and the inferred development of coherent perthite in phenocrysts of subvolcanic rocks and primary alkali feldspar in hypersolvus biotite granite. The lower value, 400°C, is near the temperature of orthoclase-to-microcline inversion, and applies to those subvolcanic and plutonic rocks bearing both types of K-feldspar. The range of temperature for the subvolcanic rocks is well below that proposed by Eugster and Wones (1962) for the intersection of annite + quartz curve (ca. 700°C) with the water-saturated granite minimum of Tuttle and Bowen (1958), and also well below the crest of the alkali feldspar solvus (650°C) at one kilobar (Smith & Parsons 1974). For the case of pristine biotite granite, the upper temperature, 480°C, is the highest temperature indicated for alkali metasomatism from fluid-inclusion studies of alkali metasomatized NYG rock units (Kinnaird et al. 1985), and is used because subsolidus equilibration is expected to have occurred close to this temperature. The lower temperature of 300°C is chosen because the majority of the plutonic rocks bear microcline ($N_{\text{Or}} > 0.9$), which suggests that equilibration (and Na-for-K exchange) might have continued to such low temperatures. For the case of altered samples of biotite granite, the upper temperature, 350°C, is chosen to include the temperature of alkali metasomatism and greisenization, and the lower temperature represents the lower limit for alteration in biotite granite as inferred from fluid-inclusion studies (Kinnaird et al. 1985). The $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ values at an assumed temperature of 400°C are used to compare the different rock units.

The plot $\text{Log } X_F/X_{Cl}$ vs X_{Mg} is shown in Fig. 4.6b. X_{Mg} is chosen as the abscissa because it is one of the variables in equation 11. The inferred values of $\text{Log}(f_{HF}/f_{HCl})$ are given in Table 4.2, and summarized in Table 4.4. Assuming that all the rocks equilibrated with fluid at the same temperature (ca. 400°C or 673.15 K), it is noted that the $\text{Log}(f_{HF}/f_{HCl})$ of the fluids that equilibrated with the dark mica increases in the sequence: subvolcanic rocks (-1.99) through pristine biotite granite (-0.33) to the altered biotite granite (-0.03). However, it is more realistic to infer a range of temperatures of equilibration, as the various rock units have different volumes and evolved to various extents. Furthermore, the field evidence provided by the drusy facies in both the subvolcanic and plutonic rocks of the Jubga-Ziem center indicates that volatile loss occurred at different stages of evolution. Hence variations in temperature of equilibration even on the scale of a single exposure are to be expected.

As is the case with inferred fugacities at the same temperature, the $\text{Log}(f_{HF}/f_{HCl})$ increase in the same sequence from subvolcanic rocks (-1.74 at 773.15 K), the pristine biotite granites (-0.13 at 753.15 K and could have been as low as -0.67 at 573.15 K), to the altered biotite granites (-0.18 at 623.15 K, and to as low as -0.52 at 533.15 K). In the subvolcanic rocks the biotite granite porphyry is characterized by the highest $\text{Log}(f_{HF}/f_{HCl})$ of -1.00 at 773.15 K.

The highest values are found for drusy biotite granite [sample 30A (0.55 at 673.15 K), which contains cassiterite and topaz, and sample 37 on the Ziem ridge (0.78 at 673.15 K)]. The Ziem ridge material provides the best indication of preferential loss of Cl in the plutonic rocks, with the resulting enrichment of fluorine in biotite granites relative to the subvolcanic rocks. The deeper, coarse-grained biotite granite (sample 26) has a lower fugacity ratio (-0.92 at 673.15 K) than the upper zones of the Ziem ridge, occupied the drusy biotite granite (sample 37, 0.78 at 673.15 K). The biotite granite porphyry (38) that forms a roof pendant on the Ziem ridge has lower ratios (-1.00 at 773.15 K). The devolatilization that led to the drusy biotite granite, and the

inferred preferential loss of Cl, are responsible for the very high contents of F. Such F-enrichment accounts well for the appearance of topaz in vugs and the observation that the largest proportion of alluvial topaz occurs by the Jubga-Ziem ridge on the eastern flank of the Kwandonkaya complex.

In most examples, the albitized biotite granite displays only slightly higher $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ than their more pristine equivalents (Tables 4.2, 4.4, Fig. 4.6b). However, in some samples bearing columbite, cassiterite, topaz, fluorite, and fluocerite (e.g., sample 23), the ratios are relatively high (0.23 at 623.15 K). In other examples of albitized biotite granite (e.g., sample 17) in which the pale mica constitutes the only mica, the fugacity ratios are even higher (0.51 at 533.15 K). These values are above average for altered plutonic rocks (Table 4.4), and indicates that albitization occurred at loci of high F contents relative to Cl.

The dark mica in the albitized host rock (35A) to greisens equilibrated with a fluid phase characterized by a higher fugacity ratio (-0.09 at 623.15 K) compared to the less altered host rock (sample 36A) (-0.84). The greisen samples bearing only dark mica display a wide range of values, from a low of -0.61 in sample 84 to -0.10 in 35C at the same, assumed, temperature of equilibration, which corresponds with a realistic temperature from fluid-inclusion studies of Kinnaird (1985, 1987). The greisens bearing predominantly a pale mica also display a bimodal distribution of ratios with, high (0.64) and low values (-0.41) at 623.15 K. The relatively low values in some samples of greisen are possibly due to the fact that: i) such pale mica replaced earlier Li- and F-rich mica, ii) low-temperature equilibration of mica can result in reduced F contents (Dodge & Moore 1968, Dodge et al. 1969), and iii) autometasomatic formation of fluorite (or topaz) can reduce the fluorine content of micas if the activity of calcium (or aluminum) is high.

Generally, because H_2O contents were determined by difference, the calculated Log

($f_{\text{H}_2\text{O}}/f_{\text{HF}}$) ratios are not emphasized. Possible sources of error in the calculation of halogen fugacities include chloritization and H_2O gains, departures from the ideal-solution models, as pointed out by Mueller (1972), lack of Fe^{3+} determination, the presence of other cations not included in experimental studies (e.g., Li, Ti, Mn), calculation of equilibrium constants, and the assumption that these other cations behave like Fe. There is so far no quantification of the effects of these sources of error on inferred halogen fugacities, and therefore relative rather than absolute differences in such fugacities are emphasized.

4.6 Chlorite

Only limited analyses were made of chlorite replacing biotite. Chlorite in the rhyolite porphyry has a composition typical of chamosite. Both biotite and chlorite are of replacement origin, and are both characterized by low F/Cl ratios. The chlorite in samples of albitized biotite granite also is chamosite, and appears to have formed at the expense of the dark mica by depletion of Si, Al, Ti, F, and F, with slight enrichment of Fe and Cl. The most striking occurrence of chlorite is in the outer zones of post-joint greisens. There, a rare pale chlorite relatively depleted in Si, Ti, K, F and Cl and enriched in Fe coexists with the dark mica.

4.7 Discussion

The crystallization of mica has been established from petrographic studies to be late-magmatic to subsolidus in both subvolcanic and plutonic rocks, and is briefly discussed first. The variations in the major compositional features of mica (Tables 4.2, 3) are discussed next. These variations lead to comparisons with occurrences in similar subvolcanic-plutonic complexes. Also discussed are halogen contents and T - $f(\text{O}_2)$ conditions of crystallization in relation to the solidus.

The occurrence of annitic biotite, quartz, orthoclase microperthite and fayalite in fine-

grained syenogranite indicates the preservation of reactants and products in the reaction fayalite + sanidine_{ss} $\rightleftharpoons$ annitic biotite + quartz (Eugster & Wones 1962), as supported by petrography. The coexistence of hornblende and biotite in the hornblende biotite granite porphyry could represent an equilibrium of the type: liquid + hornblende $\rightleftharpoons$ biotite, proposed for a similar association of ferroedenite + biotite in the Liberty Hill pluton (Speer 1987).

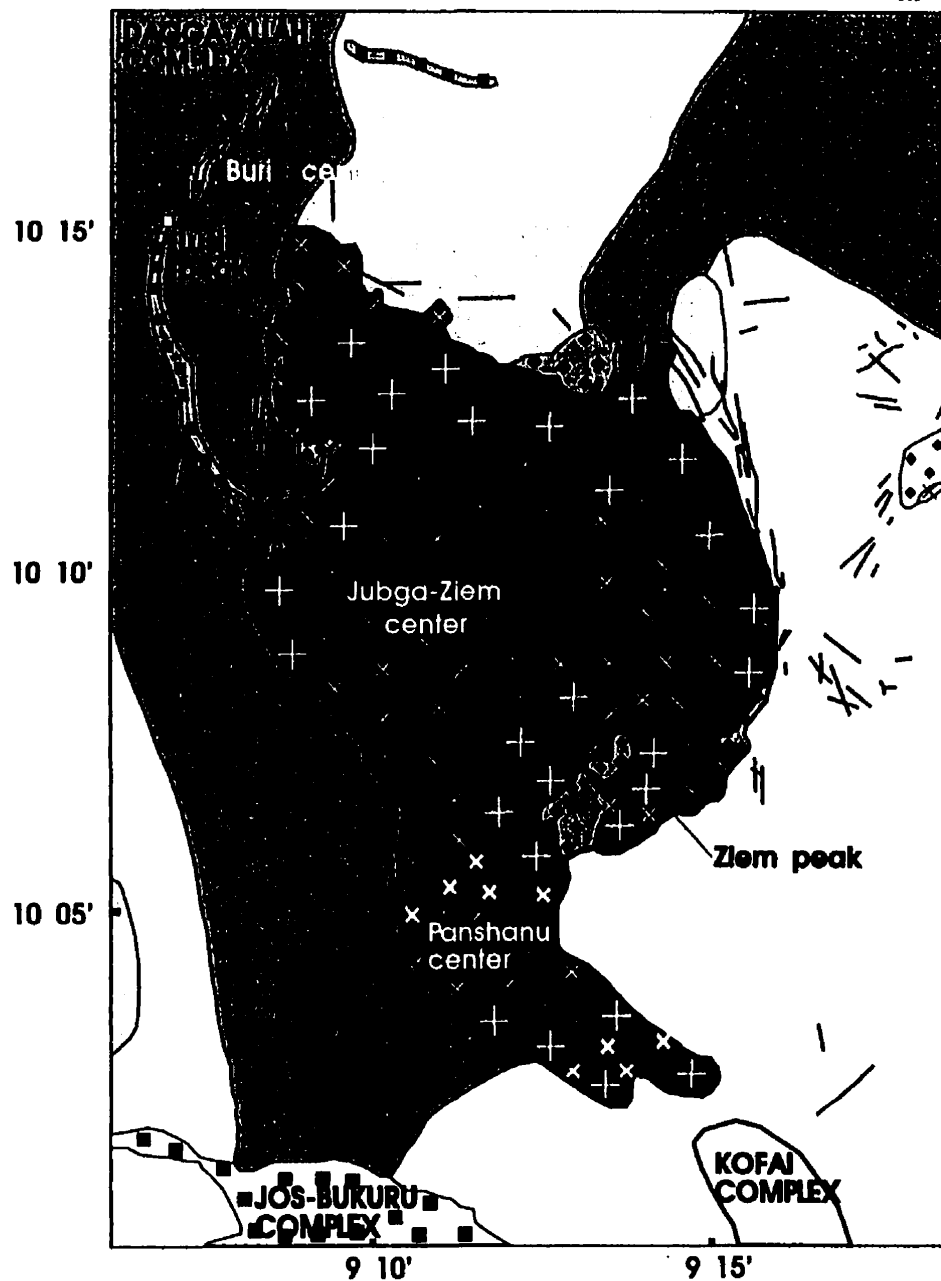
The dark mica in biotite granites represents a late-magmatic to subsolidus product of crystallization within the stability field of annite + quartz (by the reactions): i) Interstitial melt $\rightleftharpoons$ Biotite + Fe-Ti oxides, at late magmatic stages, and ii) $2\text{KAlSi}_3\text{O}_8 + 3(\text{Fe,Mg,Mn})_2\text{SiO}_4 + 2\text{H}_2\text{O} \rightleftharpoons 2\text{K}(\text{Fe,Mg,Mn})_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2 + 3\text{SiO}_2$, iii) $2\text{KAlSi}_3\text{O}_8 + 3\text{Fe}_3\text{O}_4 + 2\text{H}_2\text{O} \rightleftharpoons 2\text{KFe}_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2 + \text{O}_2$. Such crystallization occurred outside the stability field of amphiboles (cf. Eugster & Wones 1962, Speer 1984, 1987, Parsons 1981). The increase in the A/CNK [i.e., molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$] of the bulk rock, from metaluminous and mildly peraluminous subvolcanic rocks, to more strongly peraluminous plutonic biotite granites, favoured the crystallization of biotite over amphibole (or clinopyroxene).

4.7.1 Late-magmatic to subsolidus mica

The distribution of mica types and variations in relative $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ in the Kwandonkaya complex are summarized in Tables 4.3 and 4.4. In Figure 4.7, it is shown that Fe-rich micas, characterized by low $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$, occur predominantly in subvolcanic rocks, whereas micas characterized by lower Fe contents, and enhanced Al, Li (inferred), and $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ occurs in biotite granites. The increasing $^{\text{VI}}\text{Al}$ (and ΣAl) in mica with decreasing Fe (Fig. 4.5a) from that in the subvolcanic rocks to that in the plutonic rocks, reflects the fact that: i) six-fold coordination of Al is preferred in silicate minerals formed at low temperatures (cf. Thompson 1947), and ii) the activity of Al (cf. De Albuquerque 1973) or peraluminosity of the host rock (cf. Speer 1984,

Fig. 4.7: The distribution of mica types and general variations in inferred $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ of major rock units. Geological units are the same as in Fig. 4.1.

Fe-rich micas characterized by low $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ occur predominantly in subvolcanic rocks, whereas micas characterized by lower Fe contents, and enhanced Al, Li, and $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ characterize biotite granites. The localized occurrences of the albitized and drusy variants of biotite granite are characterized by even higher Al, Li, and $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$.



Basalt dyke

MESOZOIC GRANITES

M = Fp, Srp, Znw, Tl (rare Ann-Bt); moderate to high $f(\text{HF})/f(\text{HCl})$

G: Greisens

Fine- to medium-grained biotite granite

Coarse-grained biotite granite

Porphyritic biotite microgranite

M = Ann-Bt; low $f(\text{HF})/f(\text{HCl})$

"Hornblende"-biotite, biotite granite porphyry, +/- fayalite syenogranite

Hedenbergite fayalite granite porphyry

Rhyolite porphyry

Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

Biotite hornblende granite

Gneissic granite

Migmatite

1 0 2 4 6 8 km

Lalonde & Bernard, 1993) is greater in the biotite granites. In fact, the biotite hosts the excess Al in these rocks. The slight decrease in ^{IV}Al from mica in subvolcanic rocks to that in plutonic rocks (Fig. 4.5b) compares with the trends documented in the East Kemptville topaz muscovite leucogranite (Kontak 1991) and the Finnmarka complex (Czamanske & Wones 1973), interpreted to reflect increasing activity of Al in the latest fraction of magma (cf. Czamanske & Wones 1973). The overlaps in annite content of the dark mica in subvolcanic rocks (Ann_{38-94}) and least altered plutonic rocks (Ann_{17-80}), and the overlapping major and trace element trends in the whole rocks (Chapter 2), imply that mica crystallized from melts that were comagmatic. On the other hand, the overlaps in Mg content of the dark mica in the subvolcanic (Phl_{0-10}) and plutonic (Phl_{0-14}) rocks, is possibly due to similarities in $f(O_2)$ -T conditions during crystallization, and types of postmagmatic modification.

The decrease in Ti in the mica from subvolcanic rocks to plutonic rocks is largely the expression of bulk-rock variations (Fig. 4.5c). The negative correlation between Ti and ^{VI}Al in mica at Kwandonkaya arises because higher temperatures favour the entry of Ti into the biotite structure (De Albuquerque 1973). The trend in Ti content indicate substitutions of the type: $^{VI}(Al^{3+})^{IV}(Si^{4+})^{VI}(Ti^{4+})_{-1}^{IV}(Al^{3+})_{-1}$, $^{VI}(R^{2+})_2^{VI}(Ti^{4+})_{-1}^{VI}(\dagger)_{-1}$ (where $R^{2+} = Fe^{2+}, Mg^{2+}$ and $\dagger =$ vacancy), $^{VI}(R^{2+})^{IV}(Si^{4+})^{VI}(Ti^{4+})_{-1}^{IV}(Al^{3+})_{-2}$, and $^{VI}(Al^{3+})^{VI}(Li^{+})^{IV}(Ti^{4+})_{-1}$ and, most likely, also suggest the increasing role of fractionation of ilmenite and the concomitant decrease in the amount of Ti accommodated by biotite ($\pm$ hedenbergite, amphibole) with evolution of rock units, from subvolcanic to plutonic setting, in response to changing T-f(O_2) conditions. Ti, a high-field-strength element presumed to be immobile, is seen to have been depleted in the mica during alkali and acid-metasomatism. This may lend credence to the proposal that Sn (and other HFS) were mobilized during hydrothermal alteration of late-magmatic mica and subsequently deposited in mineralized zones. However, complementary evidence is required from studies using other

techniques before definite inferences can be made.

The inferred content of Li in micas increases from annitic mica in subvolcanic rocks to protolithionite-zinnwaldite in pristine plutonic rocks, and reflects a build-up in Li from subvolcanic to plutonic rocks (Fig. 4.5d). This suggests a substitution of the type $(^{\text{VI}}\text{Li}^{\text{IV}}\text{Si})(^{\text{VI}}\text{Fe}^{2+}, ^{\text{IV}}\text{Al}_{1-x})$ and $^{\text{VI}}(\text{Al}^{3+}, \text{Li}^+)^{\text{IV}}(\text{Ti}^{4+})_{1-x}$. The range of Li_2O inferred for the mica in this suite (0.3 - 4.1 wt.%) is greater than the range (of 0.8 to 1.9 wt.%) previously documented by Jacobson et al. (1958) in micas of the NYG. The pale mica in drusy samples of biotite granite is the most Li-rich, and suggests that degassing occurred in a Li-rich environment (cf. Bowden et al. 1984). Li is known to reduce melt viscosity (Manning & Pichavant 1988) and thus facilitate the escape of bubbles.

4.7.2 Secondary mica

The annitic mica in rhyolite porphyry, which is considered secondary, and ferroprotolithionite-zinnwaldite in albitized biotite granite have levels of ΣAl , $^{\text{IV}}\text{Al}$, $^{\text{VI}}\text{Al}$ and Fe similar to those of the primary mica in the more pristine rock-units (Figs. 4.5a - d). The composition of the secondary dark mica in these rocks was likely inherited from the precursor as result of low fluid/rock ratios. The dark mica in albitized biotite granite has a more restricted range of F/FM values (mostly between 0.93 to 0.96) than the pristine biotite granite. This possibly reflects the low Fe activity of the fluids, which is likely from the generally low whole-rock ΣFeO in albitized rocks (Chapter 2). The presence of two types of mica in the greisens, namely, ferroprotolithionite-siderophyllite (F/FM = 1.0), and zinnwaldite-trilithionite (F/FM = 0.88), suggests two types of greisenizing fluids: an earlier relatively reduced fluid to produce Fe-rich ferroprotolithionite-siderophyllite, and a later oxidizing fluid that produced the zinnwaldite-trilithionite along with hematite. This is in concert with the model of formation of cassiterite

deposits by oxidation of Sn^{2+} to Sn^{4+} and accompanying deposition of SnO_2 (Jackson & Helgeson 1985, Manning & Pichavant 1988). The greisens with a low F/FM mica typically develop a drusy, quartz-rich inner zone and further suggest that the mica in zones of intense alteration (higher fluid/rock ratio), which experienced mass loss, as indicated by the development of druses, is Fe-depleted, whereas mica in relatively less altered rocks (low fluid/rock ratio) has a higher F/FM. Thus the variation in F/FM in hydrothermally altered rocks may also be due to mobilization of iron by hydrothermal fluids (cf. Dodge & Moore 1968, Dodge et al. 1969).

The trilithionite has lower $^{\text{IV}}\text{Al}$, and higher $^{\text{VI}}\text{Al}$, than the mica in the more pristine rock units. The trilithionite also contains relatively little titanium, consistent with a low temperature of equilibration (Czamanske et al. 1981, Hewitt & Wones 1984). The relatively low Fe and high Al + Li contents of trilithionite mica in greisens indicate that mica compositions were buffered by that of the fluid, leading to an inference of relatively high fluid/rock ratio. The examples of patchy replacement of the dark mica by trilithionite (Fig. 4.3b) indicate localized intense subsolidus interaction with acidic fluids. A Li-Al mica is predominant in altered plutonic rocks because the Li-Al micas are limited to crystallization environments depleted in Fe^{2+} , Mg, and Mn^{2+} and cannot coexist in equilibrium with biotite (Monier & Robert 1986).

4.7.3 *Inferred halogen fugacities*

This study has shown that the ratio $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$ increases from subvolcanic to plutonic rocks (Figs. 4.6b, 4.7), and zones of degassing as indicated by drusy variants of the rock suite are characterized by the highest fugacity ratios. The enrichment in F, along with Li (also inferred from mica chemistry and illustrated in Fig. 4.5d), has important effects on the nature of felsic melts, as indicated by Manning & Pichavant (1988). The effects include, among others, displacing the granite minima toward an albite-rich composition, lowering viscosity and enabling magma to

degas more peacefully (i.e., without explosive activity), and causing the melt to persist to a late-stage (lower temperature) of magmatic evolution, thereby enabling further enrichment of incompatible elements (+ ore metals). The observed close association of F (+ Li) enrichments and Sn-W-Nb mineralization reflects the stabilization of various metal complexes, favoured by the depolymerized nature of such melts. Several authors (Munoz & Swensen 1981, Manning & Pichavant 1988, Mora & Valley 1989, Webster & Holloway 1990, van Middelaaar & Keith 1990) have shown that Cl partitions strongly in favour of the fluid phase ($0.016 \leq K_D^{\text{melt/fluid}} \leq 0.021$), whereas F prefers the melt ($3 \leq K_D^{\text{melt/fluid}} \leq 8$). This observation has an important bearing on ore-forming processes (including alterations) because Sn, known to occur as important deposits in NYG biotite granites, partitions in favour of Cl-rich aqueous phase (cf. Manning & Pichavant 1988). Although no fluid inclusion studies have been undertaken in the Kwadonkaya complex, Kinnaird (1985, 1987) in her fluid inclusion studies of several NYG complexes, has consistently shown that an NaCl-rich fluid was very important in the deposition of ore. The inferred increase in f_{HCl} with decreasing temperature suggests that low-temperature residual fluids could have the capacity to transport Sn, W, and Nb (cf. Manning & Pichavant 1988).

4.7.4 Inferred T - $f(\text{O}_2)$ conditions of crystallization.

The absence of biotite as a magmatic phase in the Buri ring dyke is attributed to the low $a(\text{H}_2\text{O})$ of the melt ($0.3 \leq a(\text{H}_2\text{O}) \leq 0.5$). In plutonic rocks, where fluorine enrichment is inferred from the mica's composition (Fig. 4.6b), the granite solidus is expected to have been depressed to perhaps less than 650° (Manning 1981), and the biotite stability field is also expanded owing to the stabilizing influence of F (Munoz & Ludington 1974, Munoz 1984) thereby intersecting the granite solidus at lower P . In the F-free system, the assemblage of annite + quartz + vapour is known to be unstable at temperatures above $705 \pm 5^\circ\text{C}$ (at 2.0 kbars) as compared to 825°C for

annite alone (Wones & Eugster 1965, Wones 1972). The annite + quartz bulk composition is stable up to $685 \pm 5^\circ\text{C}$ at about 1.0 kbar (ca. QFM), and the stability curve of annite + quartz intersects the water-saturated granite minimum of Tuttle & Bowen (1958) at 700°C and 1.5 kbar (Fig. 4.6c). Since the crystallization of mica in the Kwandonkaya complex postdates the formation of alkali feldspar in all rock units, it appears that a combination of F, Al, and Li enrichment resulted in mica crystallization from the residual melt below 650°C in both the subvolcanic and plutonic rocks (i.e., below the crest of the alkali feldspar solvus, as determined by Smith & Parsons 1974).

The range of F/FM of mica in the Kwandonkaya complex (0.8 - 1) is similar to that documented for biotite in anorogenic granites devoid of magnetite, formed at low $f(\text{O}_2)$, and subject to minimal hydrothermal alteration (Anderson et al. 1980). The late-magmatic dark mica in subvolcanic rocks and textural variants of biotite granite has a relatively high value of F/FM ($0.96 \leq \text{F/FM} \leq 1$). Such high F/FM values suggests low $f(\text{O}_2)$ conditions at the stage of mica crystallization in these rock types (cf. Czamanske & Wones 1973, Anderson et al. 1980). This inference is possible because the F/FM of biotite approaches that of the host rock either i) in the absence of magnetite (before and during biotite crystallization), or ii) at low $f(\text{O}_2)$ (ca. QFM) in cases where magnetite is present (Anderson et al. 1980). The variation of the average F/FM from 0.99 in the fayalite syenogranite to 0.90 in "hornblende"-biotite granite is consistent with different degrees of oxidation of the residual magma in the Jubga-Ziem ring dyke. Furthermore, the overall decrease in F/FM of mica with evolution from subvolcanic rock to plutonic rocks in this study (Table, 4.3, Fig. 4.5b) reflects higher levels of $f(\text{O}_2)$ in the plutonic rocks, generalized by the reaction (after Wones & Eugster 1965): $2\text{K}(\text{Fe}_2\text{Mg})_{\Sigma 3}\text{AlSi}_3\text{O}_{10}(\text{OH})_2 + \text{O}_2 \rightarrow \text{K}(\text{Mg}_2\text{Fe})_{\Sigma 3}\text{AlSi}_3\text{O}_{10}(\text{OH})_2 + \text{KAlSi}_3\text{O}_8 + \text{Fe}_3\text{O}_4 + 2\text{H}_2\text{O}$. Therefore, biotite becomes increasingly Mg-rich in an oxidizing trend, with falling temperature or progressive crystallization. In a reducing

trend, biotite would be expected to become increasingly Fe-rich (Wones & Eugster 1965).

The inferred oxidation trend, from the variations in F/FM of mica in the Kwandonkaya complex, is similar to observations in the Finnmarka complex (Czamanske & Wones 1973), and Baie-des-Moutons complex (Lalonde & Martin 1983). Oxidation with consequent Fe^{2+} -depletion can occur either at magmatic stages by i): loss of protons following degassing from magma, or a mush of crystals + melt (Czamanske & Wones 1973, Carmichael et al. 1974), ii) the interaction of magmas with atmospheric O_2 during cauldron subsidence, when gases are proposed to stream into a partly evacuated magma chamber (Haslam 1968), and iii) increases in $f(\text{O}_2)$ due to crystallization-induced build-up in water. The degassing needed to produce drusy rock units in the Kwandonkaya complex seems to have occurred from an interstitial, volatile-rich melt and might have continued below the solidus (owing to large undercoolings associated with the volatile loss). Previous work by Imeokparia (1984) indicates that the mica in the Kwandonkaya complex crystallized at $f(\text{O}_2)$ conditions slightly above the QFM buffer. From the range of F/FM values (magmatic mica: $0.9 \leq \text{F/FM} \leq 1$, and secondary mica: $0.74 \leq \text{F/FM} \leq 1$), one can infer, on the basis of $f(\text{O}_2)$ -T space contoured for F/FM for the "pure" system by Eugster & Wones (1962) and shown in Figure 4.6d, that the range of $f(\text{O}_2)$ is between 10^{-21} to 10^{-17} bars, for crystallization temperatures between 450 and 650°C, and between QFM and HM oxygen buffers.

4.8 Conclusions

1) The dark mica in subvolcanic and plutonic rocks mostly crystallized mainly from an interstitial residual melt enriched in volatile components. On the other hand, in the albitized biotite granite and greisens, the dark mica formed either by re-equilibration of pre-existing late-magmatic mica with metasomatizing fluids, by replacement of alkali feldspar, or it was deposited directly from hydrothermal fluids.

2) The variations in the dark mica chemistry include progressive increases in Si, $^{\text{VI}}\text{Al}$ (and ΣAl), Li and F/Cl with decreasing Fe^{2+} and $^{\text{IV}}\text{Al}$. Such variations in mica compositions (from annitic biotite to ferroprotolithionite - siderophyllite - zinnwaldite) suggest an evolution from subvolcanic to plutonic rocks by efficient fractional crystallization.

3) The range of F/FM (0.80 to 1) reflects the highly evolved composition of the suite of rocks, whereas the irregular variations in the F/FM of mica, even in single intrusive units, are due to an interplay of factors that include: the degree of oxidation of the residual melt in the subvolcanic rocks and the more pristine plutonic rocks, whether or not ilmenite and magnetite crystallized before, with, or after the mica, and variations in the degree of post-magmatic alteration in the case of albitized and greisenized rocks.

4) The inferred halogen fugacity ratios [$\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$] increase in the sequence: subvolcanic rocks through pristine plutonic rocks to altered biotite granites at different temperatures of equilibration. The variation in the halogen fugacities in the various rock-units reflects a dropping temperature of equilibration with fluids evolved to different extents, the inherent variation in mica chemistry (particularly the X_{Fe} or X_{Mg}) that affects the degree of accommodation of the halogens, and various extents of volatile loss. The dark mica composition in drusy as well as albitized biotite granite indicates late re-equilibration with F-rich fluids. Zoning in mica is rare and is observed mainly in drusy fine-grained biotite granite, suggesting a relationship between volatile loss and growth-zonation of the mica.

5) Increasing f_{HCl} with decreasing temperature for a given mica composition is in agreement with documented observations, from experimental studies, that Cl partitions more strongly into an aqueous fluid relative to F-bearing granitic melts.

6) Whereas the pale-coloured mica in drusy granites is considered late-magmatic, the pale-coloured trilithionite in other biotite granites, albitized biotite granite and greisens is entirely

secondary. In both cases, the pale mica is predicted to be a mixed structure (i.e., an intermediate composition between dioctahedral lithian muscovite and trioctahedral lepidolite), generally enriched in Li, Al, and Si, and inferred to have equilibrated with fluids highly enriched in F relative to Cl.

7) In general, the documented variations in the composition of mica are due to an interplay of related factors that include bulk-rock chemical variations in the rock units bearing mica (e.g., different degrees of evolution of the residual melt from which the mica crystallized), variations in $f(\text{O}_2)$ and temperature at magmatic and postmagmatic stages, and variations in the degree of volatile build-up and loss.

8) In these respects, the Kwadonkaya complex seems unusual among the NYG complexes in showing a positive correlation between the effects of high fluid-rock ratios (in localized zones) and high fugacity ratios [i.e., $\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$].

Table 4.2: Average composition of mica; **n** = number of analyses averaged, # water calculated by difference, and * Li₂O inferred according to the formula: Li₂O (wt.%) = (0.287 x SiO₂) - 9.552 (after Tindle & Webb 1990). Mole fraction of mica constituents after Gunow et al. (1980): i) the mole fraction of phlogopite ($X_{\text{Phl}} = X_{\text{Mg}} = \text{Mg} / \sum Y$ site, where $\sum Y = \text{VI}(\text{Al} + \text{Ti} + \text{Fe}^{3+} + \text{Fe}^{2+} + \text{Mg} + \text{Mn} + \text{Li})$); ii) mole fraction of siderophyllite ($X_{\text{Spr}} = [(3 - (\text{Si}/\text{Al})/1.75)](1 - X_{\text{Phl}})$); and iii) mole fraction of annite ($X_{\text{Ann}} = 1 - (X_{\text{Phl}} + X_{\text{Spr}})$).

Pristine rock units: Dark Mica in: I rhyolite porphyry, II fayalite syenogranite, III "hornblende"-biotite granite porphyry, IV biotite granite porphyry, V coarse grained biotite granite, VI hypersolvus fine-grained biotite granite, VII Fine- to medium-grained biotite granite, VIII drusy biotite granite (30A = cassiterite- and topaz-bearing; 37 is drusy hypersolvus granite on Ziem ridge with topaz in druses), IX biotite granite in Panshanu center (1A, B: porphyritic biotite granite, and 2A, B: fine- to medium-grained biotite granite). **Pale mica in:** X = Pale mica in pristine biotite granite.

Altered rock units: Dark mica in: XI albitized biotite granite, XII = host rock of greisens, XIII greisens. **Pale mica in:** XIV altered rock units (17, 22C = albitized biotite granite), and XV greisens.

The data presented in these Tables are average results of electron-microprobe analyses (see text).

	II Fa syenogranite					III Hbl Bt granite porphyry				
	I 49A	47A	47B1	47B2	58	51	55	56	68A	68B
n	2	5	3	3	1	1	9	6	5	3
SiO ₂	36.47	34.97	34.37	33.99	34.26	35.02	33.50	33.56	32.91	32.90
TiO ₂	2.62	2.54	2.64	3.03	3.52	1.44	2.25	2.82	3.25	3.27
Al ₂ O ₃	10.72	12.05	12.35	11.50	12.18	12.51	11.89	11.83	12.09	12.56
FeO	33.23	36.33	35.96	37.15	35.03	36.87	36.68	35.91	36.31	36.09
MnO	0.23	0.28	0.28	0.28	0.57	0.63	0.53	0.48	0.48	0.46
MgO	2.82	0.98	0.14	0.10	0.89	0.05	0.62	0.93	0.06	0.09
CaO	0.06	0.00	0.03	0.04	0.00	0.01	0.07	0.13	0.04	0.07
Na ₂ O	0.05	0.04	0.03	0.04	0.10	0.04	0.09	0.04	0.04	0.03
K ₂ O	8.08	8.93	9.17	9.02	9.11	8.64	8.89	8.37	8.81	8.68
H ₂ O#	3.48	3.06	3.04	2.98	3.37	3.55	3.22	3.33	3.24	3.31
F	0.18	0.59	0.32	0.35	0.06	0.03	0.24	0.17	0.01	0.02
Cl	0.55	1.06	1.54	1.61	0.87	0.25	0.91	0.73	1.22	0.99
O=F+Cl	-0.23	-0.50	-0.50	-0.50	-0.20	-0.07	-0.30	-0.20	-0.30	-0.20
Total	98.28	100.33	99.41	99.59	99.74	98.95	98.58	98.05	98.18	98.24
Li ₂ O*	0.92	0.48	0.31	0.20	0.28	0.50	0.12	0.13	-	0.01
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.94	5.75	5.74	5.73	5.67	5.82	5.69	5.69	5.63	5.60
IVAl	2.06	2.25	2.26	2.27	2.33	2.18	2.30	2.30	2.37	2.40
IVTi	0.01	-	-	0.01	-	-	0.01	0.01	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	-	0.08	0.18	0.02	0.05	0.27	0.08	0.06	0.07	0.12
VITi	0.32	0.31	0.33	0.38	0.44	0.18	0.28	0.35	0.42	0.42
Fe ²⁺	4.52	4.99	5.03	5.24	4.85	5.13	5.21	5.09	5.19	5.14
Mn ²⁺	0.03	0.04	0.04	0.04	0.08	0.09	0.08	0.07	0.07	0.07
Mg	0.69	0.24	0.03	0.03	0.22	0.01	0.16	0.23	0.02	0.02
Li	0.60	0.32	0.21	0.14	0.19	0.33	0.08	0.09	-	0.01
Y site	6.16	5.99	5.82	5.83	5.82	6.01	5.88	5.89	5.76	5.77
Ca	0.01	0.00	0.01	0.01	0.00	0.00	0.01	0.02	0.01	0.01
Na	0.02	0.01	0.01	0.01	0.03	0.01	0.03	0.01	0.01	0.01
K	1.68	1.87	1.96	1.94	1.92	1.83	1.93	1.81	1.92	1.88
X site	1.70	1.88	1.97	1.96	1.96	1.85	1.97	1.84	1.94	1.91
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.76	3.40	3.39	3.35	3.73	3.91	3.61	3.70	3.64	3.70
F	0.09	0.31	0.17	0.19	0.03	0.02	0.13	0.09	0.01	0.01
Cl	0.15	0.29	0.44	0.46	0.24	0.07	0.26	0.21	0.35	0.28
X(F)	0.02	0.08	0.04	0.05	0.01	0.00	0.03	0.02	0.00	0.00
X(Cl)	0.04	0.07	0.11	0.11	0.06	0.02	0.07	0.05	0.09	0.07
X(F/Cl)	0.63	1.04	0.40	0.41	0.13	0.24	0.51	0.48	0.02	0.04
Log(F/Cl)	-0.20	0.00	-0.40	-0.40	-0.90	-0.60	-0.30	-0.40	-0.20	-1.00
F/FM	0.87	0.95	0.99	1.00	0.96	1.00	0.97	0.96	1.00	1.00
X(Phl)	0.11	0.04	0.01	0.00	0.04	0.00	0.03	0.04	0.00	0.00
X(Ann)	0.83	0.66	0.63	0.72	0.63	0.64	0.64	0.64	0.60	0.56
X(Srp)	0.06	0.30	0.36	0.28	0.34	0.36	0.34	0.32	0.39	0.44
Log[f _(H₂O) /f _(H₂O)] at different temperatures										
773.15 K	-1.77	-1.41	-1.79	-1.72	-2.30	-1.95	-1.73	-1.84	-1.54	-2.32
673.15 K	-2.04	-1.66	-2.03	-1.96	-2.54	-2.18	-1.97	-2.09	-1.78	-2.56

	III	IV		V Coarse-gr Bt granite				VI	VII	
	79	38	subav	26A	26B	33	39Cl	83	13B1	14
n	2	5	45	6	7	6	6	6	7	7
SiO ₂	34.17	35.27	34.28	35.51	35.22	36.19	36.72	33.97	36.57	37.05
TiO ₂	2.17	2.34	2.66	1.89	2.23	2.51	1.90	2.64	2.64	1.47
Al ₂ O ₃	11.83	14.89	12.20	15.22	14.93	15.67	18.08	11.22	14.54	18.23
FeO	33.99	31.88	35.45	31.63	30.99	29.91	25.54	36.45	29.01	26.00
MnO	0.40	0.38	0.42	0.36	0.37	0.51	0.43	0.36	0.51	0.54
MgO	2.22	1.01	0.83	2.40	2.25	1.83	1.34	0.05	2.85	0.63
CaO	0.03	0.02	0.04	0.02	0.01	0.01	0.00	0.00	0.02	0.00
Na ₂ O	0.02	0.14	0.05	0.11	0.09	0.15	0.11	0.07	0.09	0.13
K ₂ O	8.82	9.16	8.81	9.24	8.81	9.34	9.62	8.47	9.85	9.73
H ₂ O#	3.25	3.14	3.25	2.38	2.60	2.61	2.67	2.83	2.62	2.47
F	0.39	0.88	0.27	2.50	2.06	2.17	2.26	1.31	2.14	2.73
Cl	0.75	0.58	0.92	0.58	0.53	0.45	0.31	0.43	0.49	0.21
O=F+Cl	-3.00	-0.50	-0.30	-1.00	-1.00	-1.00	-1.00	-0.70	-1.00	-1.00
Total	97.70	99.18	98.85	100.65	99.10	100.32	97.96	97.15	100.34	98.00
Li ₂ O*	0.25	0.57	0.31	0.64	0.57	0.84	0.99	0.20	0.94	1.08
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.75	5.70	5.73	5.64	5.67	5.69	5.76	5.82	5.74	5.81
IVAl	2.24	2.30	2.27	2.36	2.33	2.31	2.24	2.18	2.26	2.19
IVTi	0.01	-	0.00	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.10	0.54	0.13	0.50	0.50	0.60	1.11	0.08	0.44	1.19
VITi	0.27	0.28	0.33	0.23	0.27	0.30	0.22	0.34	0.31	0.17
Fe ²⁺	4.78	4.31	4.96	4.20	4.17	3.94	3.35	5.22	3.81	3.41
Mn ²⁺	0.06	0.05	0.06	0.05	0.05	0.07	0.06	0.05	0.07	0.07
Mg	0.56	0.24	0.20	0.57	0.54	0.43	0.31	0.01	0.67	0.15
Li	0.17	0.37	0.21	0.41	0.37	0.53	0.62	0.14	0.60	0.68
Y site	5.93	5.80	5.89	5.95	5.91	5.86	5.67	5.84	5.89	5.67
Ca	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.01	0.04	0.02	0.03	0.03	0.05	0.03	0.02	0.03	0.04
K	1.89	1.89	1.88	1.87	1.81	1.87	1.92	1.85	1.97	1.95
X site	1.91	1.94	1.90	1.91	1.84	1.92	1.96	1.87	2.00	1.99
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.58	3.39	3.60	2.59	2.81	2.80	2.79	3.16	2.81	2.59
F	0.21	0.45	0.14	1.26	1.05	1.08	1.12	0.71	1.06	1.35
Cl	0.21	0.16	0.26	0.16	0.15	0.12	0.08	0.13	0.13	0.05
X(F)	0.05	0.11	0.04	0.31	0.26	0.27	0.28	0.18	0.27	0.34
X(Cl)	0.05	0.04	0.07	0.04	0.04	0.03	0.02	0.03	0.03	0.01
X(F/Cl)	0.98	2.71	0.63	8.28	7.28	9.67	13.61	6.11	8.39	27.11
Log(F/Cl)	-0.01	0.42	-0.30	0.91	0.85	0.96	1.12	0.77	0.91	1.41
F/FM	0.90	0.95	0.96	0.88	0.89	0.90	0.91	1.00	0.85	0.96
X(Phl)	0.09	0.04	0.03	0.10	0.09	0.07	0.05	0.00	0.11	0.03
X(Ann)	0.62	0.42	0.63	0.38	0.39	0.38	0.26	0.75	0.45	0.27
X(Srp)	0.28	0.54	0.33	0.52	0.51	0.55	0.69	0.25	0.44	0.71
Log[F _(M) /F _(MCL)] at different temperatures										
773.15 K	-1.54	-1.00	-1.74	-0.67	-0.72	-0.57	-0.37	-0.61	-0.71	-0.02
(753.15 K for biotite granites)										
673.15 K	-1.80	-1.24	-1.99	-0.89	-0.94	-0.78	-0.58	-0.80	-0.93	-0.22
573.15 K	n.d.	n.d.	n.d.	-1.24	-1.29	-1.13	-0.92	-1.12	-1.29	-0.54

	VII		VIII			IX Bt granite (Panshanu center)				
	29a	28B	30A	37a	37b	1A1a	1A2a	1B2a	1B2b	2A1a
n	7	6	4	5	4	10	13	8	6	9
SiO ₂	35.61	36.80	39.10	40.15	39.94	37.26	38.14	37.19	37.21	37.25
TiO ₂	2.29	2.15	0.24	0.56	0.85	1.79	1.10	2.30	2.12	1.66
Al ₂ O ₃	16.47	18.15	20.92	19.15	18.35	18.95	19.70	17.69	17.69	19.09
FeO	27.86	27.44	24.39	22.33	23.03	26.56	23.80	24.93	27.34	25.07
MnO	0.68	0.47	0.52	0.46	0.46	0.52	0.49	0.34	0.34	0.52
MgO	0.94	0.58	0.10	0.24	0.25	0.88	0.99	1.64	1.66	1.10
CaO	0.01	0.00	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01
Na ₂ O	0.12	0.16	0.14	0.16	0.13	0.12	0.17	0.10	0.10	0.15
K ₂ O	9.45	9.51	9.89	10.02	9.91	9.83	9.85	9.90	9.86	9.46
H ₂ O#	2.77	2.11	2.42	1.63	1.86	2.65	2.42	2.85	2.80	2.61
F	1.82	3.38	3.14	4.82	4.29	2.43	3.02	1.95	2.06	2.54
Cl	0.47	0.37	0.04	0.04	0.05	0.18	0.14	0.26	0.22	0.16
O=F+Cl	-0.90	-2.00	-1.00	-2.00	-2.00	-1.00	-1.00	-0.90	-0.90	-1.00
Total	97.61	99.62	99.59	97.54	97.30	100.13	98.52	98.30	100.47	98.52
Li ₂ O*	0.67	1.01	1.67	1.97	1.91	1.14	1.39	1.12	1.13	1.14
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.74	5.73	5.88	6.09	6.10	5.72	5.83	5.79	5.72	5.76
IVAl	2.26	2.27	2.12	1.91	1.90	2.28	2.17	2.21	2.28	2.24
IVTi	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.86	1.06	1.58	1.52	1.41	1.14	1.39	1.03	0.92	1.24
VTi	0.28	0.25	0.03	0.06	0.10	0.21	0.13	0.27	0.24	0.19
Fe ²⁺	3.75	3.57	3.07	2.84	2.96	3.41	3.04	3.24	3.51	3.24
Mn ²⁺	0.09	0.06	0.07	0.06	0.06	0.07	0.06	0.05	0.04	0.07
Mg	0.23	0.13	0.02	0.06	0.06	0.20	0.23	0.38	0.38	0.25
Li	0.43	0.63	1.01	1.20	1.17	0.70	0.86	0.70	0.70	0.71
Y site	5.65	5.71	5.78	5.74	5.74	5.73	5.70	5.67	5.80	5.70
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.04	0.05	0.04	0.05	0.04	0.04	0.05	0.03	0.03	0.04
K	1.94	1.89	1.90	1.94	1.93	1.92	1.92	1.97	1.93	1.87
X site	1.98	1.94	1.94	1.99	1.97	1.96	1.97	2.00	1.96	1.91
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.95	2.24	2.49	1.68	1.92	2.77	2.50	2.97	2.94	2.71
F	0.92	1.66	1.49	2.31	2.07	1.18	1.46	0.96	1.00	1.24
Cl	0.13	0.10	0.01	0.01	0.01	0.05	0.04	0.07	0.06	0.04
X(F)	0.23	0.42	0.37	0.58	0.52	0.30	0.37	0.24	0.25	0.31
X(Cl)	0.03	0.02	0.00	0.00	0.00	0.01	0.01	0.02	0.01	0.01
X(F/Cl)	7.62	18.43	136.81	261.76	320.11	25.99	41.29	14.27	17.95	31.23
Log(F/Cl)	0.87	1.25	2.12	2.40	2.33	1.40	1.61	1.14	1.25	1.48
F/FM	0.94	0.96	0.99	0.98	0.98	0.94	0.93	0.89	0.90	0.93
X(Phl)	0.04	0.02	0.00	0.01	0.01	0.04	0.04	0.07	0.07	0.04
X(Ann)	0.32	0.26	0.19	0.30	0.34	0.23	0.22	0.28	0.29	0.22
X(Srp)	0.64	0.71	0.81	0.69	0.65	0.73	0.74	0.65	0.65	0.73
Log[f _(FM) /f _(FM1)] at different temperatures										
753.15 K	-0.59	-0.17	0.75	1.01	0.94	-0.04	0.15	-0.37	-0.27	0.01
673.15 K	-0.79	-0.37	0.55	0.81	0.74	-0.24	-0.05	-0.58	-0.48	-0.20
573.15 K	-1.13	-0.70	0.24	0.49	0.42	-0.58	-0.38	-0.93	-0.82	-0.53

IX	X: Pale coloured mica in Bt granite									
	2B1a	39C2	29b	1A1b	1A2b	2A1b	2B1b	9A darkav paleav		
n	15	1	1	1	1	3	1	2	132	10
SiO ₂	37.50	44.80	47.47	52.00	46.47	46.84	47.63	48.73	37.08	47.71
TiO ₂	2.00	0.29	0.09	0.01	0.34	0.14	0.23	0.04	1.80	0.16
Al ₂ O ₃	18.30	28.02	27.92	25.08	27.51	26.71	27.21	31.80	17.35	27.75
FeO	24.77	8.37	6.82	6.72	8.56	9.10	6.61	1.56	27.06	6.82
MnO	0.50	0.16	0.55	0.03	0.16	0.18	0.19	-	0.47	0.18
MgO	1.25	0.34	0.53	1.18	0.40	0.47	0.63	0.79	1.17	0.62
CaO	0.01	-	0.03	0.34	-	0.00	-	0.07	0.01	0.06
Na ₂ O	0.13	0.23	0.13	0.02	0.14	0.12	0.20	0.05	0.12	0.13
K ₂ O	9.21	10.78	11.07	5.40	11.10	10.67	9.95	6.19	9.55	9.31
H ₂ O#	2.59	3.23	3.32	2.99	3.24	3.24	3.27	4.36	2.49	3.38
F	2.59	2.30	2.28	3.32	2.35	2.35	2.42	0.66	2.62	2.24
Cl	0.18	0.03	0.01	0.02	0.04	0.01	0.00	0.00	0.28	0.02
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.30	-1.00	-0.90
Total	97.91	97.58	99.25	95.71	99.31	98.84	97.32	93.98	98.84	97.43
Li ₂ O*	1.21	3.31	4.07	5.37	3.79	3.89	4.12	4.43	1.09	4.14
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.82	6.12	6.26	6.71	6.20	6.27	6.34	6.34	5.80	6.32
IVAl	2.18	1.88	1.74	1.29	1.80	1.73	1.66	1.66	2.20	1.68
IVTi	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	1.16	2.64	2.60	2.53	2.53	2.49	2.62	3.21	0.98	2.66
VTi	0.23	0.03	0.01	0.00	0.03	0.01	0.02	0.00	0.21	0.02
Fe ²⁺	3.21	0.96	0.75	0.73	0.96	1.02	0.74	0.17	3.55	0.76
Mn ²⁺	0.07	0.02	0.06	0.00	0.02	0.02	0.02	-	0.06	0.02
Mg	0.29	0.07	0.10	0.23	0.08	0.09	0.12	0.15	0.27	0.12
Li	0.75	1.82	2.16	2.79	2.03	2.09	2.21	2.32	0.68	2.20
Y site	5.72	5.53	5.69	6.28	5.65	5.73	5.73	5.86	5.76	5.78
Ca	0.00	-	0.00	0.05	-	0.00	-	0.01	0.00	0.01
Na	0.04	0.06	0.03	0.00	0.04	0.03	0.05	0.01	0.04	0.03
K	1.82	1.88	1.86	0.89	1.89	1.82	1.69	1.01	1.90	1.58
X site	1.86	1.94	1.90	0.94	1.93	1.85	1.74	1.03	1.94	1.62
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.68	3.00	3.05	2.64	3.00	3.00	2.98	3.73	2.63	3.06
F	1.27	1.00	0.95	1.36	0.99	1.00	1.02	0.27	1.29	0.94
Cl	0.05	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.08	0.00
X(F)	0.32	0.25	0.24	0.34	0.25	0.25	0.25	0.07	0.32	0.23
X(Cl)	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00
X(F/Cl)	30.03	143.25	708.19	326.47	125.03	1038.3	902.82	329.06	54.78	510.44
Log(F/Cl)	1.46	2.16	2.85	2.51	2.10	2.72	2.96	2.45	1.35	2.54
F/FM	0.92	0.93	0.88	0.76	0.92	0.91	0.86	0.53	0.93	0.83
X(Phl)	0.05	0.01	0.02	0.04	0.01	0.02	0.02	0.03	0.05	0.02
X(Ann)	0.27	0.06	0.11	0.28	0.10	0.13	0.13	0.03	0.32	0.12
X(Srp)	0.68	0.93	0.87	0.68	0.88	0.85	0.85	0.95	0.63	0.86
Log[f _(M) /f _(M1)] at different temperatures										
753.15 K	-0.02	0.76	1.44	1.07	0.70	1.32	1.54	1.03	-0.13	1.12
673.15 K	-0.23	0.56	1.24	0.86	0.50	1.12	1.34	0.83	-0.33	0.92
573.15 K	-0.57	0.24	0.92	0.53	0.18	0.80	1.01	0.50	-0.67	0.60

XI: Dark mica in albitized biotite granite

	1C	15	22	22A	22C1	22C2	22C3	23	24	74C
n	11	9	8	8	7	4	6	8	5	3
SiO ₂	37.14	38.71	37.35	37.55	38.42	38.54	37.27	38.25	38.73	37.64
TiO ₂	1.60	0.77	1.42	1.26	1.03	1.25	1.55	1.22	2.68	1.26
Al ₂ O ₃	18.76	19.40	19.00	19.56	19.75	19.13	18.50	20.03	14.45	19.22
FeO	26.12	24.77	25.74	24.62	23.58	24.75	26.95	23.55	23.97	24.76
MnO	0.53	0.54	0.60	0.54	0.38	0.36	0.55	0.50	0.44	0.65
MgO	0.63	0.17	0.35	0.33	0.43	0.46	0.45	0.25	3.45	0.33
CaO	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.02	0.00
Na ₂ O	0.16	0.22	0.15	0.18	0.16	0.12	0.23	0.20	0.05	0.20
K ₂ O	9.73	9.63	9.84	9.53	9.46	9.44	9.40	9.64	9.82	9.56
H ₂ O#	2.28	1.83	2.19	2.21	2.23	2.40	1.99	1.86	2.34	2.18
F	3.13	4.15	3.32	3.33	3.37	3.00	3.72	4.13	3.01	3.69
Cl	0.14	0.10	0.13	0.10	0.09	0.09	0.14	0.08	0.16	0.08
O=F+Cl	-1.00	-2.00	-1.00	-1.00	-1.00	-1.00	-2.00	-2.00	-1.00	-2.00
Total	98.88	98.55	98.67	97.78	97.47	98.26	99.15	97.97	97.82	98.00
Li ₂ O*	1.11	1.56	1.17	1.23	1.47	1.51	1.14	1.43	1.56	1.25
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.77	5.93	5.80	5.83	5.91	5.91	5.79	5.87	6.00	4.56
IVAl	2.23	2.07	2.20	2.17	2.09	2.09	2.21	2.13	2.00	1.70
IVTi	-	-	-	-	-	-	-	-	-	0.01
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	6.27
VIAI	1.21	1.44	1.28	1.41	1.49	1.37	1.18	1.50	0.64	1.06
VITi	0.19	0.09	0.17	0.15	0.12	0.14	0.18	0.14	0.31	0.10
Fe ²⁺	3.40	3.18	3.35	3.20	3.03	3.18	3.50	3.03	3.11	2.49
Mn ²⁺	0.07	0.07	0.08	0.07	0.05	0.05	0.07	0.07	0.06	3.16
Mg	0.15	0.04	0.08	0.08	0.10	0.10	0.10	0.06	0.80	2.27
Li	0.69	0.96	0.73	0.77	0.91	0.93	0.71	0.88	0.97	0.63
Y site	5.70	5.77	5.69	5.67	5.71	5.77	5.76	5.67	5.90	9.70
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.05	0.06	0.05	0.05	0.05	0.03	0.07	0.06	0.01	0.05
K	1.93	1.88	1.95	1.89	1.86	1.85	1.86	1.89	1.94	1.47
X site	1.98	1.95	2.00	1.94	1.91	1.88	1.94	1.95	1.96	1.52
O	20.06	20.09	20.06	20.05	20.05	20.07	20.07	20.07	20.06	20.91
OH	2.36	1.87	2.27	2.29	2.29	2.45	2.06	1.90	2.43	1.63
F	1.54	2.01	1.63	1.64	1.64	1.46	1.83	2.01	1.47	1.44
Cl	0.04	0.03	0.03	0.03	0.02	0.02	0.04	0.02	0.04	0.02
X(F)	0.39	0.51	0.41	0.41	0.42	0.37	0.47	0.51	0.37	0.45
X(Cl)	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
X(F/Cl)	45.00	77.32	55.97	64.66	79.85	62.72	65.46	97.81	37.18	90.15
Log(F/Cl)	1.64	1.88	1.71	1.81	1.88	1.79	1.75	1.97	1.55	1.94
F/FM	0.96	0.99	0.98	0.98	0.97	0.97	0.97	0.98	0.80	0.74
X(Phl)	0.03	0.01	0.01	0.01	0.02	0.02	0.02	0.01	0.14	0.11
X(Ann)	0.24	0.25	0.24	0.21	0.23	0.26	0.26	0.21	0.51	0.23
X(Srp)	0.73	0.74	0.75	0.77	0.76	0.72	0.72	0.78	0.36	0.76
Log[f_(H₂O)/f_(H₂O)] at different temperatures										
673.15 K	0.02	0.30	0.11	0.21	0.28	0.18	0.15	0.38	-0.34	0.10
623.15 K	-0.13	0.15	-0.04	0.06	0.13	0.03	-0.002	0.23	-0.52	-0.07
533.15 K	-0.48	-0.18	-0.37	-0.27	-0.21	-0.31	-0.34	-0.10	-0.91	-0.45

	XI (cont'd)		XII Host rock of greisens				XIII Greisens		XIV	
	76A	76B	35A1	35A2	36A1	36A2	35C	84	17	22C1b
n	8	6	4	5	6	3	6	6	4	1
SiO ₂	35.92	36.90	37.77	37.88	35.76	36.45	36.62	33.35	50.82	46.55
TiO ₂	2.90	1.67	1.22	1.45	2.01	2.00	0.01	0.82	0.15	0.16
Al ₂ O ₃	18.34	19.30	18.43	18.22	17.08	17.23	19.23	20.83	27.01	27.44
FeO	27.94	26.75	26.02	26.20	29.58	28.96	27.98	28.65	5.29	6.86
MnO	0.52	0.62	0.48	0.49	0.51	0.48	0.41	0.29	0.03	0.07
MgO	0.81	0.30	0.43	0.40	0.79	0.69	0.04	0.01	0.50	0.15
CaO	0.00	0.01	0.01	0.00	0.01	0.00	0.00	0.01	0.06	-
Na ₂ O	0.11	0.13	0.22	0.20	0.21	0.23	0.12	0.09	0.03	0.15
K ₂ O	9.88	9.89	9.54	9.53	9.62	9.69	9.53	9.82	10.20	10.55
H ₂ O*	2.67	2.57	2.32	2.38	2.52	2.50	2.48	2.62	3.23	3.19
F	2.28	2.54	3.04	2.93	2.31	2.32	2.64	2.15	2.48	2.34
Cl	0.16	0.11	0.15	0.14	0.44	0.58	0.12	0.32	0.01	0.01
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00
Total	100.56	99.69	98.34	98.56	99.76	100.02	98.03	97.98	98.75	96.47
Li ₂ O*	0.76	1.04	1.29	1.32	0.71	0.91	0.96	0.09	5.03	3.81
No of cations on the basis of 24 (O, OH, F, Cl)										
IVSi	5.50	5.71	5.87	5.88	5.67	5.73	5.79	5.42	6.55	6.32
IVAl	2.43	2.29	2.13	2.12	2.33	2.27	2.21	2.58	1.45	1.68
IVTi	0.06	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	0.88	1.24	1.25	1.21	0.87	0.92	1.38	1.40	2.65	2.71
VITi	0.45	0.19	0.14	0.17	0.24	0.24	0.00	0.10	0.01	0.02
Fe ²⁺	3.58	3.47	3.38	3.40	3.93	3.81	3.70	3.89	0.57	0.78
Mn ²⁺	0.07	0.08	0.06	0.06	0.07	0.06	0.06	0.04	0.00	0.01
Mg	0.19	0.07	0.10	0.09	0.19	0.16	0.01	0.00	0.10	0.03
Li	0.47	0.64	0.80	0.82	0.45	0.57	0.61	0.06	2.61	2.08
Y site	5.63	5.69	5.75	5.76	5.74	5.76	5.75	5.45	5.94	5.62
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	-
Na	0.03	0.04	0.07	0.06	0.06	0.07	0.04	0.03	0.01	0.04
K	1.93	1.96	1.89	1.89	1.95	1.94	1.92	2.04	1.68	1.83
X site	1.96	1.99	1.96	1.95	2.01	2.01	1.96	2.07	1.69	1.87
O	20.12	20.07	20.06	20.06	20.06	20.08	20.03	19.97	20.21	20.10
OH	2.73	2.66	2.41	2.46	2.67	2.62	2.62	2.84	2.77	2.89
F	1.10	1.24	1.50	1.44	1.16	1.15	1.32	1.10	1.01	1.01
Cl	0.04	0.03	0.04	0.04	0.12	0.15	0.03	0.09	0.00	0.00
X(F)	0.28	0.32	0.38	0.36	0.29	0.29	0.33	0.27	0.27	0.26
X(Cl)	0.01	0.01	0.01	0.01	0.03	0.04	0.01	0.02	0.00	0.00
X(F/Cl)	27.92	45.61	48.24	60.50	10.61	7.59	42.87	12.79	1011.1	397.48
Log(F/Cl)	1.43	1.65	1.63	1.70	1.01	0.88	1.61	1.10	2.59	2.60
F/FM	0.95	0.98	0.97	0.97	0.95	0.96	1.00	1.00	0.86	0.96
X(Phl)	0.03	0.01	0.02	0.02	0.03	0.03	0.00	0.00	0.02	0.01
X(Ann)	0.23	0.21	0.27	0.29	0.29	0.30	0.21	0.06	0.20	0.11
X(Srp)	0.74	0.78	0.71	0.69	0.68	0.67	0.79	0.94	0.78	0.89
Log[f _(M) /f _(M1)] at different temperatures										
673.15 K	-0.21	0.06	0.03	0.10	-0.63	-0.75	0.05	-0.45	0.99	1.02
623.15 K	-0.37	-0.09	-0.12	-0.05	-0.78	-0.90	-0.10	-0.61	0.84	0.88
533.15 K	-0.71	-0.43	-0.46	-0.39	-1.13	-1.25	-0.43	-0.94	0.51	0.54

	XIV	XV			Darkav	Paleav
	22C2b	36B1b	36C1b	36C2b		
n	2	2	3	4	113	16
SiO2	42.22	48.99	48.18	48.20	37.24	47.49
TiO2	0.66	0.05	0.09	0.07	1.56	0.20
Al2O3	22.42	30.98	29.78	29.22	18.69	27.81
FeO	17.54	2.96	4.53	4.82	26.16	7.00
MnO	0.26	0.02	0.34	0.42	0.50	0.19
MgO	0.36	0.47	0.15	0.31	0.57	0.32
CaO	0.01	0.02	0.10	0.02	0.01	0.04
Na2O	0.14	0.07	0.09	0.05	0.17	0.09
K2O	9.98	9.82	9.83	10.22	9.64	10.10
H2O#	2.28	3.77	3.84	3.57	2.31	3.31
F	3.63	1.49	1.23	1.74	3.06	2.15
Cl	0.07	0.00	0.02	0.01	0.17	0.02
O=F+Cl	-2.00	-0.60	-0.50	-0.70	-1.00	-0.90
Total	98.01	98.02	97.66	97.92	98.75	97.81
Li2O*	2.56	4.51	4.28	4.28	1.14	4.08

No of cations on the basis of 24 (O, OH, F, Cl)

IVSi	6.12	6.31	6.31	6.32	5.72	6.32
IVAl	1.88	1.69	1.69	1.68	2.18	1.68
IVTi	-	-	-	-	0.00	-
T site	8.00	8.00	8.00	8.00	7.90	8.00
VIA1	1.93	3.01	2.90	2.84	1.21	2.68
VITi	0.07	0.00	0.01	0.01	0.17	0.02
Fe2+	2.16	0.32	0.50	0.53	3.37	0.81
Mn2+	0.03	0.00	0.04	0.05	0.24	0.02
Mg	0.08	0.09	0.03	0.06	0.25	0.06
Li	1.48	2.33	2.25	2.26	0.70	2.17
Y site	5.75	5.76	5.73	5.75	5.94	5.76
Ca	0.00	0.00	0.01	0.00	0.00	0.01
Na	0.04	0.02	0.02	0.01	0.05	0.02
K	1.85	1.61	1.64	1.71	1.89	1.72
X site	1.89	1.63	1.68	1.73	1.94	1.75
O	20.11	20.15	20.13	20.15	20.11	20.14
OH	2.18	3.24	3.36	3.13	2.36	2.93
F	1.69	0.61	0.51	0.72	1.48	0.92
Cl	0.02	0.00	0.00	0.00	0.05	0.01
X(F)	0.43	0.16	0.13	0.19	0.38	0.24
X(Cl)	0.00	0.00	0.00	0.00	0.01	0.00
X(F/Cl)	253.72	483.78	588.76	140.34	51.79	479.19
Log(F/Cl)	2.21	1.34	2.36	1.52	1.61	2.10
F/FM	0.97	0.78	0.95	0.91	0.95	0.90
X(Phl)	0.01	0.02	0.01	0.01	0.03	0.01
X(Ann)	0.18	0.05	0.07	0.09	0.25	0.12
X(Srp)	0.81	0.93	0.94	0.91	0.73	0.88

Log[f_(H₂O)/f_(H₂)] at different temperatures

673.15 K	0.61	-0.26	0.78	-0.07	-0.03	0.51
623.15 K	0.46	-0.41	0.64	-0.22	-0.18	0.37
533.15 K	0.13	-0.74	0.30	-0.55	-0.52	0.03

Chloritized biotite						
	22Ac	58c	22C2c	36A1c	36A2ac	36A2bc
n	2	1	1	1	2	1
SiO2	23.54	24.47	22.84	23.34	23.68	28.05
TiO2	0.07	0.15	0.07	0.07	0.09	0.01
Al2O3	19.13	18.81	19.13	18.84	19.08	19.03
FeO	44.38	45.39	45.79	45.71	45.81	41.62
MnO	0.37	0.11	0.37	0.33	0.33	0.29
MgO	0.22	0.06	0.22	0.09	0.21	0.24
CaO	0.01	0.02	0.03	0.03	0.00	0.03
Na2O	0.00	0.01	0.04	0.03	0.03	1.71
K2O	0.04	0.10	0.02	-	0.01	0.02
H2O#	3.50	3.50	3.46	3.49	3.44	3.55
F	0.10	0.06	0.12	0.08	0.18	0.20
Cl	0.01	0.08	0.01	-	0.01	0.01
O=F+Cl	-0.04	0.04	-0.05	0.03	-0.08	-0.09
Total	91.33	92.73	92.05	91.97	92.80	94.68
Li2O*	-	-	-	-	-	-

Structural formula on the basis of 24(O,OH,F,Cl)

IVSi	4.52	4.46	4.41	4.49	4.50	4.95
IVAl	3.48	3.54	3.59	3.51	3.50	3.05
IVTi	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.85	0.50	0.76	0.76	0.77	0.91
VTi	0.01	0.02	0.01	0.01	0.01	0.00
Fe2+	7.13	6.92	7.39	7.35	7.27	6.14
Mn2+	0.06	0.02	0.06	0.05	0.05	0.04
Mg	0.06	0.02	0.06	0.03	0.06	0.06
Li	-	-	-	-	-	-
Y site	5.95	7.48	5.96	5.99	6.06	6.09
Ca	0.00	0.00	0.01	0.01	0.00	0.01
Na	0.00	0.01	0.01	0.01	0.01	0.58
K	0.01	0.02	0.00	-	0.00	0.00
X site	0.01	0.03	0.02	0.02	0.01	0.59
O	19.45	20.00	19.47	19.48	19.53	19.70
OH	4.49	3.94	4.45	4.47	4.36	4.18
F	0.06	0.03	0.07	0.05	0.11	0.11
Cl	0.00	0.02	0.00	-	0.00	0.00
X(F)	0.01	0.01	0.02	0.01	0.02	0.03
X(Cl)	0.00	0.01	0.00	-	0.00	0.00
F/Cl	22.98	1.35	32.26	0.01	53.26	28.57
F/FM	0.99	1.00	0.99	1.00	0.99	0.99

Table 4.3: Range of dark mica composition(s) in the various rock units

Rock type	Range of dark mica composition(s)
Rhyolite porphyry	$\text{Ann}_{42-94}\text{Srp}_{0-6}\text{Phl}_{10-6}$, ${}^{\text{IV}}\text{Ti} = 0.32$, $0.87 \leq \text{F/FM} \leq 0.93$, $0.3 \leq \text{F/Cl} \leq 0.74$
Hypersolvus (fayalite) syenogranite:	
a) Fine-grained (late-magmatic)	$\text{Ann}_{60-69}\text{Srp}_{39-27}\text{Phl}_{0-5}$, $0.22 \leq {}^{\text{VI}}\text{Ti} \leq 0.40$, $0.90 \leq \text{F/FM} \leq 0.99$, $0.17 \leq \text{F/Cl} \leq 1.35$
b) Fine-grained (hydrothermal)	$\text{Ann}_{72}\text{Srp}_{28}\text{Phl}_0$, ${}^{\text{VI}}\text{Ti} = 0.38$, $\text{F/FM} = 1.00$, $\text{F/Cl} = 0.41$
c) Coarse-grained (secondary)	$\text{Ann}_{63}\text{Srp}_{34}\text{Phl}_4$, ${}^{\text{VI}}\text{Ti} = 0.44$, $\text{F/FM} = 0.96$, $\text{F/Cl} = 0.13$
"Hornblende"-biotite granite porphyry	$\text{Ann}_{48-71}\text{Srp}_{52-76}\text{Phl}_{0-10}$, $0 \leq {}^{\text{VI}}\text{Ti} \leq 0.45$, $0.90 \leq \text{F/FM} \leq 1.00$, $0 \leq \text{F/Cl} \leq 1.09$
Biotite granite porphyry	$\text{Ann}_{32-46}\text{Srp}_{58-49}\text{Phl}_{1-6}$, $\text{F/FM} = 0.95$, $0.27 \leq {}^{\text{VI}}\text{Ti} \leq 0.31$, $1.85 \leq \text{F/Cl} \leq 3.20$
Coarse-grained biotite granite	$\text{Ann}_{12-48}\text{Srp}_{77-41}\text{Phl}_{4-13}$, $0.06 \leq {}^{\text{VI}}\text{Ti} \leq 0.33$, $0.88 \leq \text{F/FM} \leq 0.93$, $4.56 \leq \text{F/Cl} \leq 18.89$
Medium-grained biotite granite	$\text{Ann}_{17-53}\text{Srp}_{80-36}\text{Phl}_{2-14}$, $0.85 \leq \text{F/FM} \leq 0.96$, $0.05 \leq {}^{\text{VI}}\text{Ti} \leq 0.40$, $5.18 \leq \text{F/Cl} \leq 42.43$
Fine-grained biotite granite:	
a) hypersolvus granite	$\text{Ann}_{71-80}\text{Srp}_{29-20}\text{Phl}_0$, $\text{F/FM} = 1.0$, $0.24 \leq {}^{\text{VI}}\text{Ti} \leq 0.39$, $4.27 \leq \text{F/Cl} \leq 8.4$
b) topaz-cassiterite-bearing (zinnwaldite type)	$\text{Ann}_{15-29}\text{Srp}_{85-68}\text{Phl}_{0-3}$, $\text{F/FM} = 0.99$, $0 \leq {}^{\text{VI}}\text{Ti} \leq 0.32$, $13.33 \leq \text{F/Cl} \leq 185.13$
c) drusy granite (zoned zinnwaldite-trilithionite)	Zones: a) <i>bluish green</i> :- $\text{Ann}_{29-31}\text{Srp}_{67-80}\text{Phl}_1$, $0.02 \leq {}^{\text{VI}}\text{Ti} \leq 0.1$, $184.66 \leq \text{F/Cl} \leq 384.65$, b) <i>green to pale green</i> :- $\text{Ann}_{33-35}\text{Srp}_{66-65}\text{Phl}_1$, $0.8 \leq {}^{\text{VI}}\text{Ti} \leq 0.18$, $118.45 \leq \text{F/Cl} \leq 140.5$, and c) <i>pale coloured</i> :- $\text{Ann}_{33}\text{Srp}_{66}\text{Phl}_1$, ${}^{\text{VI}}\text{Ti} = 0.04$, $\text{F/Cl} = 882.67$; $\text{F/FM} = 0.98$
Albitized biotite granite:	
a) leucocratic/hematite-speckled	$\text{Ann}_{18-30}\text{Srp}_{81-67}\text{Phl}_{1-3}$, $0.97 \leq \text{F/FM} \leq 0.99$, $0.04 \leq {}^{\text{VI}}\text{Ti} \leq 0.31$, $17.89 \leq \text{F/Cl} \leq 166.12$
b) leucocratic/drusy	$\text{Ann}_{47-57}\text{Srp}_{40-29}\text{Phl}_{13-14}$, $0.28 \leq {}^{\text{VI}}\text{Ti} \leq 0.35$, $\text{F/FM} = 0.80$, $28.03 \leq \text{F/Cl} \leq 60.85$
Post-joint greisens:	
a) host rock of greisens	$\text{Ann}_{24-35}\text{Srp}_{73-60}\text{Phl}_{1-5}$, $0.12 \leq {}^{\text{VI}}\text{Ti} \leq 0.28$, $0.95 \leq \text{F/FM} \leq 0.97$, $6.68 \leq \text{F/Cl} \leq 96.39$
b) greisen mica (type 1)	$\text{Ann}_{17-24}\text{Srp}_{82-75}\text{Phl}_1$, $\text{F/FM} = 1.00$, ${}^{\text{VI}}\text{Ti} = 1.0$, $29.92 \leq \text{F/Cl} \leq 65.02$
c) greisen mica (type 2a)	$\text{Ann}_{4-10}\text{Srp}_{94-87}\text{Phl}_{0-3}$, $0.74 \leq \text{F/FM} \leq 0.97$, $0 \leq {}^{\text{VI}}\text{Ti} \leq 0.02$, $115 \leq \text{F/Cl} \leq 1487.3$
d) greisen mica (type 2b)	$\text{Ann}_{3-10}\text{Srp}_{90-97}\text{Phl}_0$, $\text{F/FM} = 1.0$, $0 \leq {}^{\text{VI}}\text{Ti} \leq 0.27$, $8 \leq \text{F/Cl} \leq 15.85$

Table 4.4: Inferred ratio of halogen fugacities $\{\text{Log}[f_{(\text{HF})}/f_{(\text{HCl})}]\}$ for Kwandonkaya suite.

ROCK TYPE	TEMPERATURE, KELVIN (i.e., $^{\circ}\text{C} + 273.15$)	AVERAGE OF $\text{Log}[f_{(\text{HF})}/f_{(\text{HCl})}]$	RANGE OF $\text{Log}[f_{(\text{HF})}/f_{(\text{HCl})}]$
MICA IN SUBVOLCANIC ROCKS (Annetic biotite)	i) 773.15 K or 500°C ii) 673.15 K	-1.74 -1.99	-2.32 to -1.0 -2.56 to -1.24
MICA IN PRISTINE BIOTITE GRANITE	a) Dark Mica (Fp, Srp-Znw) i) 753.15 K ii) 673.15 K ii) 573.15 K b) Pale Mica (Znw - Trl) i) 753.15 K ii) 673.15 K iii) 573.15 K	-0.13 -0.33 -0.67 1.12 0.92 0.60	-0.72 to 1.01 -0.94 to 0.81 -1.29 to 0.49 0.76 to 1.54 0.56 to 1.34 0.24 to 1.01
MICA IN ALTERED PLUTONIC ROCKS	a) Dark mica (Fp, Srp-Znw) i) 673.15 K ii) 623.15 K iii) 533.15 K b) Pale Mica (Znw - Trl) i) 673.15 K ii) 623.15 K iii) 533.15 K	-0.03 -0.18 -0.52 0.51 0.37 0.03	-0.75 to 0.38 -0.90 to 0.23 -1.25 to -0.10 -0.26 to 1.02 -0.41 to 0.88 -0.74 to 0.54

Symbols of mineral phases are defined in Table 4.1. Results in italics are for inferred halogen fugacity ratio at 673.15 K (i.e., 400°C) used to compare ratios at the same temperature. The results in roman type are for the range of temperature of equilibration for the intrusive unit(s).

Chapter 5: Alkali feldspars as monitors of magmatic evolution, and postmagmatic equilibration.

5.1 Introduction

The alkali feldspar granites, characteristic of the Kwandonkaya complex, and indeed of the NYG (Nigerian Younger granites), are composed predominantly (60 to 90%) of alkali feldspar. The alkali feldspars are sensitive indicators of 1) degree of magmatic evolution in terms of their bulk composition, in particular their Ca, Na, and K contents, and 2) postmagmatic adjustments in terms of degree of Al-Si order. Badejoko (1977) undertook the most extensive studies of feldspar mineralogy of the NYG and established a relationship between the degree of Al-Si order in K-feldspar and the degree of trace-element enrichment of the NYG rock units. Specifically, he indicated a positive correlation between degree of order of the K-feldspar and the degree of enrichment in Nb, Rb, Li, Sn, and F in the rocks. Barren granites were found to contain structurally intermediate microcline, whereas "stanniferous" granites contain more of these elements and a higher degree of Al-Si order in the microcline (Badejoko 1984, 1986). In the Ririwai complex, which contains the largest lode-type Sn-W (and Pb-Zn) mineralization in the NYG, the state of the alkali feldspar has been investigated so as to shed light on the nature and extent of rock-fluid interactions (Martin & Bowden 1981). Their study shows that: a) a spectrum of composition of alkali feldspars extends from imperfectly ordered microcline + ordered albite in subequal proportions to albitites or to microclinites in zones of intense alteration; b) the feldspars followed paths of ordering intermediate between one-step and two-step schemes; and c) the peraluminosity of the granites was developed as a result of rock-fluid interaction, of which the feldspars were good indicators. Finally, in the Tibchi complex, Na-metasomatism was found to affect screens of basement founded in the granite upon cauldron subsidence (Ike 1979). The plagioclase has abnormal cell parameters interpreted to be due to "memory effects", owing to

albitization of an oligoclase or andesine precursor.

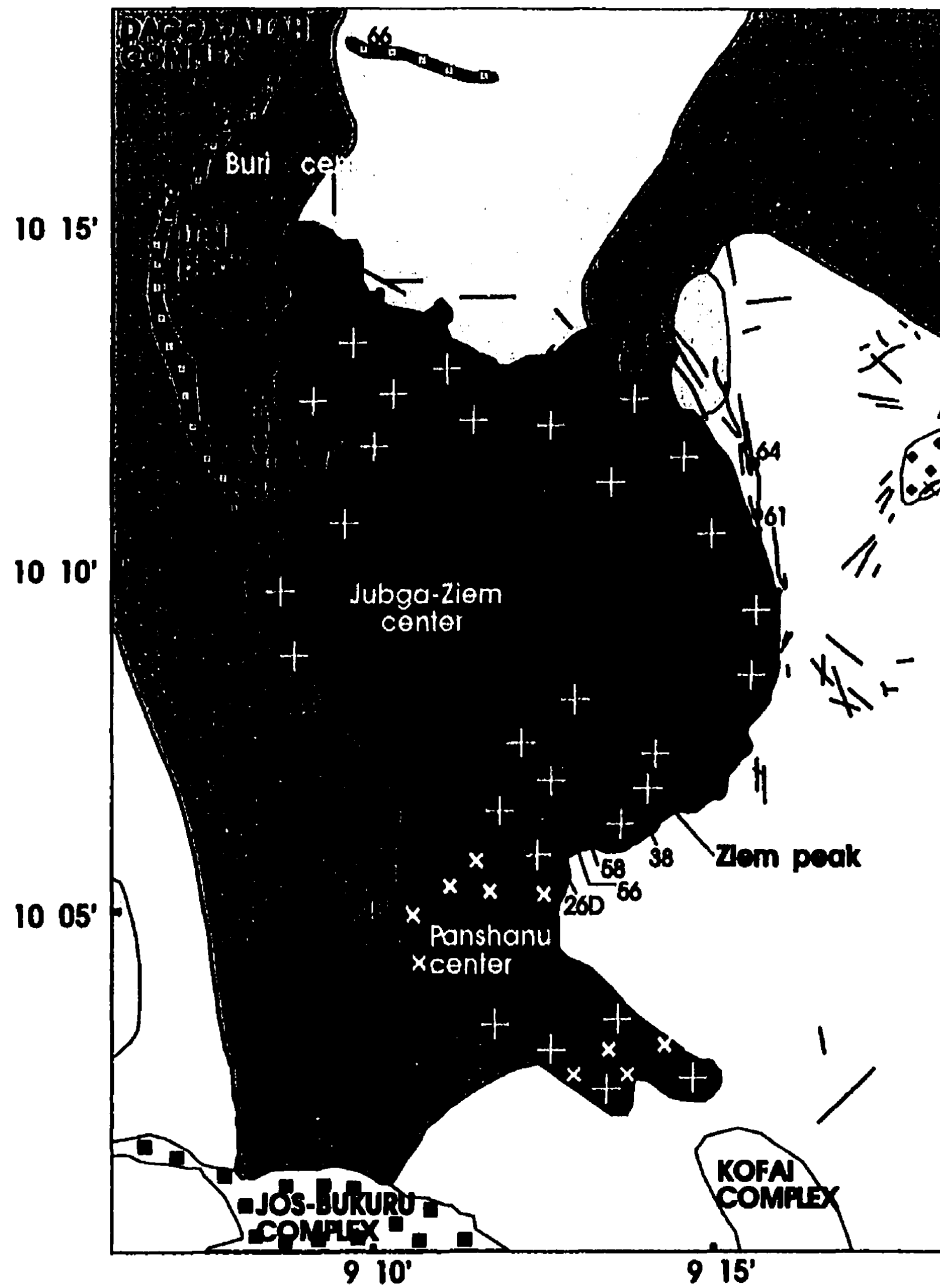
5.2 Methods of Study

The methods used in this study include petrographic characterization of alkali feldspars in 40 thin sections of various rock units. Back-scattered electron images obtained with a scanning electron microscope were used to document some of the textural features. X-ray-diffraction (XRD) analyses (powder method) of 24 samples using a Guinier-Hägg camera ($\text{CuK}\alpha_1$) led to calculation of unit-cell parameters (Appendix V.1) using indexed 2θ reflections as input for the program of Appleman and Evans (1973). A spinel standard ($a = 8.0833\text{\AA}$) was used to correct the diffraction pattern. In the section that follows, the name applied to the K-feldspar reflects the results of the XRD analysis, in view of the lack of distinguishing criteria between orthoclase and microcline in thin section.

5.3 Petrography of subvolcanic rocks

In *ryholite porphyry* from the Buri ring dyke (Fig. 5.1), the orthoclase microperthite phenocrysts constitute 25 to 50% of the rock. Other minerals are described in Table 2.1. The groundmass (40-50%) is composed mainly of a microcrystalline intergrowth of orthoclase microperthite and quartz, the assemblage probably representing devitrified glass. Typically, the translucent phenocrysts of alkali feldspar contain a clear crypto- to microperthitic core that has strain shadows and an irregular turbid albite-rich lining of fractures and cleavages (Figs. 5.2a, b). The rounded to subangular grains (Table 5.1, Fig. 5.2a) reflect abrasion owing to transport of entrained phenocrysts within a crystal-rich viscous melt, before degassing and freezing, during emplacement in the ring dyke. The feldspar phenocrysts contain inclusions of fayalite, hedenbergite, Fe-Ti oxides, chevkinite, and zircon, and thus formed after these phases.

Fig. 5.1: Geological map of Kwandonkaya complex with location of samples used in this study of feldspar mineralogy.



• Sample location

▨ Basalt dyke

MESOZOIC GRANITES

G: Grelsens

Plutonic rocks

■ Fine- to medium-grained biotite granite

⊕ Coarse-grained biotite granite

⊗ Porphyritic biotite microgranite

Subvolcanic rocks

■ Hornblende biotite-, biotite-granite porphyry, +/- fayalite syenogranite

▨ Hedenbergite fayalite granite porphyry

▨ Rhyolite porphyry

▨ Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

▨ Biotite hornblende granite

■ Gneissic granite

▨ Migmatite

1 0 2 4 6 8 km

Table 5.1: Summary of petrographic features of alkali feldspars in rock units (Fig. 5.1).

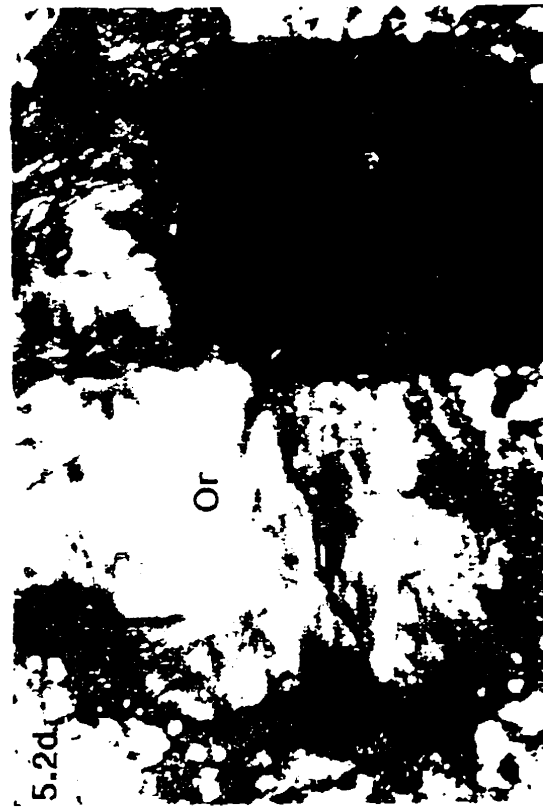
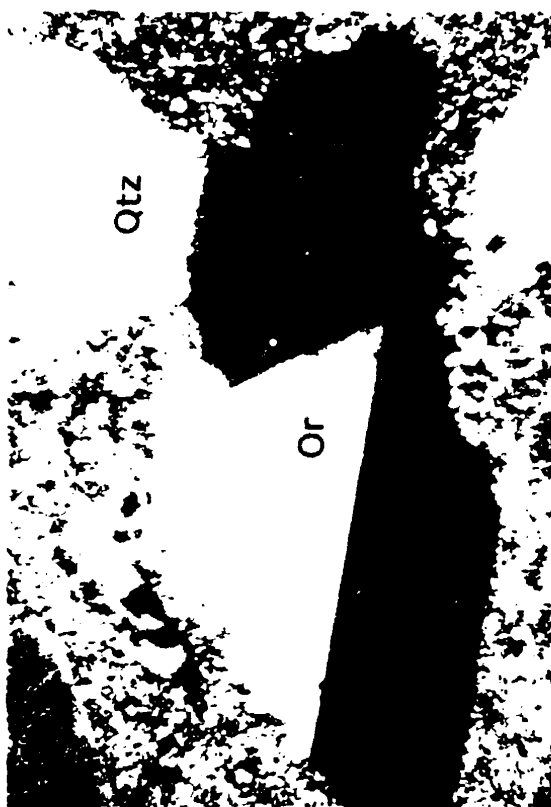
Rock Type	Petrographic characteristics alkali feldspar
Rhyolite porphyry	Phenocrysts: Green, grey, and pinkish grey; anhedral, subhedral, rounded and broken (angular) grains; simple twins, sector zoned (twinned), and twins with sinusoidal interface; Crypto- to microperthitic; Turbidity is mild (0 to 20 % of phenocrysts). Groundmass: microcrystalline intergrowth of alkali feldspar with quartz.
Hedenbergite fayalite granite porphyry. a) Normal type	Phenocrysts: Euhedral prismatic to subhedral; Synneusis is characteristic; Sector zoning (twins) and growth twins; Rhythmic zoning in some sections; Crypto to microperthitic; Pronounced turbidity (15 to 80% of grains). Groundmass: microcrystalline feldspar and quartz.
b) Reddened type	Phenocrysts: Clouded and zoned grains: cream or whitish core with red or pink rims. Both phenocrysts and groundmass feldspar are completely turbid; Groundmass also is reddened.
Felsite porphyry and leucocratic granite porphyry	Phenocrysts: Euhedral crypto- to microperthitic, and rare granophyric type; Some are sector zoned. Turbidity from 20 to 100% of grains. Groundmass: spherulitic, granophyric, and normal microcrystalline intergrowth of quartz and feldspar.
Hedenbergite hastingsite granite porphyry, hastingsite granite porphyry	Similar to those of feldspar in hedenbergite fayalite granite porphyry
Hypersolvus syenogranite	Greenish feldspar in coarse-grained type, and pinkish grey in the fine-grained variants. Alkali feldspar are mostly granular, with braid perthite; orthoclase with local granophyric intergrowths.
"Hornblende"-biotite granite porphyry and Biotite granite porphyry	Phenocrysts: green to grey, euhedral to subhedral, turbid microperthite; albite-rich domains occur. Groundmass is microcrystalline intergrowth of alkali feldspar and quartz, that locally develops a granophyric texture.
Coarse-grained biotite granite	Cream-coloured, granular to prismatic microcline microperthite; intergranular and intragranular albite. Sericite and pale mica partly replace albite.
Fine- to medium-grained biotite granite	Hypersolvus type: Cream coloured crypto- to microperthitic orthoclase, no distinct grains of albite; Drusy and less altered types: pinkish grey microcline microperthite, inter- and intragranular albite, $\pm$ cross-hatched microcline. Feldspar grains are euhedral in druses, and anhedral in other instances
Albitized biotite granite	Cream-coloured to chalk white and hematite-speckled grains. Relics of twinned albite in a matrix of untwinned albite, microcline and quartz.
Post-joint greisens	Relict turbid grains of microcline microperthite; grains largely replaced by sericite, and pale-coloured mica (siderophyllite and trilithionite).

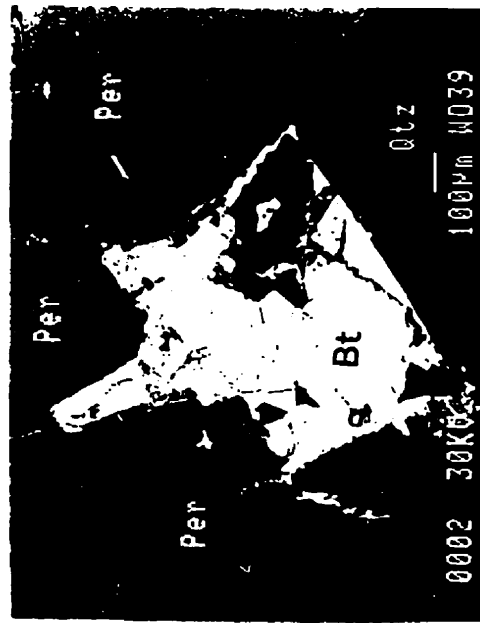
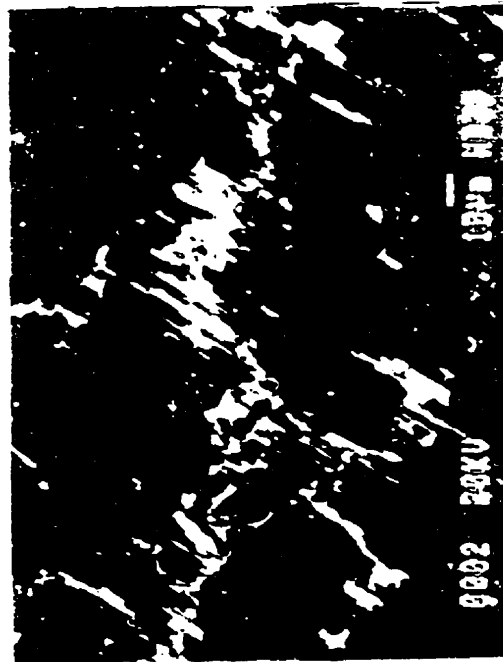
Also found in the Buri ring dyke is *hedenbergite fayalite granite porphyry*. The colour of the alkali feldspar varies from dark grey to green. Turbidity in the crypto- to microperthitic orthoclase phenocrysts is pronounced along the margins and fractures, and ranges from 15 to 40% replacement of the clear microperthite (Figs. 5.2c, d, and Appendix V.2). Oscillatory growth-zoning is locally well displayed (Fig. 5.2d). The microcrystalline groundmass is relatively coarse, compared to that in the rhyolite porphyry, and is also composed of orthoclase microperthite and quartz. The sequence of crystallization, established in Chapter 2, is similar to that of the rhyolite porphyry.

In the Jubga-Ziem center, in the *felsite porphyry* (within cone sheets), most of the volume of alkali feldspar phenocrysts is visibly microperthitic. The phenocrysts themselves may be granophyric, i.e., they may contain a regular intergrowth of orthoclase and quartz arranged in sectors (Fig. 2.4, Chapter 2). Although the quartz appears platy in some sectors, and thus possibly formerly was cristobalite now inverted to quartz (cf. Smith 1974), the shape of the quartz domains in the central sector suggests a series of rods of common orientation within a cellular alkali feldspar. Such "granophyric" phenocrysts are not a common feature, as indicated in Chapter 2. The groundmass is variable in texture, from a granophyric (rare) to microcrystalline intergrowth of orthoclase and quartz (Table 5.1). In the *hedenbergite hastingsite granite porphyry and hastingsite granite porphyry*, also within cone sheets, the orthoclase microperthite phenocrysts resemble those in the Buri ring dyke (Figs. 5.2c, d). The development of turbidity in the microperthitic phenocrysts also ranges from marginal to nearly complete (Fig. 5.2e). The alkali feldspar postdates hedenbergite, whereas it crystallized earlier than amphibole in these units.

In the *hypersolvus syenogranite*, found within the Jubga-Ziem ring dyke, the alkali feldspar grains are typically braid perthite (Fig. 5.2f) interpreted to be examples of extreme coarsening of perthite firstly involving coherent exsolution. The continuity of the braid perthite

Fig. 5.2: Petrographic characteristics of alkali feldspars in subvolcanic rocks: Photomicrographs of: **a)** Cryptoperthitic orthoclase (Or) in rhyolite porphyry (sample # 49, Buri ring dyke). Strain shadows occur in the upper right corner of the phenocryst, which has a rounded outline. The groundmass is composed of microcrystalline quartz and orthoclase. **b)** Subhedral phenocryst of Carlsbad-twinned cryptoperthitic orthoclase (Or) in rhyolite porphyry (sample 49, Buri ring dyke). The grain displays patchy turbidity in the central and marginal parts. The groundmass is microcrystalline quartz and orthoclase. **c)** Orthoclase microperthite (Or) in hedenbergite fayalite granite porphyry (sample 43, Buri ring dyke) with advanced patchy turbidity in both central and marginal parts of the phenocryst. Rhythmic concentric zoning is obvious in the grey unreplaced parts of the grain. **d)** Phenocrysts of microperthitic orthoclase (Or) displaying synneusis of grains (sample 43, Buri ring dyke). Turbidity occurs around the margin of the grains and also along the fractures. The groundmass is granophyric. Width of field of view in these photomicrographs is 3.1 mm; the sample numbers cited are shown in Figure 5.1. **e)** Orthoclase microperthite (Or) with an inclusion of chevkinite microphenocryst (Ckt) in hastingsite granite porphyry (sample 40A, cone sheet); the lower half of the grain is replaced by turbid microperthite (Per), which also forms a mantle around the grain. **f)** Carlsbad-twinned orthoclase microperthite (Per) characterized by braid perthite in hypersolvus syenogranite (sample 58, Jubga-Ziem center). The presence of microcline in addition to orthoclase is documented in this sample by XRD analyses. Interstitial grain of quartz has inclusions of altered fayalite. **g)** Back-scattered electron (BSE) image of microperthite (Or) from marginal parts of hypersolvus syenogranite (sample 57, Jubga-Ziem center). Veinlets of dark grey quartz (Qtz) are intimately associated with brighter Fe-rich chlorite and annitic biotite (Bt). **h)** BSE image of microperthite (Per). Marginal zone of dark grey albite attains 100 μm in width (see perthite grain above white biotite + amphibole). Biotite (Bt) is interstitial. Qtz = quartz.





in some sections is breached by optically continuous inclusions of granophyric quartz. Near the margin of the intrusions, microveins of albite are accompanied by Fe-rich chlorite or annite (Fig. 5.2g), indicative of postmagmatic interaction with a NaFe-bearing fluid (cf. Bowden 1985). Microphenocrysts of fayalite occur within the microperthite, indicating that the microperthite, which constitutes 60 to 70% of the rock, crystallized after fayalite but before quartz and interstitial granophyre.

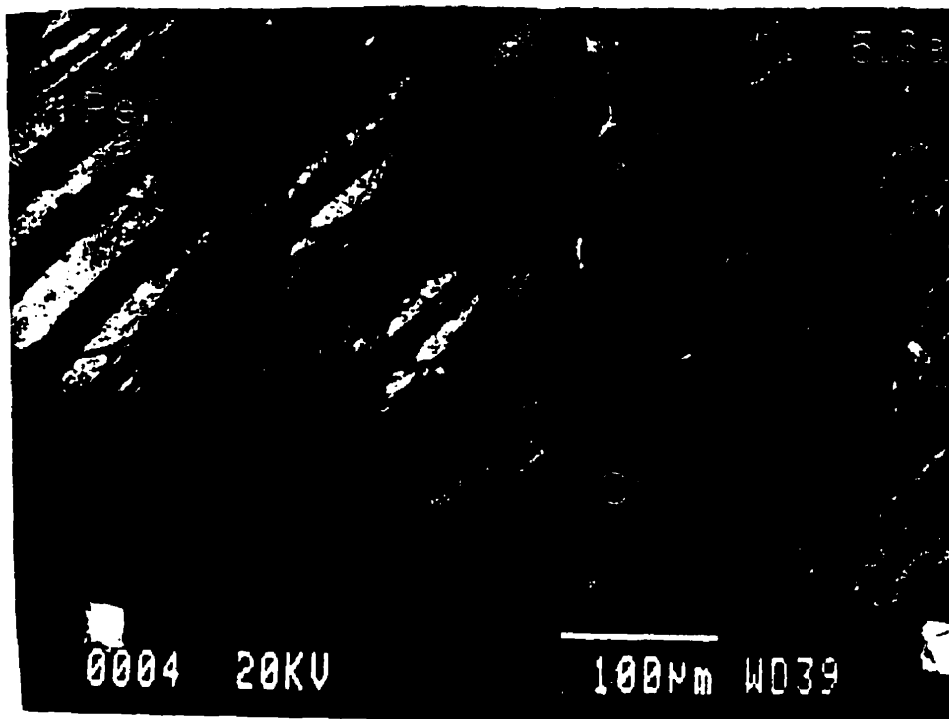
"Hornblende"-biotite granite porphyry and *biotite granite porphyry* of the same ring dyke are characterized by progressive coarsening of the perthitic texture in orthoclase and microcline. The coarsening is more advanced than in the hypersolvus syenogranite in the same ring dyke and other granite porphyries in the Buri ring dyke. Marginal albite-rich domains encroach on relics of microperthite (Fig. 5.2h), with some of the albite domains reaching a width of over 100 μm .

5.4 Petrography of plutonic rocks

These rock types occur mainly in the Jubga-Ziem and Panshanu centers, composed of different textural types of biotite granites (Fig. 5.1), which include coarse-grained, and fine- to medium-grained variants.

Coarse-grained biotite granite has a hypersolvus magmatic assemblage (Table 5.1) with the development of an interstitial (late-magmatic) and postmagmatic subsolvus assemblage. The replacement of the hypersolvus by subsolvus assemblage is only incipient. *Fine- to medium-grained biotite granite* varies from equigranular to drusy types. The hypersolvus variant is characterized by microcline microperthite grains that are cryptoperthitic in the core and microperthitic in the margin (Fig. 5.3a). The drusy type of biotite granite is also hypersolvus and has euhedral pinkish or reddish brown microcline in miarolitic cavities (along with α -quartz, biotite, $\pm$ topaz, $\pm$ aquamarine). In most cases, the drusy type of granite, along with the other less

Fig. 5.3: SEM - BSE images of alkali feldspar in plutonic rocks: a) orthoclase (Or) in hypersolvus biotite granite with a cryptoperthitic core bearing fine discontinuous lamellae, whereas the margin has relatively coarse and continuous lamellae. The Na-rich lamellae are dark grey whereas the K-rich parts are light grey in the margin. b) Drusy biotite granite bearing orthoclase microperthite (Or), albite (Ab), zoned mica (Bt), and quartz. The microperthite and albite have an interdigitating boundary.



altered variants, contain grid-twinned microcline, intragranular albite, and interstitial albite (Table 5.2, Fig. 5.3b). This indicates a hypersolvus mineralogy overprinted by a subsolvus assemblage.

Amongst the altered plutonic rocks, the *albitized biotite granite* occurs in the marginal and upper parts of the more pristine fine- to medium-grained biotite granite. Grid-twinned microcline and twinned albite occur in a matrix of untwinned fine-grained albite, microcline, and quartz. The grains of K-feldspar are preferentially replaced by albite, whereas grains of albite of a size similar to that in the less altered granite are preserved. *Post-joint greisens* develop a variety of textural types that include fine-grained, greenish grey and drusy quartz-rich types. The greisens are composed of relatively coarse quartz (with subsidiary fine-grained intergranular quartz), and aggregates of mica chloritized to various extents. The proportion of feldspar varies from 60% in the host rock of greisens (with microcline microperthite, and albite) to between 0 and 20% in the greenish grey greisen zone, where the feldspar is destroyed and replaced by sericite and pale-coloured mica (siderophyllite and trilithionite).

5.5 Degree of Al-Si order in the alkali feldspars

Unit-cell parameters were used to infer the degree of Al-Si order and the mol fraction of the K end-member (N_{K}) in the coexisting feldspars. The degree of order, τ_{10} , i.e., the Al occupancy of the T_1O site, can be derived from b , c , α' , and γ' using the equations of Kroll and Ribbe (1983). Where Al and Si are randomly distributed over the four tetrahedral sites in a given formula unit, as could be expected in a magmatic alkali feldspar (sanidine), τ_1 (or τ_2) = 0.25. Such a disordered distribution is analogous to that in the silicate melt structure, and a monoclinic alkali feldspar results (Appendix V.2a). On the other hand, complete Al-Si order characterizes a K-rich feldspar equilibrated below 450°C (Parsons 1978, Brown & Parsons 1989). Hence, τ_{10} equals 1 in well-ordered microcline (Appendix V.2b, c). Results displayed in Figures 5.4, 5.5, and 5.6 are

derived from the unit-cell parameters (Table 5.2, Appendix V.1). Where possible, phenocrysts and matrix were characterized separately.

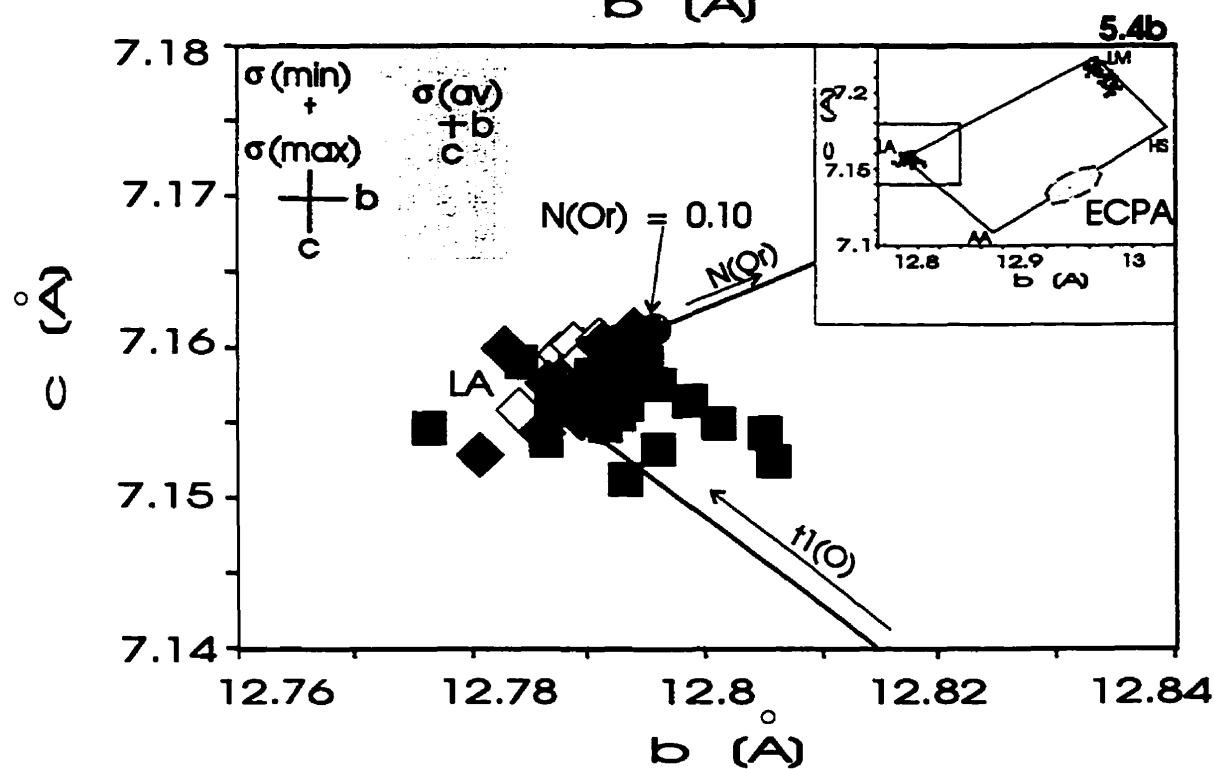
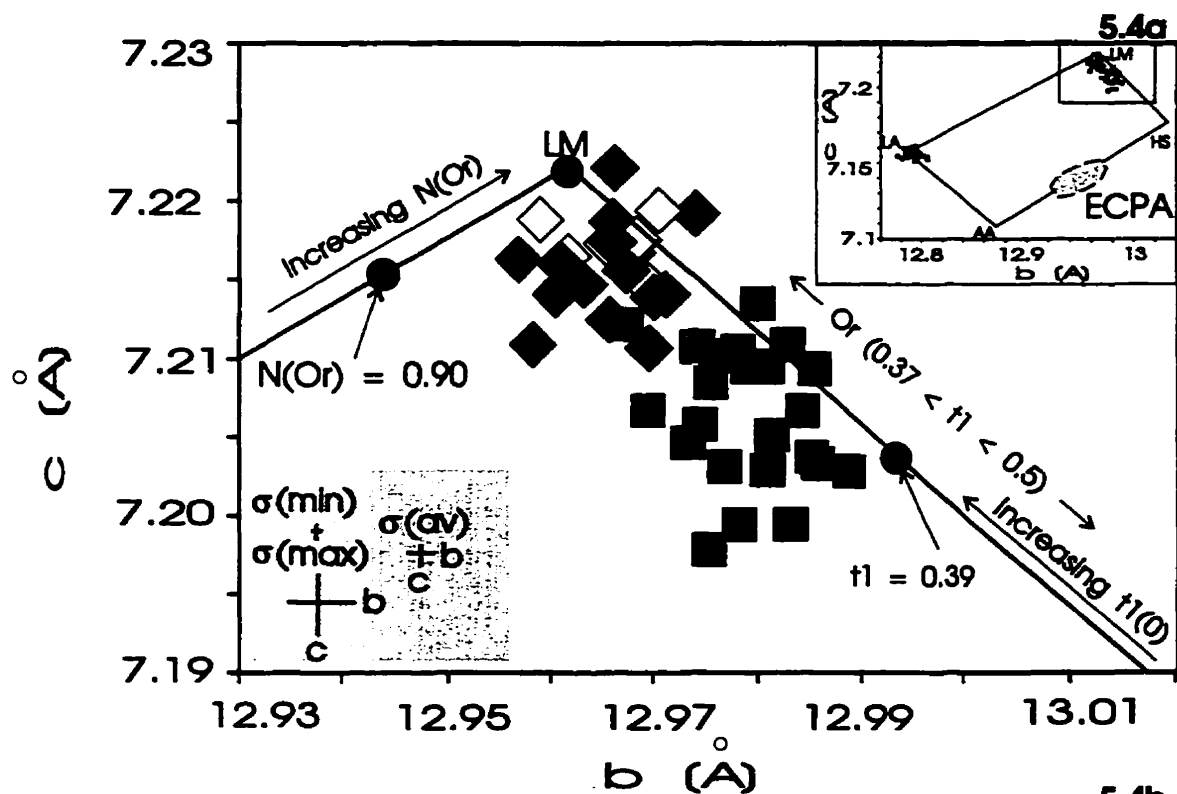
One of the most striking aspects of the data collected on the K-feldspar is the absence of sanidine and of well-ordered microcline. Whereas most of the subvolcanic rocks (Buri ring dyke and cone sheets) contain alkali feldspar apparently devoid of exsolution lamellae, the data show that all rock units in the Kwadonkaya complex, without exception, have been equilibrated to various extents in the subsolidus region. Typical values of N_{Or} are in the range 0.9 - 1 (Table 5.2, Fig. 5.4a). Thus, even where exsolution lamellae are not visible with the optical microscope, in the phenocrysts of the rhyolite porphyry, for example, the data show that such lamellae are present and developed on a submicroscopic scale (i.e., the phenocrysts are cryptoperthitic).

Also surprising is the absence of end-member low microcline in the Kwadonkaya pluton. In spite of the clear evidence that exsolution and ordering have occurred, the ordering has not reached completion. Instead, what is typical in this pluton is a large range of values of t_{10} spanning the interval orthoclase - microcline (Fig. 5.4a). The most ordered microcline has a t_{10} of 0.98. Some rocks contain orthoclase without microcline, others contain microcline without orthoclase, and in some instances both are present. Generalizations concerning the distribution of these groups will be made below.

The sodic feldspar in this suite also defines a relatively large spread (Fig. 5.4b), compared to that encountered in other anorogenic plutons of similar size. The points that depart most from the low albite corner represent values whose coexisting K-rich feldspar is orthoclase (Fig. 5.4b). It is known that the rate of ordering is very efficient in pure $\text{NaAlSi}_3\text{O}_8$ (Ribbe 1983, Parsons 1978). The disorder illustrated in Figure 5.4b reflects the presence of small amounts of Ca in the structure. In other words, the samples in which the albite departs most from the low albite corner initially contained a ternary alkali feldspar (i.e., calcic sanidine), all traces of which are now gone.

Fig. 5.4: a) Plot of $c(\text{\AA})$ versus $b(\text{\AA})$ for K-rich alkali feldspars. The inset shows areas of the plot enlarged in Figs. 5.4a, b. AA = analbite, HS = high sanidine, LA = low albite, and LM = low microcline (coordinates of Kroll & Ribbe 1983). ECPA = expected composition of primary alkali feldspar (sanidine₃₃) that crystallizes from a melt. The monoclinic (orthoclase) to triclinic (microcline) inversion occurs at about 450°C.

b) Enlargement of the b - c plot for the albite-rich composition (inset shows area of b - c plot enlarged). The sodic feldspar (coexisting with orthoclase) is relatively less ordered with significant overlaps between cone sheets and ring dyke units, whereas the albite in plutonic rocks (coexisting with microcline) is completely ordered. Error bars show average standard deviations σ_{av} along the respective axes. The maximum error bars (σ_{max}) are 2 to 3 times the symbol size.



In Figs. 5.4a & 5.6a:

◊ Low Microcline

◆ Intermediate Microcline

◆ Microcline coexisting with orthoclase

■ Orthoclase coexisting with microcline

■ Orthoclase only

In Figs. 5.4b, 5.5: symbols represent albite coexisting with above K-feldspar

The electron-microprobe data obtained on the cryptoperthitic phenocrysts do not completely agree with the XRD data. The overall range of compositions extends from Or_1 to $Or_{95.9}$, with most of the compositions in the intermediate range of the series. These intermediate compositions, for which there is no evidence in the XRD patterns, are considered artifacts of the analytical methods: the electron beam (5 - 8 μm in diameter) samples both sodic and potassic domains owing to their fineness, and gives an aggregate composition dependent on the volume proportions of the two types of domain. The electron microprobe data are useful, however, in assessing An content; the cryptoperthite contains up to 3 mol.% An, and most results are in the range $An_{0.4-1.9}$. Such small amounts are not sufficient to cause the spread of points in Figure 5.4b. It is thus likely the phenocrysts once contained more Ca, but this has been removed during subsolidus, hydrothermally mediated modifications leading to exsolution and ordering. The disorder in the sodic phase may well illustrate a metastably preserved pattern of distribution of Al over the four available T sites in the structure.

The An contents recorded in the electron-microprobe data correspond to those inferred from the plot of β^* versus γ^* ($An_{0.5}$) for the sodic plagioclase (Fig. 5.5).

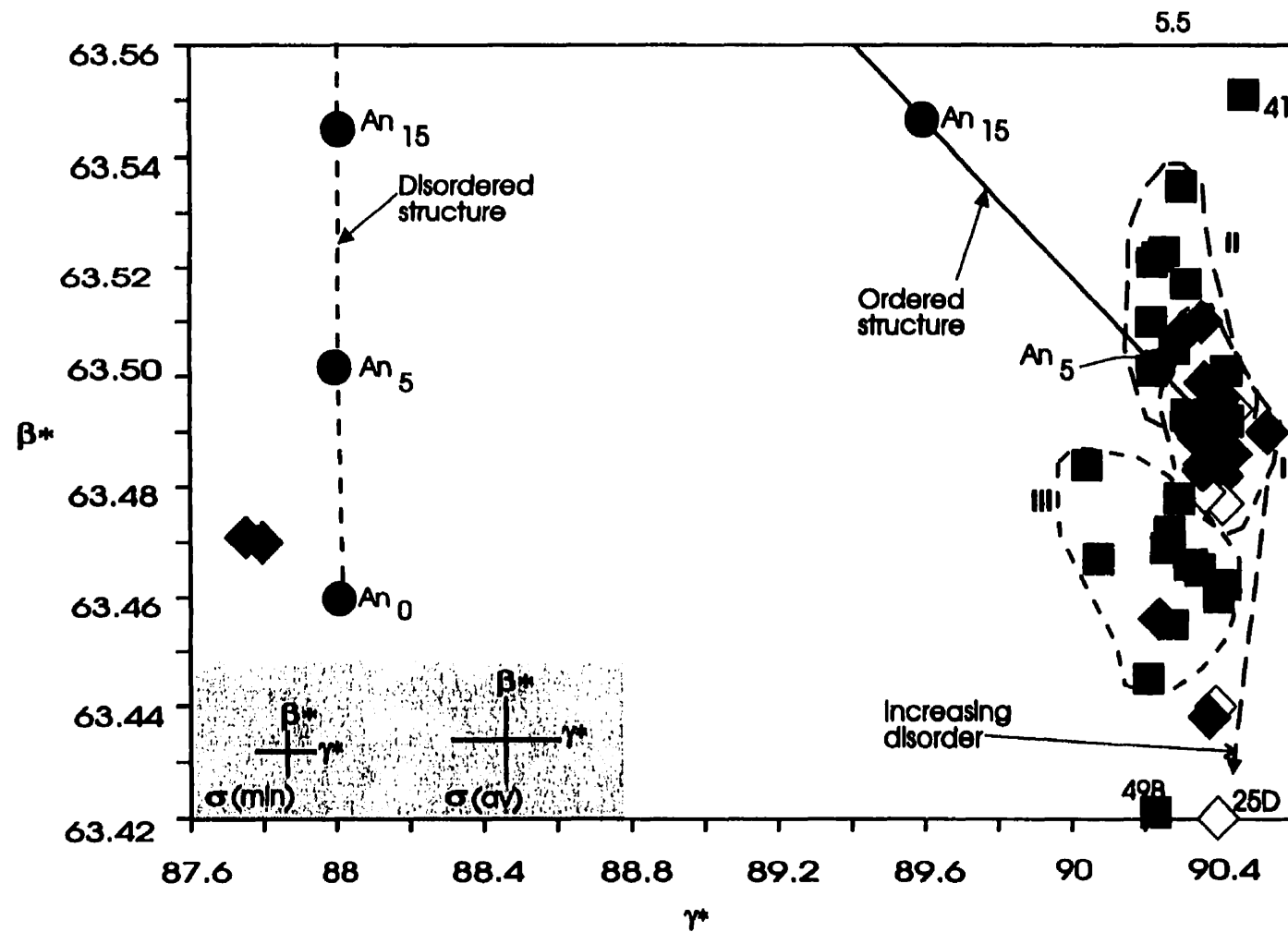
5.5.1 Orthoclase

A portion of the crystal structure of orthoclase is presented in Appendix V.2a. Typically, in orthoclase, t_1 is in the range 0.37 to 0.5, which is consistent with an average monoclinic structure with the Al distributed equally into the two T_1 sites. Ideally the Al ultimately "migrates" to the two T_1O site, and thus the Al occupancy of each t_1 site provides an indication of deviations from the disordered sanidine structure.

In the Kwadonkaya complex, orthoclase ($0.37 \leq t_1 \leq 0.46$) is the principal K-feldspar in the rhyolite porphyry (49) and granite porphyry (41) of the Buri ring dyke, in felsite porphyry

Fig. 5.5: Plot of β^* versus γ^* values for the sodic plagioclase with An contents indicated. The loci for ordered and disordered structures are indicated. The data plot mostly perpendicular to the line for ordered structures. This reflects structural states of rapidly cooled Na-plagioclase typical of subvolcanic and epizonal plutonic rocks. Sample 49B contains sodic plagioclase (in the trachyandesite enclave) with the most disordered structure, and indicates that the feldspar likely was Na-metasomatized at high temperature. Data groups I to III are discussed in the text.

Symbols are the same as in Fig. 5.4b, as they apply to the sodic phase coexisting with the K-rich feldspars shown in Fig. 5.4a.



(40D, 61A) and granite porphyry (40A, 64A) in cone sheets, some samples of granite porphyry (68B) in the Jubga-Ziem ring dyke, and the hematite-speckled biotite granite (22) in the Jubga-Ziem center. These all come from small bodies peripheral to the main complex (Fig. 5.1), with the exception of sample 22. In all cases, the orthoclase in the phenocryst and the matrix is very close to KAlSi_3O_8 , indicating efficient exsolution down to low temperatures. The t_1 value is in the range 0.39-0.44, i.e., close to the upper limit achieved by orthoclase before conversion to microcline (Kroll & Ribbe 1983). Sample 41 (hedenbergite fayalite granite porphyry) contains a matrix orthoclase that is slightly better ordered ($t_1 = 0.44$) and, in cases, purer ($N_{\text{Or}} = 1.0$) than the phenocryst ($t_1 = 0.40$, $N_{\text{Or}} = 0.95$). In sample 68B (hornblende biotite granite porphyry), the orthoclase in the phenocryst has a $t_1 = 0.40$, whereas that in the matrix has $t_1 = 0.43$. Such examples suggest that the effective grain-size of the K-feldspar was important in some instances as a control of degree of order attained, the finer fraction containing the better-ordered K-feldspar. This finding can be best explained by considering the proximity of surfaces in controlling the extent of feldspar-water interaction, leading to exsolution and ordering.

In other examples, the orthoclase in phenocrysts and matrix has a similar degree of order (and N_{Or}): e.g., rhyolite porphyry 49 (phenocryst: $t_1 = 0.43$, $N_{\text{Or}} = 0.99$, groundmass: $t_1 = 0.44$, $N_{\text{Or}} = 0.98$); hedenbergite hastingsite granite porphyry 64 (phenocryst; whitish core: $t_1 = 0.39$, $N_{\text{Or}} = 0.97$, yellowish rim: $t_1 = 0.40$, $N_{\text{Or}} = 0.96$, groundmass: $t_1 = 0.41$, $N_{\text{Or}} = 0.99$), as is also the case in some samples of felsite porphyry, e.g., 61A (mesoscopic phenocrysts: $t_1 = 0.43$, $N_{\text{Or}} = 0.98$, groundmass: $0.42 \leq t_1 \leq 0.46$, $0.95 \leq N_{\text{Or}} \leq 1.0$). Where cleavages are open in phenocrysts, perhaps owing to slight deformation during cooling, the size of the domains could in fact be similar to those encountered in the matrix, and the degree of Al-Si order would not differ from that attained in the matrix.

One other interesting style of variation is documented in sample 40A (hastingsite granite

porphyry), in which the degree of order of the orthoclase phenocrysts varies from the core to rim: the greenish core ($t_1 = 0.40$, $N_{Or} = 0.91$), is less ordered and more sodic than the white rim ($t_1 = 0.44$, $N_{Or} = 0.98$). The lower degree of order of the greenish core compared to the whitish rim could reflect a catastrophic coarsening of the perthite and improved ordering at the margin, and limited extent of recrystallization in the core of the phenocryst owing to relative impermeability. The white rim also contains traces of intermediate microcline. The matrix orthoclase ($t_1 = 0.42$, $N_{Or} = 0.97$), on the other hand, has a degree of order and composition intermediate between those of the core and rim.

Sample 49 also is interesting because it contains enclaves of trachyandesite, itself porphyritic. The matrix orthoclase ($t_1 = 0.43$, $N_{Or} = 0.97$) is the same as that in the host rhyolite porphyry, but the phenocryst and matrix plagioclase, presumably once oligoclase-andesine, now is transformed to albite. Its An content is estimated (from $\beta^* - \gamma^*$, Fig. 5.5) to be in the compositional range $An_{0.5}$.

One of the surprising occurrences of orthoclase is in the hematite-speckled granite found around miarolitic cavities (sample 22), which is from a peripheral region in the Jubga-Ziem center (Fig. 5.1). The orthoclase ($t_1 = 0.46$, $N_{Or} = 0.92$) is the most ordered one of all the samples studied. The orthoclase does contain small domains of intermediate microcline, but with an insufficient number of diffraction peaks to characterize it fully. Despite the clear signs that a fluid was available, the conversion to microcline clearly was arrested at an early stage.

5.5.2 Orthoclase coexisting with microcline

Orthoclase coexists with microcline in fayalite syenogranite (58), some samples of hornblende biotite granite porphyry (56), biotite granite porphyry (38, 82), hypersolvus fine-grained dykes (83), and in the drusy parts of biotite granite (37). Most of the orthoclase-microcline

pairs are found in the Jubga-Ziem center.

The orthoclase co-existing with microcline has a somewhat higher degree of order ($0.41 \leq t_1 \leq 0.45$) than the orthoclase occurring as the only K-feldspar (Fig. 5.4a). The documented range of Al-Si order, and compositions of the K-feldspar domains ($0.95 \leq N_{Or} \leq 0.97$) are narrow compared to cases in which the orthoclase occurs as the only K-feldspar (Fig. 5.4a).

The orthoclase in fayalite syenogranite, fine-grained granite dykes, and hematite-speckled biotite granite contain traces of intermediate microcline. These samples thus preserve the incipient stages of monoclinic - triclinic inversion. The N_{Or} contents of the orthoclase and coexisting microcline do not show any systematic differences between subvolcanic and plutonic rocks.

5.5.3 Intermediate microcline

A portion of the crystal structure of microcline is shown in Appendix V.2b, c. In the ideal case, low microcline, all Al has migrated to the T_1o site, and the occupancy of the tetrahedral sites is more clearly distinguished (Appendix V.2b), with $t_{1o} = 1.0$, and $t_{1m} = t_{2o} = t_{2m} = 0$. In the case of microcline, the Al occupancy is best described by t_{1o} , as inferred from the cell dimensions.

In this study, structurally intermediate microcline is defined as that with a t_{1o} value between 0.8 and 0.9. It occurs along with orthoclase in hornblende biotite granite porphyry (56), fayalite granite porphyry (58), biotite granite porphyry (38, 82), all in the peripheral Jubga-Ziem ring dyke, and 83 (dykes in hypersolvus biotite granite). The intermediate microcline also occurs as the principal K-feldspar in the reddened granite porphyry (66) of the Buri ring dyke, coarse-grained biotite granite (26A), hypersolvus biotite granite (83); it coexists with low microcline in porphyritic biotite granite in both the Panshanu center (1B), and the Jubga-Ziem center (36A).

The intermediate microcline coexisting with orthoclase in rock units of the Jubga-Ziem ring dyke displays a narrow range in t_{1o} , N_{Or} , and obliquity (Δ) (defined in Table 5.2). These

parameters fall in the range: $0.82 \leq t_{10} \leq 0.84$, $0.90 \leq N_{Or} \leq 0.99$, and $0.65 \leq \Delta \leq 0.81$, and overlap with those of intermediate microcline in other associations (Fig. 5.4a). The values of N_{Or} do not show a trend with the nature of the host rock. The intermediate microcline in the fine-grained dykes (83) is characterized by the lowest t_{10} (0.82), and N_{Or} (0.90) values. This most likely is due to more rapid cooling in these small bodies, which implies that the host rocks had cooled by the time the intrusion occurred. Intermediate microcline also predominates in the hydrothermally altered granite porphyry in the Buri ring dyke (66). There is no difference between the cream-coloured core ($t_{10} = 0.86$, $N_{Or} = 0.96$, $\Delta = 0.78$), and red rim of the phenocrysts ($t_{10} = 0.86$, $N_{Or} = 0.95$, $\Delta = 0.81$). The occurrence of intermediate microcline rather than orthoclase suggests relatively high fluid:rock ratios and accompanying intense hydrothermal activity in a localized area along the ring dyke. Had the recrystallization not been brief, however, a better ordered microcline should have resulted.

In the textural variants of biotite granite, the intermediate microcline shows values of t_{10} , N_{Or} , and Δ that are consistently similar in samples from both the margins and central part of the pluton. There does not seem to be any systematic variation among samples from the margin of the intrusive bodies, such as coarse-grained biotite granite (26) and porphyritic biotite granite (1B), whose range of t_{10} , N_{Or} , and Δ are: $0.81 \leq t_{10} \leq 0.90$, $0.98 \leq N_{Or} \leq 1.00$, $0.65 \leq \Delta \leq 0.82$. The samples of biotite granite from the central part of the complex also have a composition and degree of order within the above ranges. For example, the samples of porphyritic biotite granite (35A, 36A) from the central part of the complex have microcline with the range: $0.83 \leq t_{10} \leq 0.88$, $0.96 \leq N_{Or} \leq 0.98$, $0.71 \leq \Delta \leq 0.76$. The composite body of biotite granite thus seems to have experienced a rapid (uniform) rate of cooling and extent of postmagmatic equilibration, with local variations only in areas that experienced volatile loss to yield drusy biotite granites.

5.5.4 Low microcline

Low microcline in this study is defined as a K-feldspar with t_{10} between 0.91 and 1. Low microcline occurs in the drusy biotite granite (30), in miarolitic cavities in the coarse-grained biotite granite (25D), the medium-grained biotite granite (22), and in drusy biotite granite (37). Also, low microcline occurs along with intermediate microcline in the matrix (1B), and either as discrete phenocrysts (36A), or as overgrowths on intermediate microcline (36A).

The low microcline in the groundmass of the porphyritic biotite granite (1B) (the phenocrysts are intermediate microcline) is the least ordered and purest ($t_{10} = 0.91$, $N_{Or} = 1.0$, $\Delta = 0.90$), whereas the phenocrysts of low microcline in the host rock of greisens (36A) have relatively higher t_{10} and N_{Or} ($0.95 \leq t_{10} \leq 0.98$, $0.98 \leq N_{Or} \leq 0.99$, $0.89 \leq \Delta \leq 0.90$). In this example, the overgrowths on the phenocrysts, possibly of hydrothermal origin, have an even higher degree of order ($t_{10} = 0.98$).

The low microcline in biotite granite sample 22 ($t_{10} = 0.94$, $N_{Or} = 0.92$), which is from close to a miarolitic cavity, is similar in degree of order (t_{10}), but has a more sodic composition than in that found in the drusy facies (Fig. 5.4a). The latter is best exemplified by the low microcline in sample 30 ($t_{10} = 0.93$, $N_{Or} = 0.99$, $\Delta = 0.90$), which has similar t_{10} but a slightly higher N_{Or} , indicative of a more extensive equilibration to a low temperature. The low microcline in samples 37 and 26D have both t_{10} and N_{Or} ($t_{10} = 0.95$, $N_{Or} = 0.98$, $0.90 \leq \Delta \leq 0.93$) more like the one in sample 30. The low microcline in the drusy cavity of biotite granite (37) contains traces of orthoclase, which is similar to observations on some of the microcline coexisting with orthoclase. The drusy cavities (in 37) also contain hexagonal books of mica, topaz, quartz, and kaolinite. The occurrence of kaolinite (characterized by XRD studies) and topaz in the drusy cavity is consistent with the local importance of highly acid (F-rich) fluids that had relatively short residence times, and limited interaction with the adjacent feldspar, with which it presumably was

not stable.

5.5.6 Sodic plagioclase

Usually, albite orders easily and completely in the presence of an aqueous fluid. It is therefore surprising that the b - c plot (Fig. 5.4b) shows the albite to be incompletely ordered, although the apparent mol fraction of orthoclase (N_{Or}) is mostly less than 0.25. The albite coexisting with orthoclase is relatively more disordered than that co-existing with microcline (Fig. 5.4b). The b - c plot does not give an indication of An-contents, which is known to affect the extent to which Al-Si order can proceed. Therefore, to further explore the causes of incomplete order of the albite, a plot of β^* versus γ^* (Fig. 5.5) is used (cf. Smith 1974). Three patterns of distribution can be discerned in Figure 5.5 (labelled **I**, **II**, and **III**): **I**) in plutonic rocks, the albite coexisting with microcline plots mostly close to the line marking the locus of ordered structures (Fig. 5.5). Such a distribution is explained by the relatively slow cooling rates in most of the plutonic rocks (and microcline-bearing) rock units that also underwent various extents of rock-fluid interaction. The exceptions are sodic plagioclase coexisting with orthoclase in hypersolvus biotite granite (and drusy biotite granites that lost volatiles early), because they experienced little or no rock-fluid interaction. **II**) This is comprised of the sodic plagioclase coexisting with orthoclase, and plot close to the line defining ordered structures, but mostly above such a line. This data set is also characterized by $An_{0.5}$, and overlaps slightly with data for albite coexisting with microcline. The data points plotting above the ordered structure are not fully explained by the cooling history or An-contents. However, the most deviant sample of sodic plagioclase in this class is within two standard deviations of the β^* value of pure ordered albite (ca. 63.49°), and such data may therefore be due to higher-than-normal standard errors that reflect relatively small number of peaks or the presence of coherency strain further discussed below. **III**) The third grouping of data consists of

sodic plagioclase coexisting with orthoclase plotting toward the higher degree of disorder but still characterized by $An_{0.5}$. This group of data points may represent more strongly quenched rocks or sodic plagioclase that initially contained more calcium and was subsequently albitized. Similar departures from the locus of ordered plagioclase, in the direction of decreasing β^* , are attributed by Martin (1984) and Schandl et al. (1986) to Al-Si disorder, with the departures increasing with decreasing An-content in albitized plagioclase. The disorder in these cases is "residual", reflecting the former distribution of Al in a more Ca- and Al-rich primary plagioclase that is now more sodic than it once was. It is only in the case of albite in trachyandesite enclave (sample 49B in Fig. 5.5) that magma mixing can be invoked to explain the more calcic original composition. Most of the other samples are from rock types that bear no enclaves.

5.6 Strain in the K-feldspars

The index of strain (Δa) (Stewart & Wright 1974, Kroll & Ribbe 1983) gives an indication of the strain state in alkali feldspar (where $\Delta a = a_{obs} - a_{est}$, with terms defined in footnotes in Table 5.2). The early stages of perthite formation at Kwandonkaya likely involved the formation of coherently intergrown coexisting K-rich and Na-rich domains (Appendix V.3). Elastic coherency strains arise from the change in cell dimensions as a function of composition (Yund & Tullis 1983); the changes in b and c amount to about 1%, whereas the change in a is about 5%.

The Δa value of coexisting K- and Na feldspars in the Buri ring dyke, Jubga-Ziem ring dyke and biotite granite (Table 5.2) suggests that the feldspars in these rock units are completely unstrained. In samples 49A, 41, 64, and 40A (rhyolite porphyry and granite porphyry in cone sheets and ring dykes), the phenocrysts, and rarely the matrix, have Δa values between 0.05 and 0.09, i.e., barely beyond the range (± 0.05) taken as being indicative of unstrained structures. Thus

the original sanidine_{ss} that first exsolved according to the coherent solvus (inset in Appendix V.3), has been efficiently converted to strain-free assemblage owing to what Parsons (1978, 1981) has called "catastrophic coarsening" of the intergrowth. The feldspar structures now have compositions that are "dictated" by the strain-free solvus (Appendix V.3). Accordingly, the compositions mentioned above are inferred from equations that apply to strain-free intergrowths.

5.7 Discussion

The sequence of discussion includes a brief interpretation of textural features displayed by grains of alkali feldspar intergrown with other minerals (particularly quartz). This is followed by a discussion of the implications of the variations in the degree of Al-Si order and composition state of the K-feldspar in relation to the magmatic to postmagmatic evolution of their host rocks.

5.7.1 Textural features

Mantling of microperthitic orthoclase by turbid albite-rich feldspar occurs in phenocrysts within both cone sheets that do not contain enclaves or other evidence of mixing, and ring dykes. The foregoing, along with the fact that the mantle contains, in some instances, Na-rich plagioclase (low albite) that is very poor in Ca, make it doubtful that the mantling was produced by mixing, as proposed by Hibbard (1981). The turbid mantle on orthoclase microperthite phenocrysts is highly irregular and pervades fractures and cleavages, with the turbid margin making up from 5 to 80% of the phenocrysts. SEM-BSE images (Fig. 5.3a) reveal the presence of a clear cryptoperthitic core and a margin consisting of relatively coarse and continuous microperthite. The texture can be explained by the following reconstruction: i) The sanidine_{ss} phenocrysts passed through a stage of coherent exsolution. ii) The outer parts of the grains then experienced "catastrophic" coarsening of perthite and loss of coherency due to incipient rock-fluid interaction.

This is comparable to observations by Parsons (1981) in some of the sections of Klokken feldspar, in which braid perthite abruptly coarsens to irregular patch perthite around the margins. iii) The grains become turbid, according to Worden et al. (1990), owing to the presence of myriads of tubular micro-inclusions, either fluid-filled micropores or sites of previous fluid inclusions, and is associated with coarsening of microperthite and the development of subgrains. According to Martin and Lalonde (1979), the fluid inclusions (vacuoles) are interpreted to result from volume decrease undergone by the high-temperature feldspar due to anisotropic thermal contraction, exsolution, nucleation of strain-free Na-rich and K-rich domains, increased degree of Al-Si order, and removal of matter by dissolution. The latter suggests the importance of secondary porosity in promoting the recrystallization. If water-rock interactions occurred at 450°C, then the turbid feldspar is expected to be orthoclase (as is the case with the Kwandonkaya subvolcanic rocks in this study), whereas turbid microcline should result from protracted interaction of rocks with water at less than 450°C. The reddening in granite porphyry (sample 66) bearing microcline is interpreted to be due to hematite flakes coating vacuole walls, suggesting also interaction with oxidizing fluids that led locally to the complete breakdown of ferromagnesian minerals (Martin & Lalonde 1979, Lalonde & Martin 1983a).

Relatively low undercoolings at initial stages resulted in the formation of phenocrysts, typically seen in all subvolcanic rocks, owing to the low density of nucleation and high rates of growth (Swanson 1977). Subsequently, higher undercoolings of partly crystallized melt at crystal:melt ratios of 0.3 to 0.6 produced the porphyries. In water-undersaturated melts, euhedral (faceted) crystals are stable within 40°C of the liquidus (Fenn 1977), whereas the spherulitic texture in the groundmass is indicative of greater undercoolings. Prevailing low pressures, in cone sheets and ring dykes, favour the loss of volatiles from a magma, thereby increasing viscosity and compounding the effects of rapid cooling (Brandeis & Jaupart 1987, Toramaru 1991), and also

enhancing the chances of forming the spherulitic texture. The latter results from the efficient dissipation of latent heat generated by a freely growing crystal away from the interface, causing the temperature to decrease away from the crystal. Protrusions from the crystal grow faster than the rest, thereby resulting in cellular morphology, leading to the spherulitic texture. The spherulitic phenocrysts in the felsite and granite porphyries were produced by crystallization from a melt (cf. Fig. 2.3c). In felsite porphyry, the unusual development of granophyric phenocrysts with sectors (Fig. 2.3c) suggests rapid simultaneous crystallization of nonfaceted quartz and cellular alkali feldspar (Carstens 1983). In the groundmass of cone sheets, some of the granophyric texture that produces a plumose pattern could arise from forced simultaneous crystallization of a relatively large proportion of the melt (Fenn 1977, Swanson 1977, Carstens 1983), or by devitrification of a glassy groundmass. It seems that the spherulitic texture in most of the subvolcanic rocks was produced by forced simultaneous crystallization. This occurs owing to an imbalance between the growth rate of the feldspar and the diffusivity of silica in the bulk melt, which creates a silica-enriched boundary layer that in turn causes the interface to degrade from planar to cellular. Upon further growth of the cellular interface, the grooves between adjoining cells are greatly enriched in SiO_2 , and this enrichment causes the nucleation and growth of quartz along with the feldspar (Fenn 1977). Fenn emphasized that simultaneous growth, in this case, is not necessarily due to eutectic or cotectic crystallization, but rather is caused by the sluggishness of diffusion in the boundary layer adjacent to the interface. Although it is not possible to clearly distinguish spherulitic textures produced by devitrification, the pervasive spherulitic groundmass with no defined feldspar morphology, observed in rare cases, is inferred to be a product of devitrification.

The aggregation of phenocrysts of alkali feldspar ($\pm$ pyroxene, quartz, or amphibole) to produce a glomeroporphyritic texture in the granite porphyries (Fig. 2.4d) is termed synneusis. Synneusis is absent or arrested at different stages in rhyolite porphyry and felsite porphyry,

whereas it is invariably present in more slowly cooled granite porphyries. Phenocrysts in a synneusis relationship invariably have a planar interface and, according to Kirkpatrick (1975), resulted from interface-controlled growth at low undercoolings. Melt convection or some type of differential motion of crystals is required for synneusis to occur, and the driving force for the adherence of crystals is the decrease in surface energy brought about by the conversion of two crystal-liquid interfaces into one (Fenn 1977). Synneusis seems to be a product of inhomogeneous nucleation and preferential drifting together of phenocrysts of similar composition.

In the least altered plutonic biotite granite, the different textural variants are characterized by a hypersolvus assemblage of feldspars replaced by a subsolvus assemblage. This indicates magmatic crystallization at temperatures above the crest of the alkali feldspar solvus, consistent with low confining pressure. Subsequently, the initial single feldspar crystallized was transformed to microcline + albite. Within the albitized biotite granite, the subsolvus assemblage is essentially metasomatic, whereas in post-joint greisens, alkali feldspar is rarely preserved.

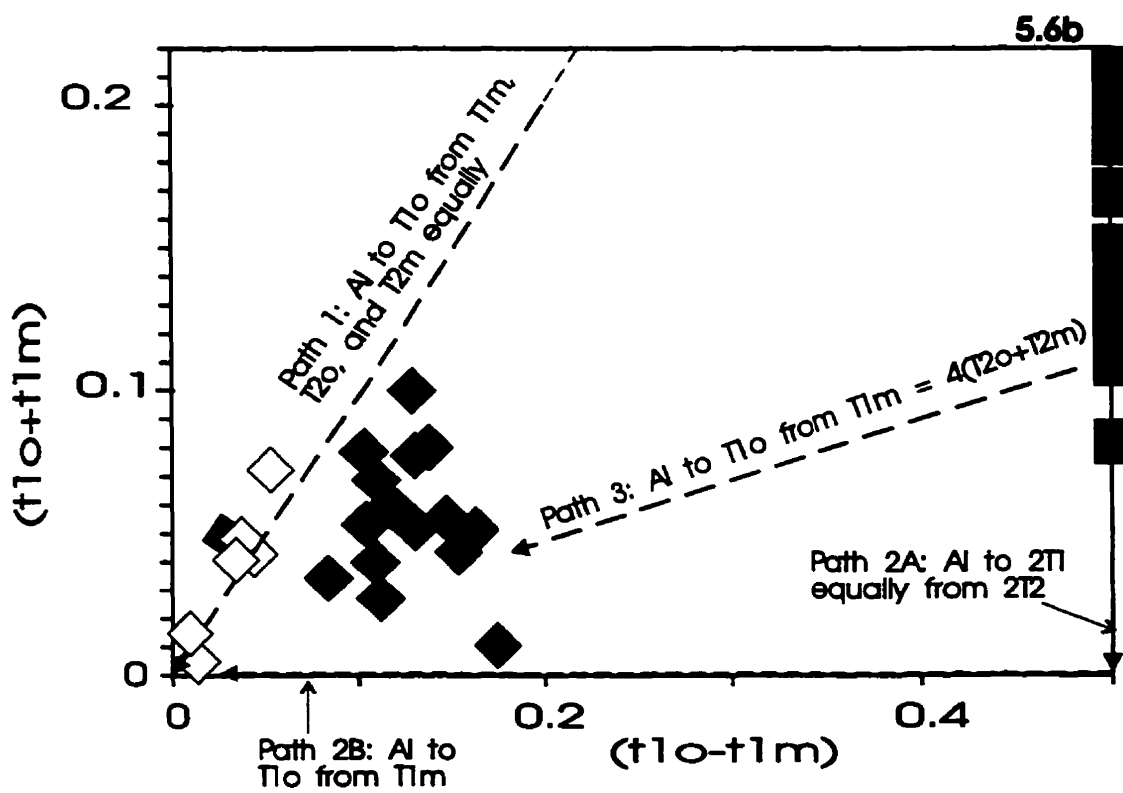
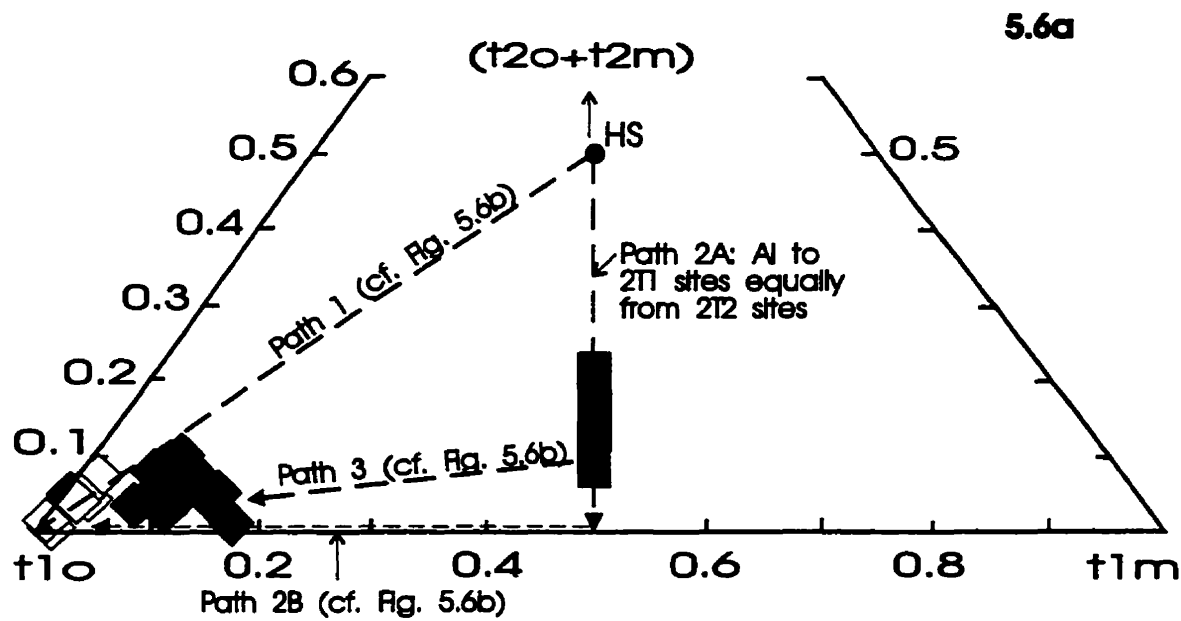
Drusy or miarolitic cavities in marginal part of hornblende biotite granite porphyry and biotite granite indicate volatile build-up and loss.

5.7.2 Implications of observed structural states of K-feldspar

Sanidine_{ss} that crystallized from melt ordered firstly to orthoclase along path 2A (Figs. 5.6a, b), in which Al "moves" to $2T_1$ sites equally from the two $2T_2$ sites. This is one limb of the two-step ordering paths, and no K-feldspar is documented that lies along the second part (path 2B) of the ordering path. Microcline lies along ordering path 3, which is intermediate between path 1 and path 2B. Along path 3, Al migrates to T_1o from the T_1m site at a rate four times the sum of migration from T_2o and T_2m (i.e., $T_1o - T_1m = 4(T_2o + T_2m)$). Ordering to "ideal" low microcline was nearly attained in drusy cavities within biotite granite (Fig. 5.6b).

Fig. 5.6: Plots illustrating ordering paths for K-rich alkali feldspars: **a)** triangular diagram of $t_1o - t_1m - (t_2o + t_2m)$ for K-feldspars. The different paths or Al-Si order are indicated, with further explanations in the text. **b)** $(t_1o - t_1m)$ versus $(t_1o + t_1m)$ for K-feldspar.

Symbols are the same as in Fig. 5.4a.



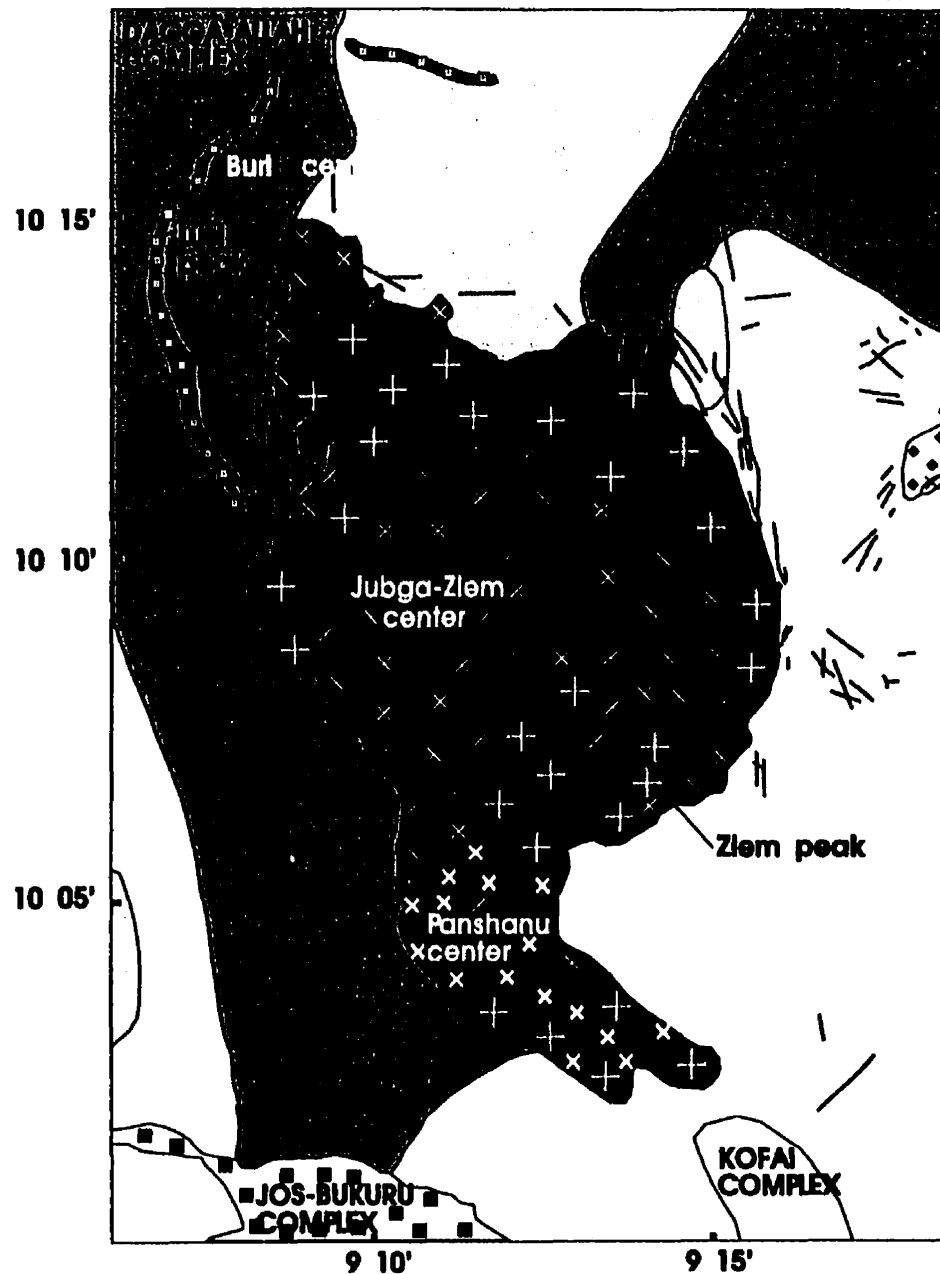
Sanidine_{ss} crystallized directly from the melt, in subvolcanic rocks and hypersolvus biotite granites, and first underwent coherent exsolution to produce crypto- to microperthitic orthoclase + albite. Evidence for this early stage has been erased by later fluid-dominated recrystallization. The relatively low values of the index of strain ($0 \leq \Delta a \leq 0.09$) indicate that the assemblage then recrystallized to incoherently intergrown minerals, i.e., the strain-free solvus applies. In the biotite granites, equilibration progressed further to yield intermediate (+low) microcline + albite.

The distribution of the K-rich alkali feldspar types is summarized in Fig. 5.7, which shows the Buri ring dyke and cone sheets (mostly small-volume intrusions) that underwent high undercoolings to be characterized by orthoclase, whereas subvolcanic rocks in the Jubga-Ziem ring dyke to contain orthoclase + microcline, possibly because of their relatively larger size and interaction with fluids exsolved from the biotite granite. The larger intrusions of biotite granite, which also display evidence of volatile build-up and various extents of fluid-rock interaction, contain mainly microcline except in some localized occurrences of hypersolvus granite and drusy variants.

The persistence of orthoclase either as the main K-feldspar in rock units in the Buri ring dyke, cone sheets, and hematite-speckled biotite granite or orthoclase + imperfectly ordered microcline in rock units of the Jubga-Ziem ring dyke and drusy biotite granite, represents a clear case of what Ribbe (1983) described as a metastable state. Such orthoclase represents a state of Al-Si order that is stranded kinetically during the sanidine-to-microcline inversion dominantly because of sudden loss of the orthomagmatic fluid, and also perhaps because of initially higher Ca-contents of the sanidine_{ss}. Orthoclase persists in the cooling interval in which microcline becomes a stable phase because fluids were not present (or sufficiently plentiful) to facilitate the coarsening of triclinic domains. The fact that the orthoclase is documented in both the phenocryst and groundmass of pristine units in the Buri ring dyke and cone sheets (Figs. 5.4, 5) supports the

Fig. 5.7: The distribution of alkali feldspars in major rock units of the Kwandonkaya complex. This shows the Buri ring dyke and cone sheets (mostly small-volume intrusions), which underwent high degree of undercooling, to be characterized by orthoclase, whereas subvolcanic rocks in the Jubga-Ziem ring dyke contain orthoclase + microcline, possibly because of their relatively larger size and interaction with fluids exsolved from biotite granite. The larger intrusions of biotite granite contain mainly microcline, except in some localized occurrences of hypersolvus granite and drusy variants.

The symbols for rock types are the same as in Fig. 5.1.



Basalt dyke

MESOZOIC GRANITES

G: Grelsens

Microcline (unimodal/bimodal)

■ Fine- to medium-grained biotite granite

⊕ Coarse-grained biotite granite

⊗ Porphyritic biotite microgranite

Orthoclase + microcline

■ Hornblende biotite-, biotite-granite porphyry, +/- fayalite syenogranite

Orthoclase only

■ Hedenbergite fayalite granite porphyry

■ Rhyolite porphyry

⊕ Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

□ Biotite hornblende granite

■ Gneissic granite

□ Migmatite

1 0 2 4 6 8 km

hypothesis of relatively low levels of water and other volatile components inferred for these rock units, both at the magmatic and postmagmatic stages (ca. $0.3 \leq a_{\text{H}_2\text{O}} \leq 0.5$, Chapter 2). The preservation of orthoclase may also be explained by the fact that the original alkali feldspar was relatively enriched in Ca ($> \text{An}_2$), and thus could not become completely ordered (Kroll & Ribbe 1983). The electron-microprobe data, and estimates from the plot of β^* versus γ^* (Fig. 5.5), show that most of the sodic feldspar consists of $\text{An}_{0.5}$. Furthermore, Figure 5.5 suggests that the sodic feldspar might have lost its Ca during exsolution, and thus preserves the ghost-structures inherited from a more An-rich (i.e., ternary) alkali feldspar. The occurrence of microcline only in the reddened granite porphyry (within rocks bearing only orthoclase) indicates that the transition from disordered sanidine_{ss} to low microcline + albite was aided by localized occurrence of high fluid:rock ratios, in loci of hydrothermal alteration. The completely turbid character of the feldspar in the reddened granite porphyry could be a side effect of catastrophic coarsening of perthite.

The preservation of orthoclase and microcline in fayalite syenogranite, hornblende biotite granite porphyry, biotite granite porphyry, drusy biotite granite, and hypersolvus biotite granite dyke indicates an arrested monoclinic-to-triclinic inversion of K-feldspar in the sequence: (sanidine_{ss} $\rightarrow$) orthoclase + albite $\rightarrow$ intermediate microcline + albite. This group of rocks progressed further than those discussed above owing to their slower cooling rates as well as higher degrees of volatile build-up.

The rock units bearing microcline and free of orthoclase represent the highest degree of low-temperature interaction with an aqueous fluid. The rock units with microcline suggest local variations in the extent of ordering involving differences in grain size or in permeability along cleavages and in density of fluid inclusions, thereby making the effective grain-size very much smaller than the overall dimensions of the grains (Parsons 1981). Most of the samples of biotite granite contain microcline as the only K-feldspar. The microcline perthite is a "pseudomorph" of

the orthoclase perthite, and indicates that feldspar-fluid interaction, in fact a series of solution-redeposition reactions, proceeded further, perhaps due to an increase in magmatic water (+Cl, F) in the environment as the rocks cooled. The extent of Al-Si order attained is known to depend on many factors, including composition of the fluid phase and amount of fluid (water-rock ratio).

It has been well documented in the peralkaline Strange Lake granitic complex in Quebec-Labrador (Nassif 1993), that extreme enrichment in Fe, Rb, HFS and LILE has affected the feldspar composition so as to shift such compositions beyond the microcline apex in the b-c plot. Because enrichments in these elements are considerably lower in the metaluminous-peraluminous Kwadonkaya suite, the suite of K-rich alkali feldspar are mostly contained within the Analcite - High sanidine - Microcline - Albite quadrilateral in the b-c plot (Fig. 5.5a)

It is apparent that K-feldspar phenocrysts in subvolcanic rocks and hypersolvus biotite granites crystallized from the melt (as sanidine₁₀₀) possibly at >900°C, and were later equilibrated to orthoclase $\pm$ microcline (+ albite). According to Stewart and Wright (1974), the range of crystallization of orthoclase (with $2x_1 = 0.8 \pm 0.1$) is 750 to 625°C, whereas the upper stability of maximum microcline occurs at about 450°C. The orthoclase and intermediate microcline probably represent stranded structures that failed to "keep up" with decreasing temperature. The molar fraction of orthoclase (N_{Or}) varies within a very limited range ($0.91 \leq N_{Or} \leq 1.0$), with no systematic variation in XRD-determined compositions between subvolcanic and plutonic rocks (Table 5.2). Compositionally, the feldspar (orthoclase) has equilibrated successfully to the field of microcline (Appendix V.3). From the composition-temperature (X-T) diagram of the strain-free solvus at 1 kbar, it is estimated that the temperature of "closure" of rocks bearing orthoclase only, in terms of N_{Or} , is between 450°C to 300°C. Thus, Na and K exsolution continued to relatively low temperatures (cf. Parsons 1981), long after the cessation of Al-Si ordering, and therefore a metastable extension of the stability field of orthoclase (within the field of microcline) is shown

(Appendix V.3). The composition of microcline also varies within a limited range ($0.95 \leq N_{Or} \leq 1.0$) and, as shown in Appendix V.3, suggests a temperature of equilibration of between 450 and 200°C. The observation that there is variation of N_{Or} and degree of order within individual phenocrysts, and in phenocryst-matrix pairs within a single sample of rock, or in samples bearing both orthoclase and microcline, suggests the absence of a pervasive fluid phase (i.e., the fluid:rock ratio was low) needed to effect complete or homogeneous transformation of the K-feldspar, particularly in subvolcanic rocks. Alternatively, this may reflect variations in "effective" grain size for fluid-rock interaction. Plutonic rocks, on the other hand, contain mainly microcline that seems to have "kept up" in terms of both Al-Si order and exsolution of Na-rich phase, presumably owing to slower cooling, more voluminous fluid, and enhanced permeability.

5.8 Conclusions

1) A variety of textural features are displayed by feldspars in the subvolcanic and plutonic rocks, which include synneusis aggregates, spherulitic intergrowths, and marginal turbidity of grains. Whereas synneusis and spherulitic textures are inferred to reflect the crystallization history of the rock units, marginal turbidity and drusy facies arose from late-magmatic to subsolidus processes. For example, the spherulitic texture indicates initially slow growth initially, succeeded by high undercoolings and forced simultaneous crystallization of quartz and alkali feldspar, which produced phenocrysts of sanidine_{ss} (now orthoclase microperthite) surrounded by a microcrystalline, spherulitic to granophyric groundmass. Marginal turbidity in phenocrysts likely developed as result of catastrophic coarsening of perthite and loss of coherency in the subsolidus range, in some instances accompanying volatile loss.

2) Despite the fact that most subvolcanic rocks contain alkali feldspar apparently devoid of exsolution lamellae, the data show that all rocks in the Kwandonkaya complex, without

exception, have been equilibrated to various extents in the subsolidus region. Furthermore, there is no end-member microcline. That is, in spite of evidence that exsolution has invariably occurred, the Al-Si ordering has not reached completion.

3) Small-volume intrusions, such as the Buri ring dyke and cone sheets (bearing fayalite, hedenbergite, and hastingsitic hornblende), which underwent high undercoolings, are characterized by orthoclase without microcline. On the other hand, subvolcanic rocks in the Jubga-Ziem ring dyke (bearing amphibole and biotite) contain orthoclase + microcline in their phenocryst-matrix pairs, possibly because of their relatively larger size, higher degrees of volatile build-up and interaction with fluids exsolved from biotite granite. These rock types, along with hypersolvus and drusy biotite granite, preserve intermediate stages of the monoclinic-to-triclinic inversion accompanying the Al-Si order, owing to slower cooling rates and intermediate degrees of rock-fluid interaction.

4) The larger intrusive masses of biotite granite contain mainly microcline and display evidence of volatile build-up and various extents of fluid-rock interaction, except in some localized occurrences of hypersolvus and drusy variants. Overall, the textural variants of biotite granite display consistent similarity in their t_0 , N_{O_2} , and Δ , both at the margins and central part of the pluton. This suggests that the composite biotite granite experienced a rapid (uniform) rate of cooling and extent of postmagmatic equilibration, except in localized areas that experienced volatile loss to yield drusy biotite granite.

5) The three individual centers are mostly displaced laterally (southeastward) from each other as shown in Figs. 5.1, 5.7). There is no evidence in any of the earlier centers (Buri and Jubga-Ziem) being heated up to result in the resetting of the Al-Si order of earlier rock units. However, the effect of a fluid phase (exsolved from the plutonic biotite granite) on the relics of the Jubga-Ziem ring dyke cannot be precluded.

6) The sodic feldspar from the Kwandonkaya complex departs from low albite and is inferred to have contained a ternary alkali feldspar (or calcic sanidine), traces of which are now gone. This implies that the observed disorder is "residual", and reflects the former presence of Al in a more Ca- and Al-rich primary plagioclase that is now more sodic than it once was.

7) The K-feldspar in the subvolcanic and plutonic rocks of the Kwandonkaya complex were mostly equilibrated at compositions defined by the strain-free solvus between 200 and 450°C.

Table 5.2: Parameters calculated from unit-cell (XRD) measurements. The unit-cell volume (V in Å) and number of lines (#) used in determination of unit cell measurements are also given. In all samples *Mc* = microcline, *Or* = orthoclase, *Ph* = phenocryst, and *Mtx* = matrix. Other abbreviation in order of their first usage include: *Fsg* = fine-grained subsolvus granite, *Hsg* = hematite-speckled part of granite, *Wpr* = white perthite, *Lgd* = leucocratic granite dyke in hypersolvus biotite granite, *Chg* = coarse-grained hypersolvus granite, *Ocr1* & 2 = orange core of phenocryst, *Omg* = orange feldspar in miarolitic cavity of coarse-grained biotite granite, *Pcr* = pink core of phenocryst, *Rph* = red phenocryst, *Tph* = translucent phenocryst, *Pph* = pink phenocryst, *OgS* = overgrowth and substrate, *Gcr* = green core, *Wrm* = white rim, *Yrp* = yellow rim of phenocryst, *Wph* = white phenocryst, *Rpr* = red perthite, *Ccr* = cream core, *Incl* = inclusion of phenocryst, and *Trand.* = trachyandesite enclave matrix. The raw data are presented in Appendix V.1.

K-feldspar											Albite										
	Δb_c	$\Delta\alpha^*\gamma^*$	t1o	t1m	t2o+t2m	N _{Or}	Δa	$\Delta(Or)$	V(A ³)	#		Δb_c	$\Delta\alpha^*\gamma^*$	t1o	t1m	t2o+t2m	N _{Or}	Δa	$\psi(Ab)$	V(A ³)	#
Hematite speckled biotite granite																					
22(Fsg)-Mc	0.95	0.92	0.94	0.01	0.05	1.00	-	0.92	722.37	46	0.99	0.97	0.98	0.01	0.01	0.01	0.00	1.13	664.57	48	
22(Hsg)-Or	0.92	-	0.46	0.46	0.08	0.92	0.00	-	720.04	22	1.00	0.98	0.99	0.01	0.00	0.00	0.00	1.11	664.33	58	
Drusy biotite granite																					
30A-Mc	0.96	0.91	0.93	0.02	0.04	0.99	-	0.90	722.01	26	1.00	0.99	1.00	0.01	0.00	0.01	-	-	664.59	52	
37(Wpr)-Or	0.89	-	0.45	0.45	0.11	0.96	-	-	721.56	26	0.99	0.98	0.99	0.01	0.01	0.01	0.00	1.11	664.54	46	
37-Mc	0.95	0.94	0.95	0.01	0.05	0.98	-	0.90	721.97	37	0.99	0.97	0.98	0.01	0.01	0.01	0.00	1.03	664.20	44	
Hypersolvus biotite granite																					
83(Lgd)-Or	0.89	-	0.44	0.44	0.11	0.95	-	-	720.92	25											
83(Lgd)-Mc	0.99	0.65	0.82	0.17	0.01	1.00	-	0.81	722.65	17	0.99	0.98	0.99	0.01	0.01	0.01	0.00	1.12	664.54	48	
83(Chg)-Mc	0.93	0.78	0.86	0.08	0.07	0.98	0.07	0.75	721.82	22	0.99	0.98	0.99	0.01	0.01	0.00	0.00	1.13	663.99	45	
Coarse-grained biotite granite																					
26B-Mc	0.95	0.79	0.87	0.08	0.05	1.00	-	0.77	722.52	50	1.00	0.99	0.99	0.01	0.00	0.01	0.00	1.10	664.56	50	
Porphyritic biotite granite																					
1B(Ocr1)-Mc	0.95	0.68	0.81	0.14	0.05	0.98	-	0.65	721.89	37	1.00	0.96	0.98	0.02	0.00	0.02	0.00	1.13	664.72	60	
1B(Ocr2)-Mc	0.97	0.83	0.90	0.07	0.03	0.99	-	0.82	722.14	41	0.98	0.98	0.98	0.00	0.02	0.00	0.00	1.12	664.06	58	
1B(Mtx)-Mc	0.93	0.89	0.91	0.02	0.07	1.00	-	0.90	722.51	36	0.99	0.98	0.99	0.01	0.01	0.00	0.01	1.13	663.82	47	
25D(Cmg)-Mc	0.96	0.93	0.96	0.02	0.04	0.98	0.01	0.93	721.91	52	1.00	0.97	0.99	0.02	0.00	0.01	0.00	1.12	664.14	47	
35A(Pcr)-Mc	0.97	0.78	0.88	0.10	0.03	0.95	0.02	0.76	720.68	31	1.02	0.98	1.00	0.02	0.02	0.00	0.00	1.10	663.97	50	
35A(RPh)-Mc	0.95	0.71	0.83	0.12	0.06	0.98	-	0.72	721.65	24	0.99	0.96	0.97	0.01	0.02	0.01	0.00	1.13	664.65	40	
35A(Mtx)-Mc	0.92	0.74	0.83	0.09	0.08	0.96	0.01	0.75	721.18	32	1.00	0.97	0.98	0.02	0.01	0.02	0.00	1.13	664.69	43	
36A(Tph)-Mc	0.94	0.77	0.85	0.09	0.06	0.98	-	0.71	721.82	31	1.00	0.96	0.98	0.02	0.00	0.02	0.01	1.15	665.11	52	
36A(Ppr)-Mc	0.96	0.93	0.95	0.02	0.04	0.99	-	0.90	722.33	56	1.00	0.98	0.99	0.01	0.00	0.01	0.01	1.11	664.68	56	
36A(Ogs)-Mc	0.99	0.98	0.98	0.00	0.02	0.98	-	0.89	721.82	60	1.00	0.97	0.98	0.02	0.00	0.00	0.01	1.14	663.96	47	
Biotite granite porphyry																					
38(Ph)-Or	0.86	-	0.43	0.43	0.14	0.97	-	-	721.87	30	1.01	0.95	0.98	0.03	0.01	0.01	0.00	1.13	664.16	29	
38(Mtx)-Mc	0.95	0.74	0.84	0.10	0.05	0.97	0.03	0.77	721.32	23	0.98	0.97	0.97	0.01	0.02	0.01	0.01	1.15	664.25	38	
82(Ph)-Or	0.82	-	0.41	0.41	0.19	0.96	0.03	-	721.62	30	0.96	0.93	0.95	0.01	0.04	0.01	0.01	1.18	664.55	39	
82(Mtx)-Mc	0.92	0.73	0.83	0.10	0.08	0.99	-	0.65	722.02	20	0.98	0.96	0.97	0.01	0.03	0.00	0.01	1.14	663.99	47	
Hornblende biotite granite porphyry																					
56(Ph)-Or	0.87	-	0.43	0.43	0.13	0.97	-	-	721.70	31	0.98	0.95	0.96	0.02	0.02	0.02	0.00	1.14	665.00	29	
56(Mtx)-Mc	0.90	0.74	0.82	0.08	0.10	0.94	0.01	0.69	720.42	23	0.97	0.96	0.96	0.01	0.03	0.00	0.00	1.13	663.98	41	
68B(Ph)-Or	0.81	-	0.40	0.40	0.19	0.97	0.02	-	721.66	27	0.98	0.90	0.94	0.04	0.02	0.02	0.00	1.17	664.69	39	
68B(Mtx)-Mc	0.87	-	0.43	0.43	0.13	0.93	-	-	720.45	22	0.97	0.92	0.95	0.02	0.03	0.01	0.00	1.17	664.13	34	
Hypersolvus fayalite syenogranite																					
58-Or	0.88	-	0.44	0.44	0.13	0.95	-	-	721.12	26	0.99	0.93	0.96	0.03	0.01	0.01	-	1.15	664.65	47	
58-Mc	0.96	0.69	0.83	0.13	0.04	0.96	-	0.65	721.06	18	1.00	0.96	0.98	0.02	0.00	0.02	0.00	1.15	664.80	38	
Felsite porphyry																					
40D(Mtx)-Or	0.83	-	0.41	0.41	0.17	0.98	0.06	-	722.28	22	0.98	0.95	0.96	0.02	0.02	0.01	0.00	1.15	664.29	36	
61A(Ph)-Or	0.87	-	0.43	0.43	0.13	0.98	-	-	722.07	35	0.95	0.94	0.94	0.01	0.06	0.02	0.01	1.16	665.02	34	

Table 5.2 (cont'd)

K-feldspar											Albite									
	Δbc	$\Delta\alpha^*\gamma^*$	t_{10}	t_{1m}	$t_{20}+t_{2m}$	N_{Or}	Δa	$\Delta(Or)$	$V(\text{\AA}^3)$	#	Δbc	$\Delta\alpha^*\gamma^*$	t_{10}	t_{1m}	$t_{20}+t_{2m}$	N_{Or}	Δa	$\psi(\text{\AA}^\circ)$	$V(\text{\AA}^3)$	#
Felsite porphyry (cont'd)																				
61A(Mtx)-Or	0.92	-	0.46	0.46	0.08	0.95	-	-	720.92	17	0.94	0.92	0.93	0.01	0.06	0.01	0.01	1.17	664.23	34
61A(Mtx)-Or	0.83	-	0.42	0.42	0.17	1.00	-	-	722.83	19	0.96	0.92	0.93	0.02	0.05	0.01	0.00	1.20	664.56	29
Hastingsite granite porphyry																				
40A(Gor)-Or	0.80	-	0.40	0.40	0.21	0.91	0.09	-	719.75	25	0.92	0.82	0.87	0.05	0.08	0.02	0.01	1.23	664.86	27
40A(Wrm)-Or	0.88	-	0.44	0.44	0.12	0.98	-	-	722.13	17	0.98	0.96	0.97	0.01	0.02	0.01	0.00	1.18	664.17	39
40A(Mtx)-Or	0.83	-	0.42	0.42	0.17	0.97	0.01	-	721.68	30	0.97	0.92	0.94	0.02	0.04	0.01	0.00	1.15	664.33	37
Hedenbergite hastingsite granite porphyry																				
64(Yrp)-Or	0.80	-	0.40	0.40	0.21	0.96	-	-	721.53	34	0.98	0.90	0.94	0.04	0.02	0.01	0.01	1.19	664.46	42
64(Wph)-Or	0.79	-	0.39	0.39	0.21	0.97	0.06	-	721.64	35	0.99	0.95	0.97	0.02	0.02	0.06	-	1.15	664.72	26
64(Mtx)-Or	0.81	-	0.41	0.41	0.19	0.99	0.01	-	722.40	27	0.97	0.90	0.93	0.04	0.03	0.00	0.01	1.17	663.98	25
Reddened granite porphyry																				
66(Rpr)-Mc	0.96	0.78	0.87	0.09	0.04	0.95	0.00	0.81	720.92	40	0.98	0.94	0.96	0.02	0.02	0.01	0.01	1.16	663.53	38
66(Cor)-Mc	0.92	0.79	0.86	0.06	0.08	0.96	-	0.78	721.11	30	0.97	0.92	0.95	0.03	0.03	0.00	-	1.19	663.88	32
Hedenbergite fayalite granite porphyry																				
41(Ph)-Or	0.80	-	0.40	0.40	0.20	0.95	0.07	-	721.13	22	0.98	1.00	0.99	0.01	0.02	0.01	0.00	1.08	664.14	28
41(Mtx)-Or	0.87	-	0.44	0.44	0.13	1.00	0.06	-	722.85	15	0.99	0.99	0.99	0.00	0.01	0.09	0.00	1.12	664.34	30
Rhyolite porphyry																				
49A(Ph)-Or	0.85	-	0.43	0.43	0.15	0.99	0.05	-	722.50	25	0.97	0.90	0.93	0.04	0.03	0.02	0.00	1.14	664.71	33
49A(Mtx)-Or	0.87	-	0.44	0.44	0.13	0.98	0.00	-	722.25	21	1.00	0.91	0.95	0.05	0.02	0.01	0.02	1.19	663.52	26
Enclave in rhyolite porphyry																				
49B(Incl)	-	-	-	-	-	-	-	-	-	-	0.97	0.92	0.95	0.03	0.03	0.01	0.00	1.20	664.62	22
49B(Trand.)	0.85	-	0.43	0.43	0.15	0.97	0.07	-	721.85	20	0.93	0.84	0.89	0.05	0.07	0.01	0.01	1.23	664.63	26

Definition of terms and notations in Table above and text

The terms used in this Table and the text are the same as defined by Ribbe (1983), and Kroll & Ribbe (1983). This includes: 1) Δbc or $(t_{10} + t_{1m}) = (b - 0.7138 - 1.7505c)/(-7.7245 + 1.0150c)$; 2) $\Delta\alpha^*\gamma^*$ or $(t_{10} - t_{1m}) = (\alpha^* + 89.118 - 1.9902\gamma^*)/(-24.691 + 0.2229\gamma^*)$; 3) $t_{10} = [\Delta(bc) + \Delta(\alpha^*\gamma^*)]/2$ from equations 1 and 2. Substituting the t_{10} term into either the expression for Δbc or $\Delta\alpha^*\gamma^*$ the t_{1m} term can be solved; and by difference, 4) $(t_{20} + t_{2m}) = 1 - (t_{10} + t_{1m})$; 5) $\Delta = 12.5(d_{131} - d_{131})$, where Δ is the triclinicity of K-feldspar in $\text{\AA}$; and 6) $\psi = 2\theta_{131} - 2\theta_{131}$, where ψ is the obliquity of Na-feldspar (plagioclase) in degrees. 6) The mole fraction of orthoclase in intermediate alkali feldspars (N_{Or}) = $-872.330 + 3.82791V - (5.61695 \times 10^{-3})V^2 + (2.75683 \times 10^{-6})V^3$. 7) The mole fraction of K-feldspar in microcline and low albite (N_{Or}) = $-1209.141 + 5.28104V - (7.70522 \times 10^{-3})V^2 + (3.75650 \times 10^{-6})V^3$. Equations 6 & 7 are after Kroll & Ribbe (1983). 8) $\Delta a = a_{obs} - a_{est}$, where $a_{obs} = a$ values from XRD measurements, and $a_{est} = 8.634 - (11.437 - 0.12226bc)^{1/2}$ (Kroll & Ribbe, 1983).

Chapter 6: Petrographic and SEM study of chevkinite, allanite, and other accessory minerals and their significance in the evolution of the complex

6.1 Introduction

Accessory minerals can exert a significant control on the composition of silicate melts. For example, they may or may not be residual during partial melting, and may well be involved in fractional crystallization (Exley 1980, Brooks et al. 1981, Watson & Harrison 1984, Michael 1988, Czamanske 1988, Bacon 1989). Experimental studies have also been conducted to relate the concentration of essential structural constituents (ESC) in felsic melts to the saturation of accessory minerals (Zr: Watson 1979, P: Harrison & Watson 1984, Ti, Zr, Hf: Ellison & Hess 1986, REE: Schnetzler & Philpotts 1970, Green & Pearson 1986, 1988, Michael 1988, Sawka et al. 1984). Also, the hydrothermal mobility of REE, Zr, Th, Nb, Ta, and Ti has been demonstrated in some granitic suites through the study of accessory minerals, for example, in the intrusive units and country rocks of the Skye Tertiary igneous complex, Scotland (Exley 1980). In the French Massif Central, Cathelineau (1987) demonstrated a high degree of Zr mobility in a hydrothermal environment, contrary to its characterization in most geological environments as an immobile element (Rubin et al. 1989).

In addition to the distinction of the NYG complexes from Pan-African granites in the basement on the basis of their ferromagnesian minerals, the suite of accessory minerals also is unusual. In the NYG province, some suites of accessory minerals (e.g., columbite, cassiterite, xenotime, and other REE-phases) have been related to episodes of alkali and acid metasomatism (Kinnaird 1976, 1987, Kinnaird et al. 1985, Ixer et al. 1987). Ixer et al. (1987) first investigated the accessory mineralogy of the Ririwai complex; further studies provided evidence for metastable solid-solution of thorite and coffinite and indications of zircon-thorite_{ss} (Pointer et al. 1988a), and mobility of Zr, Th, and U during albitization and greisenization (Pointer et al. 1988b).

This study employs petrographic features and SEM-BSE images of accessory minerals to indicate processes that affected the rocks at magmatic and subsolidus stages. The role of the accessory minerals in the magmatic evolution is qualitatively appraised by relating observed trace-element contents of their host rock(s) to the relative sequence of crystallization of such minerals. Furthermore, petrographic evidence is sought for the remobilization of the essential structural constituents (ESC) of accessory minerals at subsolidus, hydrothermal stages. From these studies, limited mobility of the REE is indicated in subvolcanic rocks and pristine biotite granites, and more important mobility of REE, Zr, Th, Nb, Ta, and Ti is demonstrated in the albitized and greisenized biotite granites. The quantitative appraisal of mass gains and losses associated with subsolidus to hydrothermal effects was not undertaken because of the following problems: 1) there is a lack of evidence for substantial changes in whole-rock chemical compositions due to alteration; in fact, mobility can in most cases only be indicated on the scale of individual minerals, 2) minor gains and losses of elements cannot be effectively accounted for because of their low magnitude in this instance relative to analytical errors.

6.2 Petrography and SEM studies of accessory minerals

In this section, chevkinite and allanite are characterized first, followed by Fe-Ti oxides, zircon, thorite, and other accessory minerals. Table 6.1 provides a list of rock types and the suite of accessory minerals in such rock units. Most of the accessory minerals are distinguished mainly by scanning electron microscopy (SEM) (Appendix VI.1, 2); results of both petrographic and SEM studies are presented together.

6.2.1 Chevkinite

Chevkinite has the general formula $[A_4BC_2D_2O_8(Z_2O_7)_2]$ (Ito 1967, Ito & Arem 1971,

(McDowell 1979, Haggerty & Mariano 1983, Green & Pearson 1988), where A = REE³⁺, Ca, Na, B = Fe²⁺, Mg, C = Ti⁴⁺, Mn²⁺, Fe³⁺, Al, D = Ti⁴⁺, and Z = Si. A typical EDS spectrum is shown in Appendix VI.1a.

(Within the rhyolite porphyry and hedenbergite granite porphyry (of the Buri ring dyke), chevkinite grains occur as microphenocrysts in the groundmass (Figs. 6.1a, b), or intergrown with and trapped as inclusions in hedenbergite (Figs. 6.1c, d). The most extensive occurrence of chevkinite is within the hastingite granite porphyry, where it forms microphenocrysts in the groundmass (Figs. 6.1e, f, Appendix VI.3a-c), and inclusions in orthoclase microperthite (Fig. 6.2a), hastingitic hornblende (Fig. 6.2b), allanite (Fig. 6.2c), and also in an assemblage of magnetite + quartz (Fig. 6.2d). Its development as microphenocrysts, and as inclusions in orthoclase microperthite, fayalite, and hedenbergite, indicates that chevkinite crystallized as an early magmatic phase, whereas the inclusions in amphibole are considered to be late magmatic. Some of the chevkinite grains show an irregular development of a low-REE mantle and light patches (Figs. 6.1a, b), which are interpreted as being of postmagmatic origin. It is suspected, but not established, that perrierite occurs along with chevkinite (darker, low-REE patches and zones in Figs. 6.1a to 6.1f). Perrierite is related to chevkinite and has a similar structural formula, and typically forms at the expense of chevkinite by decomposition, oxidation, and recrystallization after the expulsion of Ce and oxidation of Fe²⁺ to Fe³⁺; also, it typically is richer in Al and Ca (Bergen 1984). In other grains, the chevkinite displays oscillatory zoning (Figs. 6.1c - f, 6.2a, b, d, Appendix VI.3c) or patchy zoning superimposed on oscillatory zoning (Figs. 6.1c, e, f, 6.2d). The oscillatory zoning is considered a magmatic feature, whereas patchy zoning is attributed to subsolidus mobilization of the REE. The rim is not metamict in chevkinite from the rhyolite porphyry (Figs. 6.1a - d), whereas in some grains in hedenbergite hastingite granite porphyry, the chevkinite invariably is metamict, i.e., it is typically nonbirefringent, with cracks infilled by

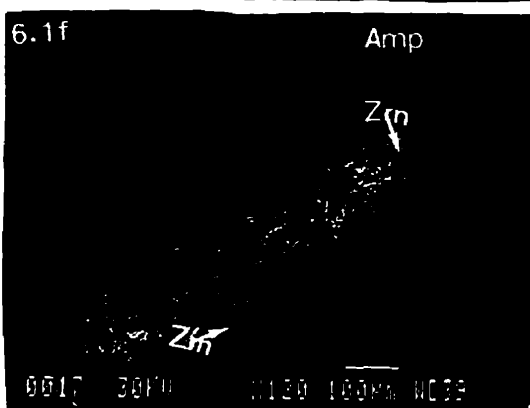
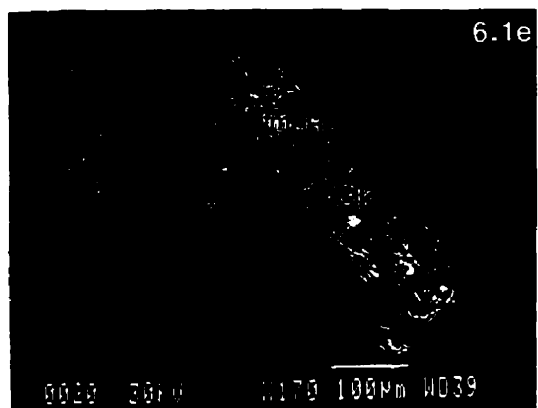
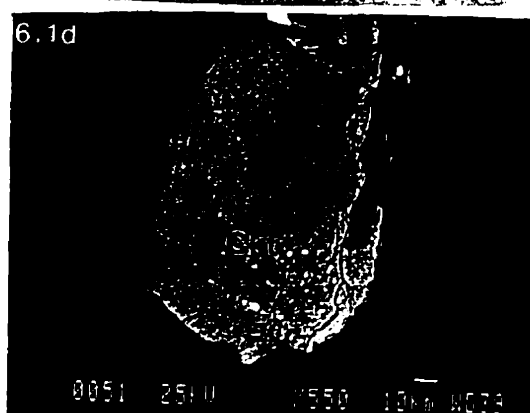
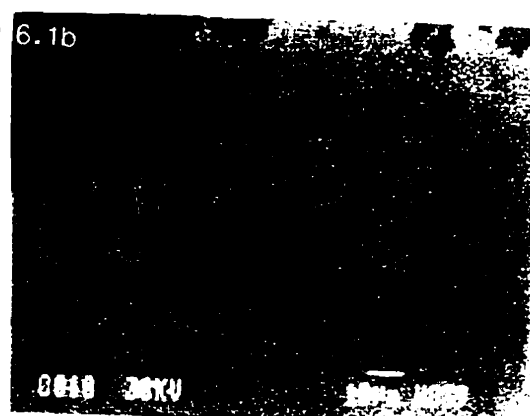
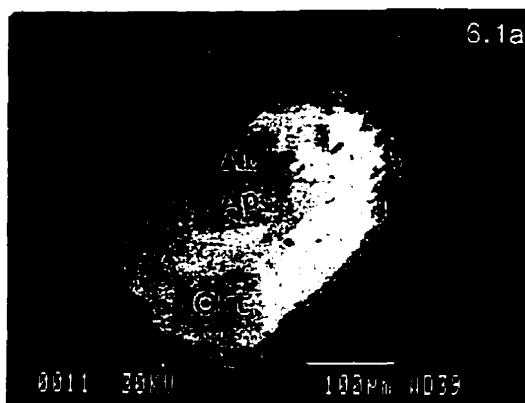
Table 6.1: Accessory mineral suite in the main rock types of the Kwandonkaya complex

Rock type	Suite of accessory minerals
BURI RING DYKE Rhyolite porphyry and hedenbergite fayalite granite porphyry	Ckt, Zrn, Ilm, Mag (Ti-rich), [<i>Ap, Bhl, Pyr</i>]
CONE SHEETS (Jubga-Ziem center) Hastingsite granite porphyry	Ckt, Aln, Zrn, Ilm (Mn- & Nb-rich), Mag, [<i>Col, Ttn, Ap, Fl, Brt</i>]
Other cone sheets	Aln, Ckt, Zrn, Ilm, Mag, [<i>Fl</i>]
JUBGA-ZIEM RING DYKE a) Hypersolvus (fayalite) syenogranite	Ilm, Fst, Zrn, Mnz, Aln, Mag, Fl, Flc, Mnz, Tht, Bst, [<i>Fe-Tgs, Bst, Spl, Pyr, & Cpyr</i>]
b) Hornblende biotite granite porphyry	Aln, Zrn, Fl, Ilm, Mag, Bst, [<i>Fst, Xnt, Flc</i>]
c) Biotite granite porphyry	Mag, Ilm, Zrn, Fl
BIOTITE GRANITE a) Coarse-grained biotite granite	Zrn, Ilm, Fl, Prt, Flc, Mnz, [<i>Bst, Aln</i>]
b) Fine-grained (hypersolvus) biotite granite	Ilm, Mag, Zrn, Fl, Toz, Y-Flc, [<i>Fst, Xnt, Bst</i>]
c) Other fine-grained biotite granite	Ilm, Mag, Fl, Toz, Hem, Zrn, Mnz, [<i>Xnt, Ap, & Flc</i>]
ALTERED PLUTONIC ROCKS a) Albitized biotite granite	Zrn, Tht, Hem, Cst, Col, Toz, Fl, Flc [<i>Mag, Brl, Brt, Bst</i>]
b) Greisens	Zrn, Tht, Nb-Rut, Mnz, Col, Cst, Flc, Aln, Hem, [<i>Hf-Zr-Tht, U-Tht, Nb-Tgs, Nb-Mn-mineral</i>]

Symbols:

Aln: allanite {Ca(LREE,Y)(Fe)(Al)₂(SiO₄)₃(OH)}, *Ap*: apatite {(Ca,REE)₅[(P,Si)O₄]₃(OH,F)}, *Brt*: barite (BaSO₄), *Bst*: bastnäsite {REE(CO₃)F}, *Brl*: beryl {Al₂(Be₃Si₆)O₁₈}, *Bhl*: britholite {(Ce,Y,Ca)₅[(Si,P)O₄]₃(OH,F,Cl)}, *Cst*: cassiterite (SnO₂), *Cpyr*: chalcopyrite (CuFeS₂), *Ckt*: chevkinite (see text), *Col*: columbite {(Nb,Ta,Ti)₂(Fe,Mn)O₆}, *Fst*: fergusonite {(LREE,Y)(Nb,Ta,Ti)O₄}, *Fe-Tgs*: ferritungstite {(Fe,Mn)WO₄}, *Flc*: fluocerite {(REE)F₃}, *Fl*: fluorite (CaF₂), *Hem*: hematite (Fe₂O₃), *Ilm*: ilmenite (FeTiO₃), *Mag*: magnetite (Fe₃O₄), *Mnz*: monazite {(LREE,Y,Ca,Th)PO₄}, *Prt*: parisite {Ca(REE)₂(CO₃)₃F₂}, *Pyr*: pyrite (FeS), *Nb-Rut*: Nb-bearing rutile {(Ti,Nb)O₂}, *Spl*: sphalerite (ZnS), *Tht*: thorite {(Th,Zr,Hf,U)SiO₄}, *Toz*: topaz {Al₂SiO₄(OH,F)₂}, *Ttn*: titanite (CaTiSiO₅), *U-Tht*: uranothorite {(Th,U)SiO₄}, *Xnt*: xenotime (YPO₄), *Nb-Tgs*: niobotungstite (Nb-MnWO₄), *Zrn*: zircon {(Zr,Th,Hf)SiO₄}. The less commonly occurring accessory minerals are listed in italics within square brackets.

Fig. 6.1: SEM-BSE (scanning-electron-microscope - back-scattered-electron) images of chevkinite in various mineral associations within subvolcanic rocks: **a)** Microphenocryst of chevkinite (Ckt) in the groundmass of rhyolite porphyry. The chevkinite is characterized by brighter (higher $\Sigma\text{REE}_2\text{O}_3$, ΣFeO , and low Nb_2O_5 , and TiO_2), and irregularly distributed darker parts (slightly lower $\Sigma\text{REE}_2\text{O}_3$, and ΣFeO , and higher Nb_2O_5 , and TiO_2). The brighter part contains inclusions of REE-bearing silicon-bearing apatite (Ap) and albite (Ab). The irregular darker grey, Nb-rich and low- ΣREE phase may be perrierite (see text), and the sharp but irregular boundary of the mantle suggests an origin by postmagmatic replacement. **b)** An enlargement of parts of chevkinite grain in Figure 6.1a. This photo illustrates the sharp boundary between the predominant brighter (REE-rich) and darker (REE-poor) patches. This texture suggests a postmagmatic replacement of the phenocrysts. The presence of trapped albite along with apatite is anomalous here, because both the groundmass and phenocrysts contain orthoclase microperthite with no distinct grains of albite. **c)** Chevkinite (Ckt) included in replaced hedenbergite (annite + magnetite) in rhyolite porphyry. The upper grain displays patchy low- ΣREE (dark grey) zones, whereas the lower grain has a regular low- ΣREE zone. **d)** Enlargement of the lower grain of chevkinite in Figure 6.1c, displaying a core zone that is darker (poorer in REE) and an outer rim that shows rhythmic zoning. **e)** Prismatic section of chevkinite (Ckt) along the outer margin of orthoclase microperthite within hastingite granite porphyry. Inner part of grain has patchy brighter (high- ΣREE), and darker grey (ΣREE -poor) zones, and the grain has a continuous (about 10 μm wide) metamict rim (Mtm-R). **f)** Metamict microphenocryst of chevkinite (Ckt) in the devitrified groundmass of hastingite granite porphyry. The grain contains inclusions of allanite (Aln) and zircon (Zrn), with a metamict rim containing Fe-oxides. The patchy zones indicate subsolidus redistribution of REEs. Amp = amphibole. (*Figs. 6.1a - d: rhyolite porphyry sample 49, Buri ring dyke; Figs. 6.1e, f: hastingite granite porphyry sample 40A in cone sheet). Samples located on Fig. 2.1).*



Fe-oxides, and has a low-intensity (darker) BSE image (Figs. 6.1e, f). It is notable that, within the hastingsite granite porphyry, there is no such rim on chevkinite grains included in orthoclase microperthite (Fig. 6.2a) and fayalite (Fig. 6.2d), but a metamict rim is developed on chevkinite microphenocrysts in the matrix (Figs. 6.1e, f), and on inclusions in amphibole (Figs. 6.2b) as well as in allanite (Fig. 6.2c). Also, inclusions of zircon in chevkinite (Fig. 6.1f) and amphibole (Figs. 6.2b, c) in the same thin section do not have a metamict rim. This observation suggests that chevkinite is more susceptible to radiation damage than zircon.

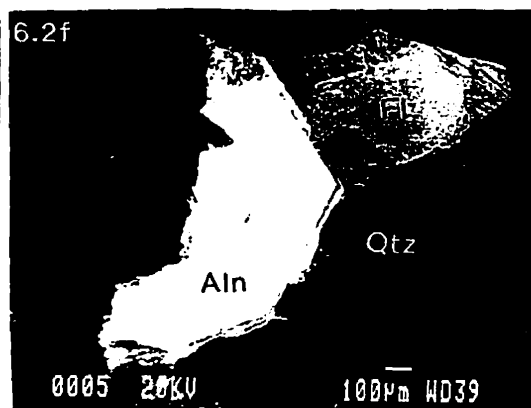
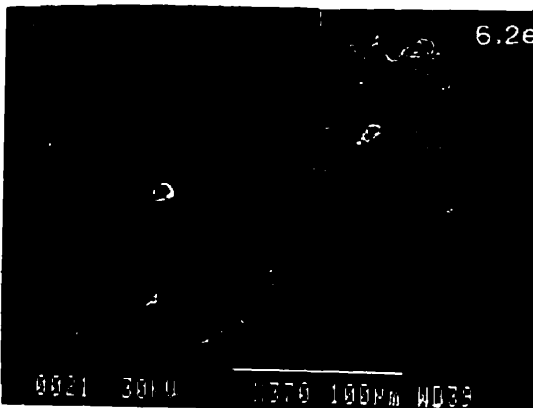
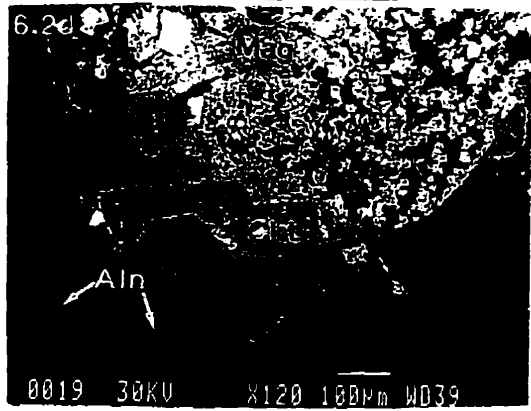
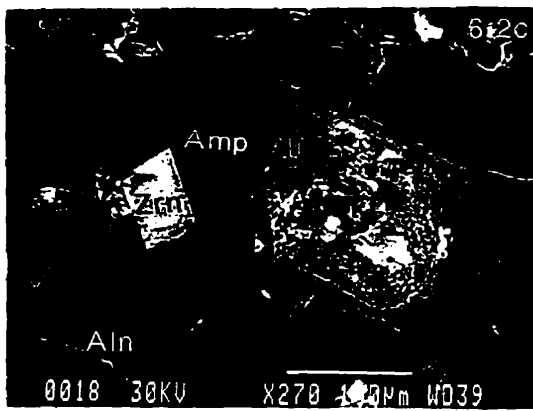
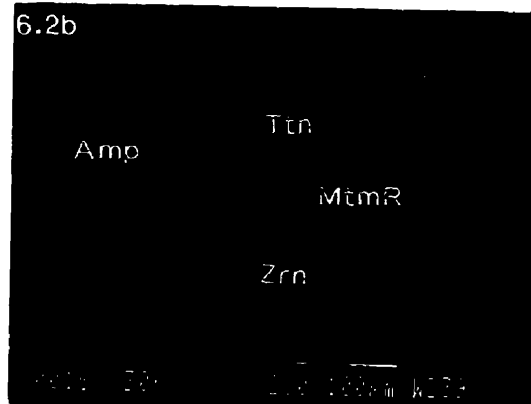
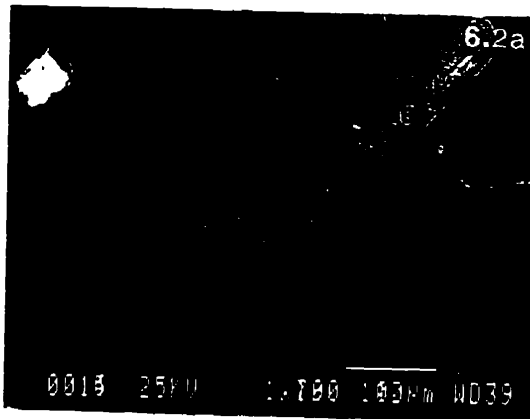
Semiquantitative SEM-EDS analyses show chevkinite in the rhyolite porphyry (Buri ring dyke) to be characterized by relatively dark, low-REE areas ($38.1 \leq \Sigma\text{REE}_2\text{O}_3 \leq 39.5\%$) and brighter high-REE areas, in cases in oscillatory zones ($44.1 \leq \Sigma\text{REE}_2\text{O}_3 < 46.8\%$). In the hastingsite granite porphyry (cone sheets), both the darker zones ($20.8 \leq \Sigma\text{REE}_2\text{O}_3 \leq 29.5\%$) and brighter zones ($31.4 \leq \Sigma\text{REE}_2\text{O}_3 \leq 37.3\%$) have lower ΣREE values than the chevkinite in rhyolite porphyry. The difference in chevkinite between both rock types is discussed below.

The mineral chevkinite is related to titanite by the generalized substitution scheme: $\text{REE}^{3+}_A\text{M}_C(\text{Ca}^{2+}_A\text{Ti}^{4+}_C)_{-1}$, where $\text{M}_C = \text{Fe}^{2+}, \text{Mg}, \text{Fe}^{3+}, \text{Zr}, \text{Nb}, \text{P}, \text{and K}$, of aggregate valence 3+. One of the compositions of chevkinite is anomalous, with 31.4 wt % TiO_2 and 20.8 % $\Sigma\text{REE}_2\text{O}_3$. This has over 5% excess TiO_2 than the range normally encountered ($13.2 \leq \text{TiO}_2 \leq 24.8\%$). Furthermore, the titanite intergrown with this sample of chevkinite has only 1% $\Sigma\text{REE}_2\text{O}_3$. The REE content of titanite rarely exceeds 5% (Speer & Gibbs 1976). The REE-rich "titanite" may indicate a partial preservation of a hitherto undocumented solid-solution between titanite and chevkinite, or may represent compositions of chevkinite with submicroscopic inclusions of titanite.

6.2.2 Allanite

Allanite has the chemical formula $\{\text{Ca}(\text{LREE}, \text{Y})(\text{Fe})(\text{Al})_2(\text{SiO}_4)_3(\text{OH})\}$, and occurs as

Fig. 6.2: SEM - BSE images of chevkinite (a - d), and allanite (b, e - f) in subvolcanic rocks: **a)** Drumstick-shaped chevkinite (Ckt) inclusion within orthoclase microperthite in hastingite granite porphyry. The grain of chevkinite is relatively homogeneous, unzoned, and has no metamict (opaque) rim. A drumstick-shaped microinclusion of apatite (Ap) occur within the chevkinite. **b)** Chevkinite (Ckt) hosted by hastingitic hornblende (in hastingite granite porphyry) with relatively homogeneous inner part, and a rhythmically zoned as well as patchy and darker (low ΣREE) outer part. The metamict rim of chevkinite bears inclusions of zircon and titanite. **c)** Inclusion of zoned allanite (Aln) within amphibole in hastingite granite porphyry. The allanite is patchy in the inner part and has relict inclusions of replaced chevkinite (Ckt); zircon (Zrn) also occurs in the amphibole. Some cleavages in the amphibole are filled by secondary allanite that has higher levels of ΣREE . It seems that the core of the grain was initially chevkinite that was later overgrown and replaced by allanite producing the complex intergrowth within amphibole. **d)** Replaced assemblage of magnetite (Mag) and quartz (Qtz) with inclusions of ilmenite (Ilm) and rhythmically as well as patchily zoned chevkinite (Ckt). Fractures close to this intergrowth are filled with secondary allanite (Aln). **e)** Microphenocryst of allanite (Aln) in the groundmass of hastingite granite porphyry. Inner parts of allanite have patchy zoning, whereas the pyramidal termination is rhythmically zoned. **f)** Zoned allanite (Aln) intergrown with dark grey fluorite (Fl) that is partly rimmed by annitic biotite (in hornblende biotite granite porphyry). The zoned grain of allanite is partly replaced along its margin by mica and bright specks of bastnäsite. The zone of replacement is overgrown by low-REE allanite that contains brighter grains of Th-bearing bastnäsite. The thin bright rim around the fluorite is fluocerite ($\pm$ xenotime). (*Figs. 6.2a-e: hastingite granite porphyry sample 40A in cone sheet; Fig. 6.2f: hornblende biotite granite porphyry sample 68B, Jubga-Ziem ring dyke. Samples are located in Fig. 2.1).*



ehedral, rhythmically zoned grains in metaluminous to mildly peraluminous subvolcanic rocks in the cone sheets and Jubga-Ziem ring dyke. Allanite is conspicuously absent in the essentially metaluminous Buri center, and is rare in the pristine biotite granite. In most sections containing both allanite and chevkinite, the allanite is later (Figs. 6.1f), and in a few instances, both minerals seems to have cocrystallized (Fig. 6.2c). The allanite occurs as inclusions in chevkinite (Fig. 6.1f), and amphibole (Fig. 6.2c), as microphenocrysts (Fig. 6.2e, Appendix VI.3d, e), or as interstitial grains (Fig. 6.2f). The most common occurrence of allanite is as interstitial grains in hornblende biotite granite porphyry. The interstitial allanite is interpreted as being late magmatic, whereas the allanite occurring in veinlets (Figs. 6.2c, d, f) clearly is subsolidus (hydrothermal). The late-magmatic allanite is invariably characterized by oscillatory zoning in all rock units.

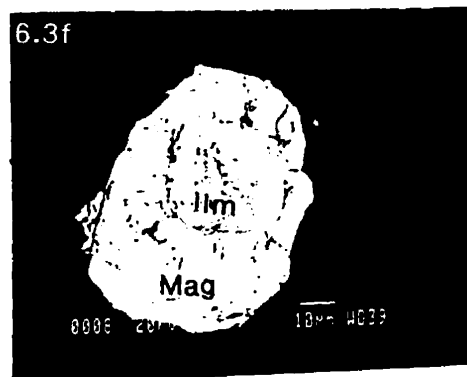
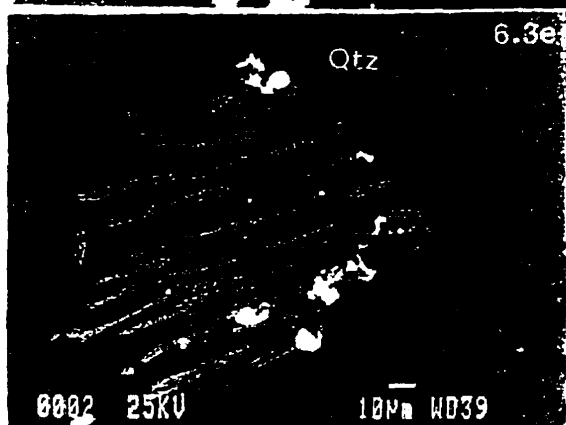
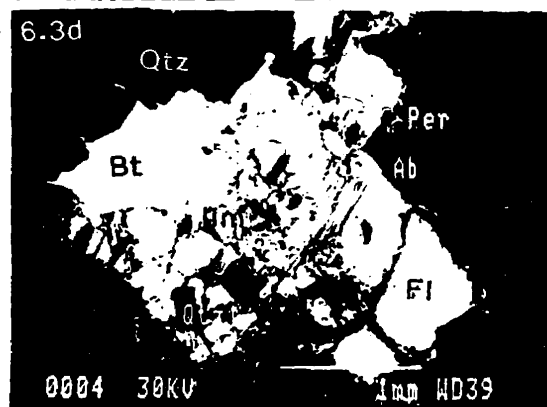
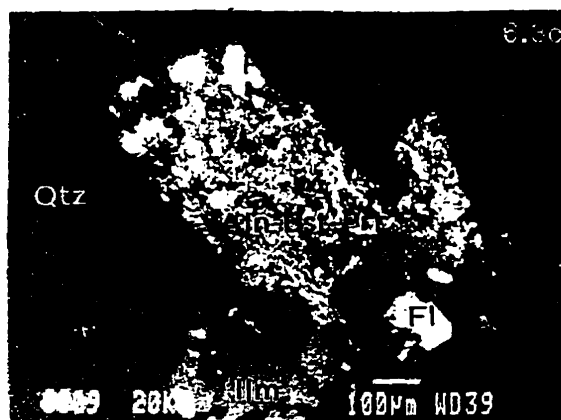
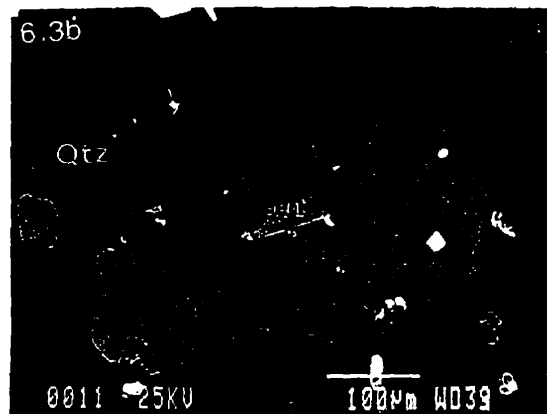
In biotite granite, allanite is a minor accessory phase and mainly secondary, usually occurring in replacement assemblages such as illustrated in Figure 6.3c. In greisen zones, allanite occurs in veinlets intergrown with hydrothermal biotite.

Semiquantitative SEM-EDS analyses show the allanite to consistently have lower ΣREE , Ti, and Nb, and higher Si, Fe, and Al than chevkinite (Appendix VI.2b). The darker zones are poorer in REE ($16.3 \leq \Sigma\text{REE}_2\text{O}_3 \leq 17.6\%$), whereas the brighter zones are richer ($19 \leq \Sigma\text{REE}_2\text{O}_3 \leq 25.6\%$). The lack of standardized (quantitative) analyses makes these statements tentative.

6.2.3 Fe-Ti oxides

Ilmenite (FeTiO_3) is widespread in the subvolcanic rocks and more pristine samples of biotite granite; it displays a variety of habits. In the subvolcanic rocks, the ilmenite occurs as inclusions in altered fayalite (Fig. 6.2d), as platy and skeletal grains (Fig. 6.3a), and as interstitial microphenocrysts (Figs. 6.3d, f). The ilmenite in subvolcanic rocks spans the early to late magmatic range of crystallization. In most other rock types, the ilmenite is included in biotite

Fig. 6.3: SEM - BSE images of ilmenite displaying different textural features and associated with a variety of accessory minerals: **a)** Anhedral grains of ilmenite (Ilm) in hypersolvus syenogranite with marginal fergusonite (Fg) and fluocerite (Fc). The ilmenite grain is intergrown with secondary annitic mica (Ann), monazite (Mnz), and magnetite (Mag). This association suggests crystallization of syenogranites at an $f(\text{O}_2)$ less than the quartz-ulvöspinel-ilmenite-fayalite (i.e., $\leq \text{QUIF}$) and subsolidus alteration at $f(\text{O}_2)$ greater than quartz-fayalite-magnetite (i.e., $\geq \text{QFM}$) buffers. **b)** Subhedral grain of ilmenite (Ilm) with inclusion of zircon (Zrn), bastnäsite (Bst) and biotite (Bt), which also rims the ilmenite (26A; coarse-grained biotite granite). Left of the ilmenite grain, normal granular quartz (Qtz) occurs, whereas a "granophyric" intergrowth of quartz and biotite can be seen in the upper right corner. **c)** Biotite (Bt) bearing inclusions of ilmenite (Ilm), fluorite (Fl), quartz, and fluocerite (Fc) in coarse-grained biotite granite. Fluocerite is replaced partly by parisite (Prt), bastnäsite (Bst), and allanite. Intergrown quartz and biotite occur in the upper right corner of the field of view. **d)** An assemblage of ilmenite (Ilm), biotite (Bt), quartz (Qtz), and zoned fluorite (Fl) in "hornblende"-biotite granite porphyry. The association may be produced by the late-magmatic replacement of calcic amphibole in a drusy cavity. Perthite (Per) grains have dark grey domains of albite (Ab). **e)** Multigranular biotite (Bt) with an interstitial intergrowth of lamellar ilmenite (Ilm) and fluorite (Fl) seen only in coarse-grained biotite granite. The ilmenite grew along $\{111\}$ cleavage planes in fluorite, presumably rapidly. Monazite occurs close to or along the ilmenite-fluorite intergrowth. **f)** An intergrowth characterized as "sandwich-type" (cf. Haggerty 1991). The ilmenite (Ilm) in the core contains inclusions of zircon, and is surrounded by brighter magnetite (Mag) containing trellis-type, exsolved lamellae of ilmenite (in "hornblende"-biotite granite porphyry). The latter was an initially homogeneous rim of Usp-Mag_{ss} that exsolved (by oxidation-exsolution) into magnetite + ilmenite (see text). (*Fig. 6.3a: sample 58; Figs. 6.3b, c, e: sample 26; Figs. 6.3d, f: sample 68B. All samples are shown in Fig. 2.1).*



(Figs. 6.3b, c, e). Some of the ilmenite grains in these examples themselves have inclusions of zircon, biotite, and monazite, which are interpreted as being late-magmatic and as having crystallized by boundary-layer saturation (Fig. 6.3b). The finer-grained and anhedral ilmenite in mica is subsolidus, formed by oxidation-induced breakdown of the magmatic biotite.

A rarely observed feature, restricted to coarse-grained biotite granite, is a trellis-type intergrowth along {111} cleavage planes of magmatic fluorite (Fig. 6.3e). The lamellar intergrowth presumably formed as a result of local supersaturation and coprecipitation of the two phases. The more normal ilmenite-magnetite intergrowth is well displayed in the sandwich-type intergrowth (Fig. 6.2f) in "hornblende"-biotite granite porphyry, and trellis-type intergrowth (Fig. 6.3a) in hypersolvus biotite granite. The sandwich-type intergrowth contains an ilmenite core (with micro-inclusions of zircon) mantled by titaniferous magnetite that contain exsolved trellis-type lamellae of ilmenite (Fig. 6.3a). Such exsolved, intergrown magnetite and ilmenite also occur in hypersolvus biotite granite, where the magnetite seems to have undergone subsolidus alteration (Fig. 6.3b).

6.2.4 Zircon and thorite

Zircon $\{(Zr,Th,Hf)SiO_4\}$ is the most widely occurring accessory mineral. It is found in subvolcanic rocks as inclusions in chevkinite (Figs. 6.1e, f), amphibole (Figs. 6.1e, 6.2a), and quartz; as well, it forms microphenocrysts in the groundmass. In most sections of fayalite syenogranite and "hornblende"-biotite granite porphyry, zircon develops a peculiar skeletal (cuneiform) habit. This reflects relatively high growth-rate associated with high undercooling. The zircon occurring as inclusions in hedenbergite, amphibole, and K-feldspar, as well as microphenocrysts in subvolcanic rocks of Buri and Jubga-Ziem centers, is clearly early magmatic.

In biotite granite, zircon occurs mostly as inclusions within biotite (Fig. 6.3e). In this

instance, the zircon is considered late magmatic. Within the albitized biotite granite and greisen, the zircon (bearing micro-inclusions of thorite) is associated with thorite of similar size (Figs. 4.4d, 6.4e). The brighter parts of zircon crystals in albitized biotite granite have relatively less U, Fe, and Al than the darker patches (Fig. 6.4e), and the rim of zircon is partially metamict and infilled by Fe-oxides (the Fe-oxides in zircon in turn host Th-oxides). The implication of these observations is that the greisen did not form from a geochemically exotic magma, but rather at a postmagmatic stage, by interaction of jointed rocks with hydrothermal fluids.

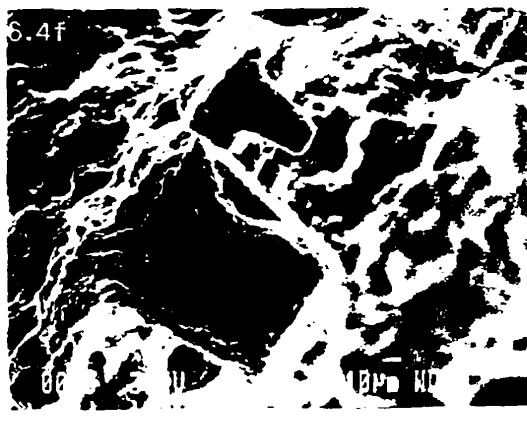
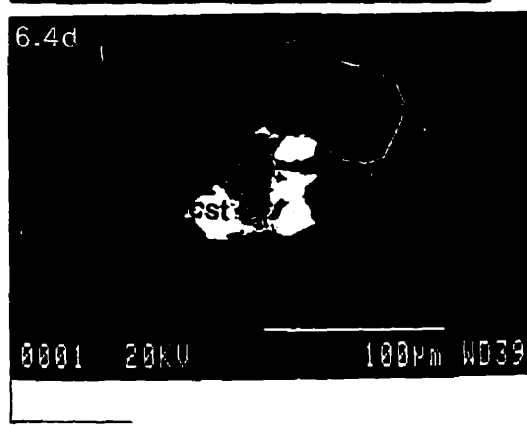
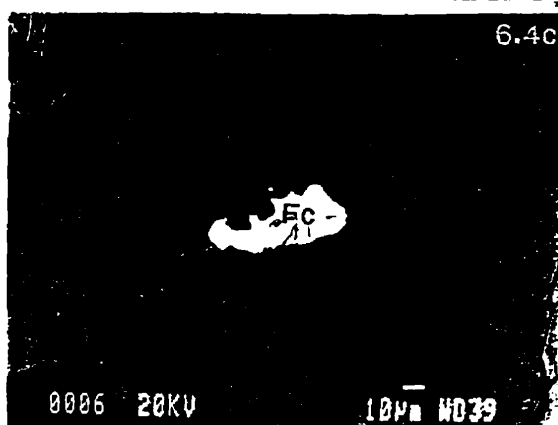
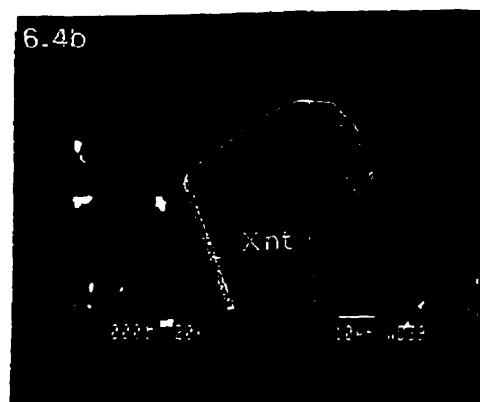
Some of the zircon grains in albitized biotite granite and greisen have equal concentration of Zr and Th in their EDS spectra, and also substantial levels of Hf and U. This may represent partial preservation of a metastable solid-solution between zircon and thorite (cf. Pointer et al. 1988a). Such zircon also contains relatively rare micro-inclusions of uranothorite. The zircon containing thorite, confined to albitized rocks and greisen, may result from the protracted interaction with a fluid phase, during which exsolution will be favoured (Mumpton & Roy 1961).

Thorite $\{(Th,Zr,Hf,U)SiO_4\}$ occurs in fayalite syenogranite, albitized biotite granite, and greisen. Thorite with micro-inclusions of bastnäsite occurs along the margin of allanite located in veinlets within hypersolvus syenogranite. In albitized biotite granite (Fig. 6.4e) and greisen (Fig. 4.4d), the thorite is invariably intergrown with biotite.

6.2.5 Fluorite

Fluorite (CaF_2) is ubiquitous, associated with altered fayalite, annitic mica, monazite, and fluocerite in hypersolvus syenogranite. Fluorite forms interstitial grains (Figs. 6.2f, 6.3d) in "hornblende"-biotite granite, or occurs as inclusions in biotite within coarse-grained biotite granite (Figs. 6.3c), and albitized biotite granite (Fig. 6.4c). The interstitial fluorite in subvolcanic rocks and the inclusions in biotite within pristine biotite granite are considered late magmatic (i.e.,

Fig. 6.4: SEM-BSE images of other accessory minerals (ilmenite, xenotime, fluocerite, columbite-cassiterite, zircon-thorite, and an unknown Th-REE mineral): **a)** Trellis-type intergrowth between ilmenite (Ilm) lamellae (dark) and magnetite (Mag) within biotite (Bt) in hypersolvus biotite granite (sample 83). This has the same significance as indicated for occurrences in "hornblende"-biotite granite porphyry. The bright grains around the intergrowth are fluocerite with rare fergusonite. **b)** Rhythmically zoned xenotime (Xnt) in biotite within fine- to medium-grained biotite granite (sample 13). The outer surface of the xenotime appears to be partly corroded. **c)** Biotite (Bt) grain with an inclusion of fluorite (Fl), which in turn contains an inclusion of fluocerite (Fc) within albitized biotite granite (sample 23C). Topaz (Toz) occurs in the lower left corner of the field of view. **d)** Zoned columbite - tantalite (Col) intergrown with cassiterite (Cst) in albitized biotite granite (sample 23C). The larger grain of columbite reveals at least four stages of growth. The inner (darker) zones are Nb- and Mn-rich, whereas the outer (brighter) overgrowths are enriched in Ta and Ti. This suggests dissolution and reprecipitation by surges of corrosive hydrothermal fluids. Ta and Ti might have been preferentially mobile over a short range, and were precipitated at later stages. **e)** Zircon (Zrn) displaying patchy zones and different degrees of metamictization in albitized biotite granite (sample 23C). Inclusions of thorite (Tht) have micro-inclusions of columbite, and the thorite grains are zoned with Th-rich and Zr $\pm$ Hf-rich areas. The boundary between zircon and feldspar (+ quartz, and biotite) experienced radiation damage, which is more pronounced where thorite inclusions occur. This boundary is also infilled by Fe-oxides. **f)** Surface of reddened alkali feldspar within a drusy cavity of coarse-grained biotite granite (sample 26D). The rough surface is a coating of Fe-silicate (possibly a clay), whereas the euhedral projections from the feldspar surface are due to an unidentified Th-rich REE mineral, possibly törnebohmite, as inferred from EDS spectrum. (*All samples in Figs. 6.4a - f are from the Jubga-Ziem center, and the sample locations are given in Fig. 2.1).*



deposited from the silicate melt late, in interstices), as are the inclusions in coarse-grained biotite granite, whereas in the altered biotite granite, the fluorite is clearly hydrothermal, as shown by its replacement of mica. A rare and unusual trellis-type intergrowth of fluorite and lamellar ilmenite has been mentioned earlier (Fig. 6.3e).

6.2.6 Other accessory minerals

Most of the other accessory minerals (Table 6.1) are very scarce.

Apatite $\{(Ca,REE)_3[(P,Si)O_4]_3(OH,F)\}$ occurs as micro-inclusions in chevkinite within the Buri ring dyke (rhyolite porphyry and hedenbergite fayalite granite porphyry), and hastingsite granite porphyry. Generally, apatite does not form a prominent accessory phase, reflecting the low P contents of these rocks (Chapter 2). Thus, the P content consistently remained below the saturation level for apatite, which is known to be more soluble in peraluminous melts (Watson & Harrison 1984).

Fluocerite $\{(La,Ce)F_3\}$ (Appendix VI.1c) occurs around the margin of ilmenite grains in fayalite syenogranite (Fig. 6.3a), along an interface boundary between fluorite and biotite in "hornblende"-biotite granite porphyry (6.2f), or associated with the trellis-type magnetite - ilmenite intergrowth and biotite in hypersolvus biotite granite (Fig. 6.3a). Another prominent occurrence of subhedral fluocerite is as inclusions in mica in coarse-grained biotite granite (Fig. 6.3c, Appendix VI.3f). The foregoing types of occurrence span the late magmatic to subsolidus range of crystallization. Fluocerite is also common in albitized biotite granite in association with fluorite (Fig. 6.4f) and greisenized zones. In these examples, the fluocerite is clearly hydrothermal. *Parisite* $[Ca(REE)_2(CO_3)_3F_2]$ (Appendix VI.3f) is limited to coarse-grained biotite granite, where it replaces fluocerite (Fig. 6.3c).

Bastnäsite $[REE(CO_3)F]$ (Appendix VI.1d) occurs as discrete grains that are ubiquitous

in some marginal samples of "hornblende"-biotite granite porphyry (sample 79); it may contain inclusions of allanite (sample 57) or skeletal zircon (sample 79). Such grains are interpreted as being late magmatic. Bastnäsite also occurs as inclusions in mica, thorite and allanite within hypersolvus syenogranite, whereas Th-bearing bastnäsite replaces allanite in hastingsite granite porphyry and "hornblende"-biotite granite porphyry. The latter type is interpreted to be of subsolidus origin, like the bastnäsite associated with fluocerite and parisite (Fig. 6. 3c).

Fergusonite $(LREE,Y)(Nb,Ta,Ti)O_4$ (Appendix VI.2a) is associated with ilmenite (+ fluocerite) in hypersolvus syenogranite (Fig. 6.3a), and with magnetite-ilmenite intergrowths in hypersolvus biotite granite (Fig. 6.4a). Fergusonite is rare; it also seems to span the late magmatic to subsolidus range.

Monazite $(La,Ce,Nd,Y,Ca,Th)PO_4$ occurs as inclusions in allanite in hypersolvus syenogranite (Fig. 6.2d) associated with the magnetite + annitic biotite + quartz assemblage (Fig. 6.2f), and as disseminations in biotite within textural variants of biotite granite (Fig. 6.3f). The occurrence of monazite as inclusions in biotite and a host of other minerals indicates that it crystallized early as well as late. In greisen, Th-bearing monazite occurs as discrete grains, and as micro-inclusions in zircon and Nb-bearing rutile. The discrete grains (of monazite), and micro-inclusions in rutile suggest crystallization from a hydrothermal fluid, whereas the occurrences along microveins in the greisens are due to post-greisen hydrothermal vein-filling.

Xenotime (YPO_4) is very limited in occurrence (Table 6.1), mostly in medium-grained biotite granite as discrete grains that are rhythmically zoned (Fig. 6.4b). The xenotime (Appendix VI.2b) also occurs as inclusions in zircon along with REE-rich apatite in the same thin section.

Topaz $\{Al_2SiO_4(F,OH)_2\}$ occurs in drusy cavities in miarolitic biotite granite, as prismatic grains with pyramidal terminations. Most of the topaz occurs as euhedral to massive grains in alteration assemblages in albitized biotite granite (Fig. 6.4c) and greisens. Colourless and bluish

green varieties are encountered (Appendix VI.4). The most extensive occurrence of alluvial topaz is found along the eastern flank of the Jubga-Ziem center. The provenance of this alluvial topaz (and cassiterite) is the drusy biotite granite in the upper zones of the Ziem ridge and the mineralized granites around the fine- to medium-grained granite of this center. Aquamarine (beryl) is limited to pegmatitic stringers in the Panshanu center, where it is exploited (Appendix VI.4).

Cassiterite (SnO_2) occurs in albitized biotite granite and greisen. In the albitized biotite granite, the cassiterite is invariably intergrown with columbite-tantalite (Fig. 6.4b). Cassiterite is of major economic interest, as indicated by alluvial mining in the complex for over eight decades.

Columbite - tantalite $(\text{Nb,Ta,Ti})_2(\text{Fe,Mn})\text{O}_6$ also occurs in albitized biotite granite and greisen. The columbite (Appendix VI.2c) shows a corroded Nb-, Fe, and Mn-rich core truncated and surrounded by later Ta- and Ti-rich zones (Fig. 6.4b, Appendix VI.5). Zoned crystals of columbite - tantalite have darker inner parts with higher Nb and Mn, and brighter outer parts with more Ta and traces of Ti (tantalite). The implications of these features are discussed below.

An *unidentified REE-Th-silicate* (Appendix VI.2d) is limited to miarolitic cavities in coarse-grained biotite granite. It occurs on K-feldspar surfaces as euhedral grains that attain 100 x 200 μm (Fig. 6.4f). The REE-Th-silicate was deposited from a hydrothermal fluid onto the surface of euhedral microcline in a drusy cavity. The SEM-EDS spectrum and results of semiquantitative analyses suggest that the mineral may be either yttrallite $\{(\text{Y,LREE,Th})_2(\text{Si}_2\text{O}_7)\}$, with Fe-rich clay contributing some Al to the spectrum, or possibly a Th-rich t rnebohmite $(\text{La,Ce})_2\text{Al}(\text{SiO}_4)_2(\text{OH})$.

6.3 Discussion

The different modes of magmatic crystallization of accessory minerals are first assessed. This is followed by a discussion of subsolidus processes that affected the accessory phases. The

textural characteristics discussed in this study include zoning documented in chevkinite, allanite, zircon, columbite-tantalite and thorite, and metamictization of chevkinite, zircon and thorite.

6.3.1 Magmatic crystallization of accessory minerals

A variety of models have been proposed for the crystallization of accessory minerals. This includes bulk saturation in melt, local buildup of incompatible elements at interfaces between phenocrysts of the major phases crystallizing and residual melt, and heterogeneous nucleation, such that accessory minerals nucleated on the surface of the phenocrysts so as to reduce the surface energy by reducing two crystal-melt interfaces to one.

The variation of trace elements that comprise the ESCs (essential structural constituents) of accessory minerals has been presented in Harker-type diagrams (Fig. 2.9 - 2.11), and the REE-variations presented in chondrite-normalized plots (Fig. 2.12). In these presentations, it was shown that concentrations of Ti, Ba, Sr, Nb, Y, and LREE decrease, whereas those of Th, U, and HREE increase from subvolcanic to plutonic rocks. The noncorrelation between F-enrichment (or differentiation index) and ESCs (+ HFS) elements rules out their build-up by enrichment in residual F-rich melts or fluids as a result of fractionation from a common parental melt.

The chevkinite documented as a prominent accessory in some of the subvolcanic rock units of the Kwandonkaya complex is the second reported occurrence in the NYG province. The mineral is likely more widespread in the province and may have been misidentified as titanite or allanite. The petrographic evidence in the Kwandonkaya occurrence consistently indicates the chevkinite to be an early to late magmatic phase (Figs. 6.1a - f, 6.2a). This interpretation corresponds to that proposed for its occurrence in the Cenozoic ashbeds of the western United States (Izett & Wilcox 1968), the Devil's Slide fayalite quartz syenite near Stark, New Hampshire (Jaffe et al. 1958), Little Chief granite porphyry stock, California (McDowell 1979), Mt. Amiata

(trachytes and rhyolites) in Italy (Bergen 1984), and St. Kilda microgranite in Scotland (Harding et al. 1982). Despite the differences in the degree of enrichment of REE and Ti in samples 49 (rhyolite porphyry), 43 (hedenbergite fayalite granite porphyry), and 40A (hastingsite granite porphyry), as seen in Figure 6.5, they both crystallized chevkinite at an early (magmatic) stage. The levels of REE in chevkinite in the rhyolite porphyry are similar to those reported in chevkinite-bearing syenite from the Shira NYG complex (cf. Bennett et al. 1984).

Although allanite is related to epidote by the coupled substitution: $\text{Ca}^{2+}\text{Fe}^{3+} \rightleftharpoons \text{REE}^{3+}\text{Fe}^{2+}$, there is no evidence that the allanite replaced epidote by metasomatic reaction. Usually, magmatic allanite crystallizes in place of chevkinite if the melt has a relatively high concentration of Ca, and ΣREE , and low levels of Ti (Green & Pearson 1986, 1988). The crystallization of allanite was suppressed in the metaluminous rhyolite porphyry and hedenbergite fayalite granite porphyry owing to relatively low Al contents. In the hypersolvus syenogranite and "hornblende"-biotite granite porphyry, the allanite is predominant, mostly of late-magmatic origin, and appears to have resulted from build-up of REEs and Al in residual melt (Fig. 6.6). Allanite was still in its field of stability below the solidus, as it forms veinlets with biotite in hypersolvus syenogranite and "hornblende"-biotite granite porphyry. In the coarse-grained biotite granite, allanite occurs in subsolidus replacement assemblages, whereas in greisen zones, allanite has crystallized from the acidic fluid medium.

Both the ilmenite and magnetite in subvolcanic rock units are early magmatic, whereas the occurrences in biotite granite are late magmatic. This reflects the progressive depletion of Fe and Ti from the subvolcanic to plutonic rocks. The ilmenite core in the "sandwich-type" intergrowth (Fig. 6.3a) is interpreted as being magmatic and earlier than the trellis-type ilmenite-magnetite intergrowth (Figs. 6.3a, b), produced by oxidation-induced exsolution (cf. Haggerty 1991) and discussed below. The trellis-type intergrowth of ilmenite and fluorite (Fig. 6.3e) was

Fig. 6.5: Plots of REE and trace element data for cone-sheet units. The data are normalized to those of rhyolite porphyry (49A, Buri ring dyke). Cone sheet samples 61A (felsite porphyry), 64 (hedenbergite hastingite granite porphyry), and 40A (hastingite granite porphyry) are located on Fig. 2.1. Of all the cone-sheet units, only sample 40A is slightly LREE- and Eu-enriched compared to the rhyolite porphyry. Also, the cone-sheet units are depleted in all elements except Th, U, Sr, and Ni.

Fig. 6.6: Plots of REE and trace element data for ring dyke units. The data are normalized to those of rhyolite porphyry (49A, Buri ring dyke). Buri ring dyke: Sample 43 (hedenbergite fayalite granite porphyry), Jubga-Ziem ring dyke: samples 58 (hypersolvus fayalite syenogranite), 56 and 68 (hornblende biotite granite porphyry). Samples are located in Figure 2.1. Whereas all the samples are slightly HREE-depleted (with the exception of sample 58), only samples 43 and 56 are slightly LREE-enriched.

Fig. 6.7: Plots of REE and trace element data for pristine biotite granite. The data are normalized to those of coarse-grained biotite granite (26A). Samples 1B and 36A (porphyritic biotite granite), 30A and 37 (drusy biotite granite). The biotite granite are depleted in all elements (relative to the coarse-grained biotite granite) except Rb, Th, U, Ta, Yb, Lu, and Ni.

Fig. 6.8: Plots of REE and trace element data for altered biotite granite units. The data are normalized to those of coarse-grained biotite granite (26A). All the samples (35A, 23C: albitized biotite granite; 36B: greisen) are depleted in all elements (relative to the coarse-grained biotite granite), with the exception of Rb, Th, U, Ta, Yb, Lu, and Ni.

In Figures 6.5 and 6.6, data are normalized against sample 49A because it is the least aluminous subvolcanic rock ($ASI = 0.96$), least affected by subsolidus re-equilibration, and it contains magmatic accessory minerals. In Figures 6.7 and 6.8, the data are normalized against sample 26A, which is one of the most pristine samples of biotite granite.

Fig. 6.5

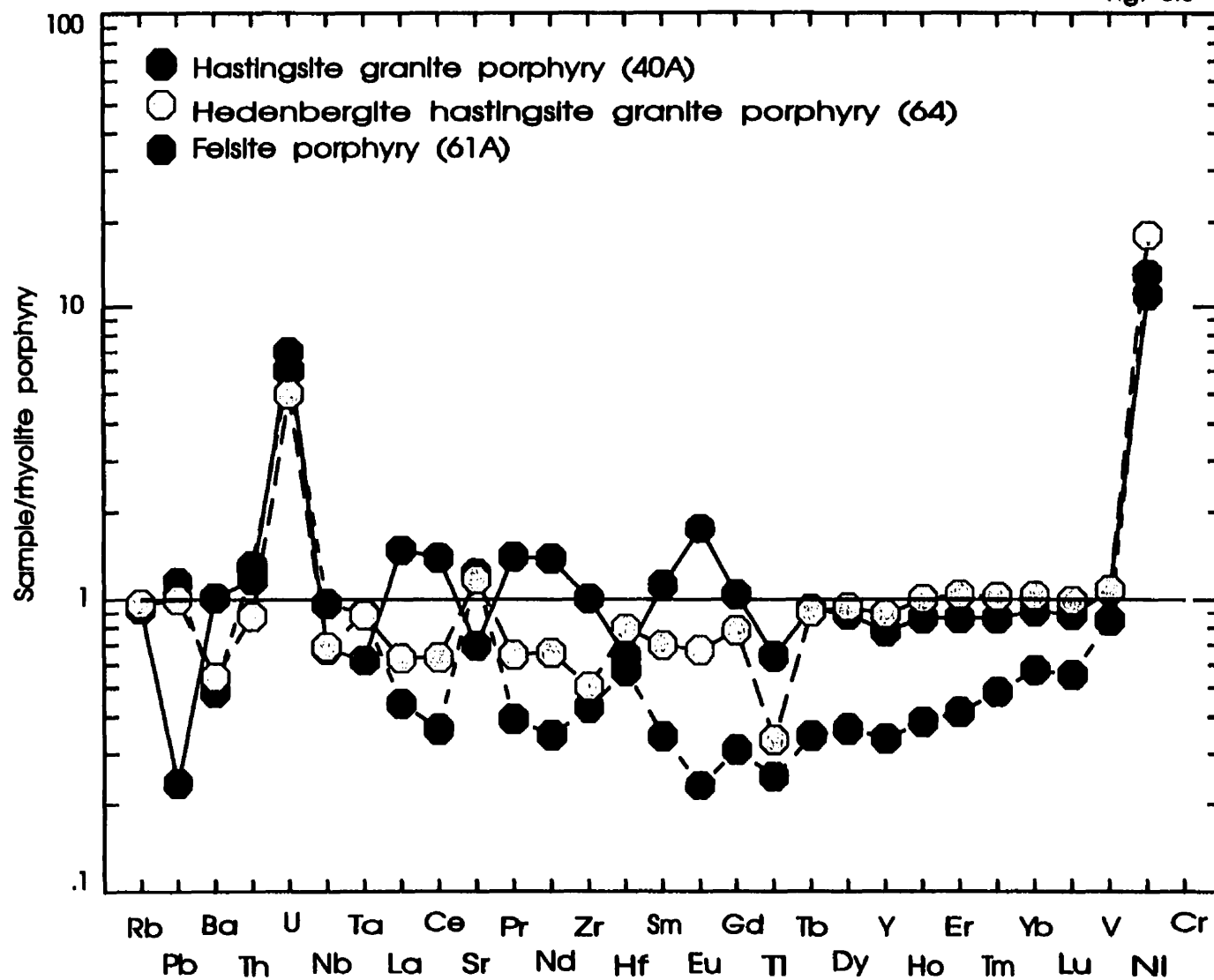


Fig. 6.6

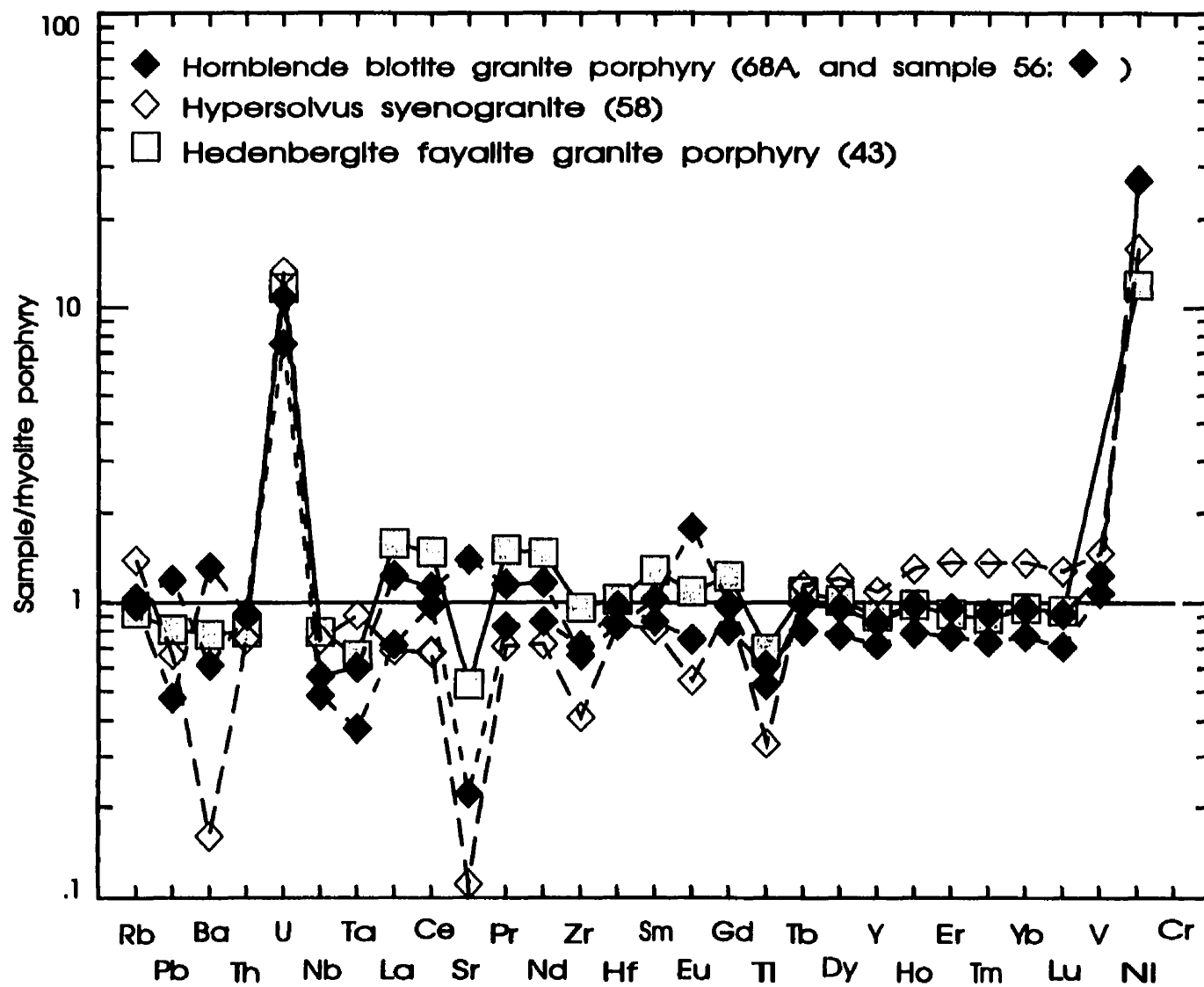


Fig. 6.7

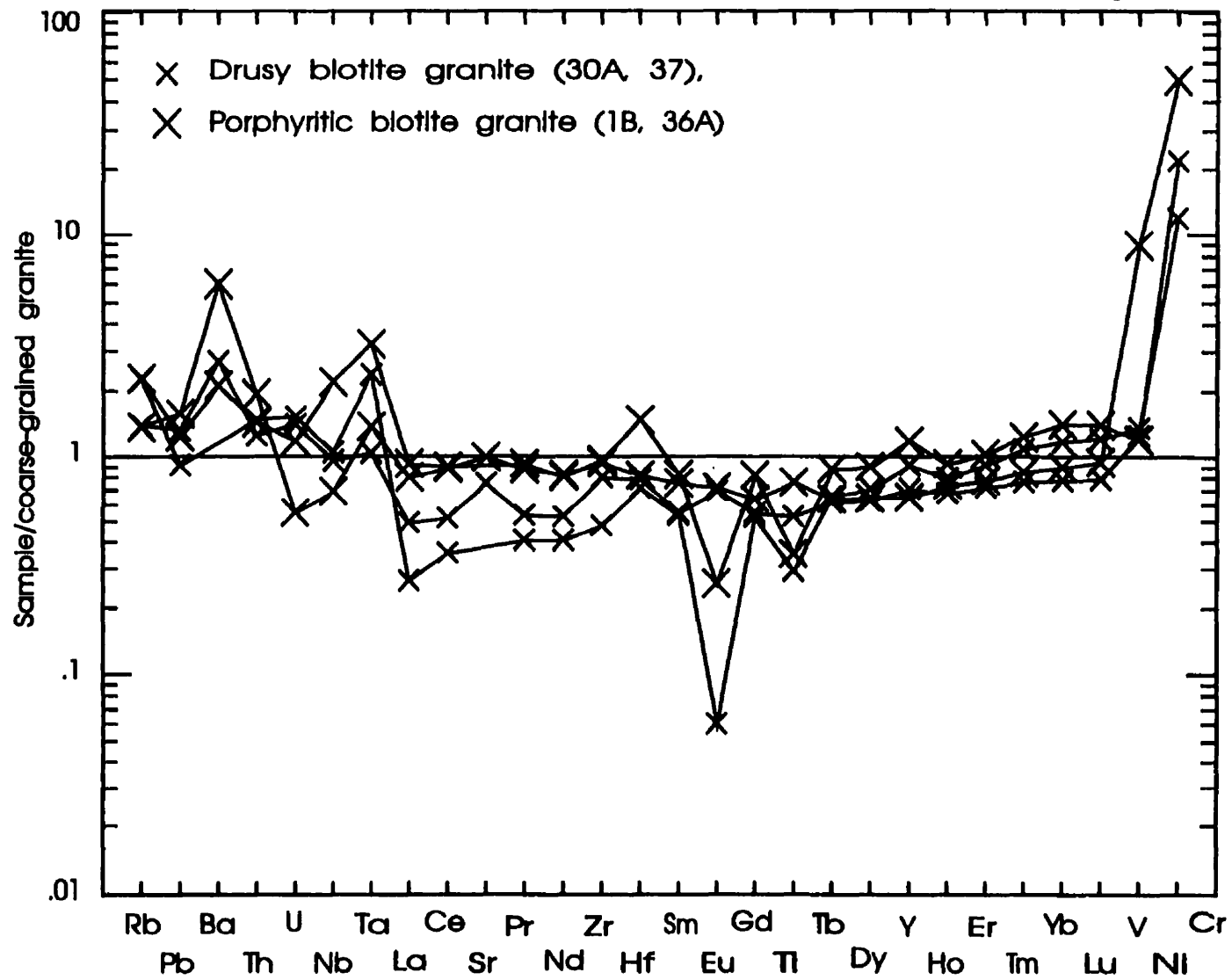
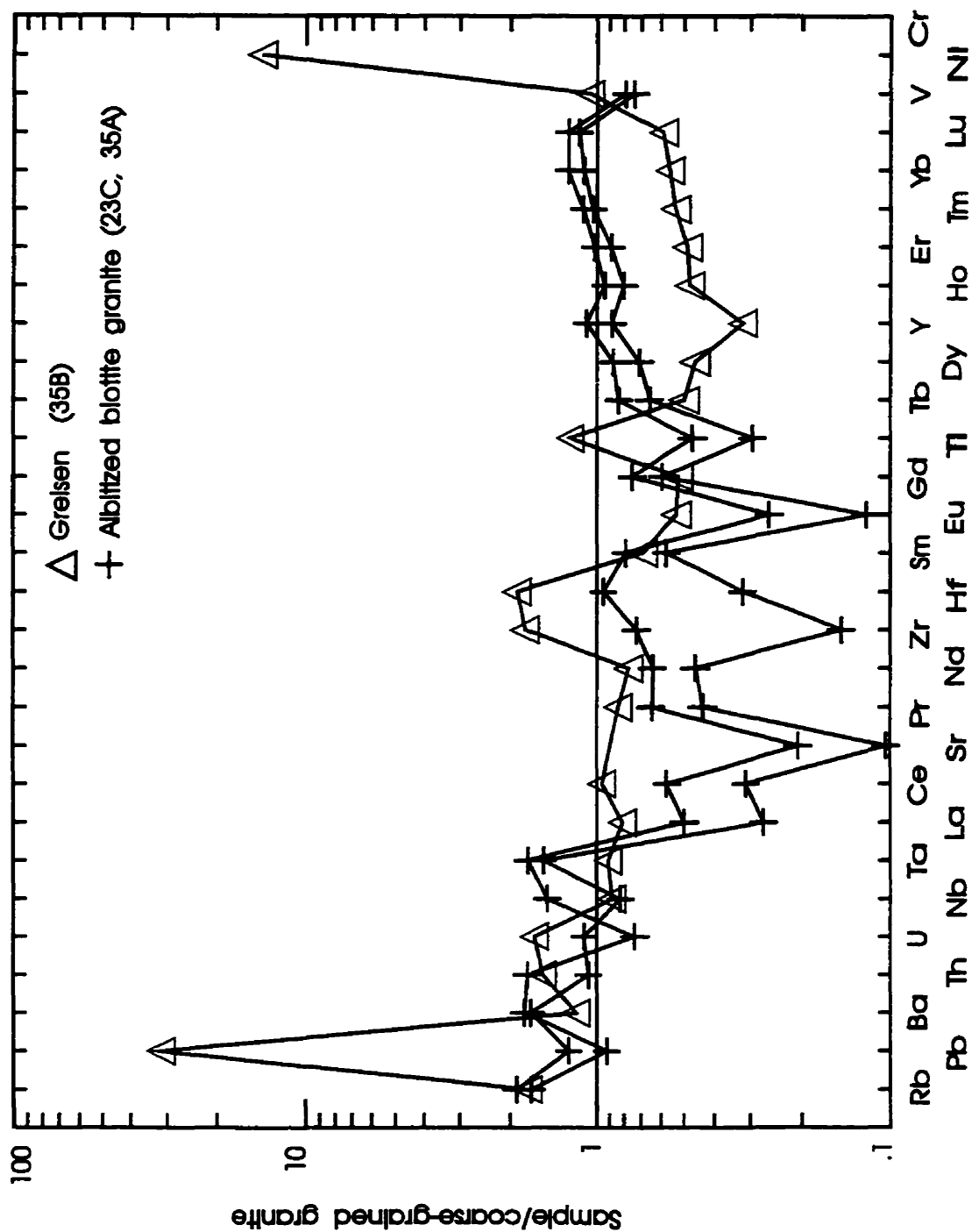


Fig. 6.8



produced at a late-magmatic stage by the preferential growth of the ilmenite along {111} planes of the fluorite, similar to the case of the ilmenite-magnetite intergrowths (cf. Haggerty 1991).

Zircon crystallization also spanned the early to late magmatic range. The subvolcanic rocks have Zr content of 108 to 622 ppm. This is above the zircon saturation value of 100 ppm in mildly peraluminous melts near the quartz-orthoclase-albite plane (Watson 1979, Watson & Harrison 1984). It is known that Zr concentrations greater than 100 ppm in felsic (non-peralkaline) rocks generally do not represent Zr abundances in melt, but a composite value of melt + zircon crystals (Watson & Harrison 1984). The Zr content of the metaluminous to peraluminous melt(s) that produced the subvolcanic rocks in the Kwandonkaya complex seems to have been kept above the zircon-saturation value, owing to generation of the parental magma from a Zr-enriched source. Zr was subsequently depleted in the source during the production of successive batches of melt, hence the biotite granites have lower Zr content (pristine biotite granite: 81-170 ppm); in this case, zircon crystallized mostly at a late magmatic stage (with or after interstitial biotite). It is also known that zircon saturation in the melt is affected by the F content, owing to increased solubility of Zr and HFSE (Keppler 1993). This factor also was likely an important factor in the crystallization of zircon (and other accessory minerals) at an early magmatic stage in the relatively F-poor subvolcanic rocks, and at a late magmatic stage in F-rich plutonic rocks. The greisen samples show evidence of hydrothermal enrichment in Zr (85-302 ppm) relative to the pristine biotite granite.

Other accessory minerals that are inferred to have crystallized as a result of saturation in the residual melt(s) include fluocerite, fergusonite, monazite, and fluorite. The crystallization of fergusonite may be related to slight enrichments in HREE within hypersolvus syenogranite and "hornblende"-biotite granite porphyry. Fluocerite, on the other hand, seems to have crystallized in REE- and F-rich residual melts that formed some subvolcanic rocks and biotite granites. Where

P and REE saturation are achieved, monazite occurs in place of chevkinite and allanite (Green & Pearson 1988).

The similarity in the composition and size of free microphenocrysts and inclusions of accessory phases (e.g., zircon, chevkinite, and allanite), and the fact that empirically measured K_D values of REEs (and a host of ESCs) are generally greater than 1 for most ferromagnesian minerals (e.g., at La Primavera and Bishop Tuff high-silica rhyolites: Mahood & Hildreth 1983) all favour the formation of such accessory minerals by wholesale saturation in the melt. However, Figures 6.5 - 6.8 do not give an indication of any direct relationship between accessory mineral crystallization and degrees of ESC enrichment in all cases, and therefore necessitates the exploration of other mechanisms to explain the formation of some of these accessory minerals.

6.3.2 *Local saturation*

Some accessory minerals that crystallized at a late magmatic stage and are enclosed in other accessory minerals owe their existence to local build-up of and saturation in incompatible elements at the interface between host crystals and melt. The local saturation involves a change in liquid composition, such that supersaturation in the accessory phases occurs at the crystal/liquid interface accompanied by thermal and constitutional undercooling (Bacon 1989). The supersaturation can result from the development of a differentiated boundary-layer because of the exclusion of the ESCs from the phenocrysts and their slow diffusion away from the interface. The effect described in the foregoing is referred to as the "snow-pile effect" (Harrison & Watson 1984). If $K_D^{x_{tal}/melt} < 1$, then ESCs will be significantly enriched at the interface, and nucleation may be initiated. In this situation, boundary-layer saturation can account for the inclusion of some accessory minerals in major phases.

In the Kwadonkaya complex, where accessory minerals are invariably associated with

the ferromagnesian minerals (inferred to have $K_D^{xtal/melt}$ for ESCs > 1), a decrease in the solubility of the ESCs, as shown by Michael (1988), with increasing boundary-layer differentiation, seem to have been an overriding factor. This might have resulted in saturation when such ferromagnesian minerals were still liquidus phases, and explains why the accessory minerals are more commonly included or attached to the ferromagnesian rather than salic phases having lower K_D for ESCs ($K_D^{xtal/melt} < 1$). Furthermore, accessory minerals might have nucleated on the surface of the phenocrysts so as to reduce surface energy by the conversion of two crystal/liquid interfaces into one (Bacon 1989).

For apatite saturation in the melt, between about 0.01 up to 0.15 wt% P_2O_5 are required (Harrison & Watson 1984). The rock suite of the Kwadonkaya complex lies on the lower end of this range (0 to 0.04 wt.% P_2O_5), except for rhyolite porphyry, with 0.13 to 0.16 wt.% P_2O_5 . This may explain why apatite is not a commonly observed accessory phase, although it is documented in other ring complexes. The apatite occurring as inclusions in chevkinite, zircon, and K-feldspar is interpreted to be due to local saturation. The apatite occurring as needle-like inclusions in chevkinite is similar to the occurrence from the Cenozoic ashbeds of the western United States (Izett & Wilcox 1968), and recalls observations that chevkinite in most natural occurrences is usually accompanied by a small amount of apatite (Ito & Arem 1971).

6.3.3 Zoning in some accessory minerals

Petrographic observations and SEM studies show chevkinite, allanite, and columbite to display different scales of zoning (Figs. 6.1, 6.2, 6.4d). Zoning in general indicates nonequilibrium crystallization from parental melt or fluid. The REE zoning in chevkinite and allanite could be caused solely by in situ fractionation-induced changes in the melt (cf. Sawka et al. 1984, Lahti 1987). For example, reversals in LREE partitioning trends, which usually result from changes in

the melt structure due to increasing volatile contents. Volatile constituents such as F and water break T-oxygen bonds, and distort the framework of the aluminosilicate melt, thereby introducing larger sites and hence allowing larger highly charged cations (e.g., the REE) to more readily remain in the melt (Sawka et al. 1984, Manning & Pichavant 1988, Keppler 1993).

Oscillatory zoning in chevkinite, allanite, and columbite (Figs. 6.1, 6.2a-d, and 6.4d) is a function of diffusion and interface kinetics between growing crystals and magma or fluid (Haase et al. 1980, Allègre et al. 1981, Lahti 1987, Halden et al. 1993). The observed oscillatory zoning can be produced either by cyclical changes in partition coefficients corresponding to fluctuations in temperature or melt composition, or by variations in the rate of crystal growth. As noted by Halden et al. (1993), the frequency and extent of variations in temperature and K_D values that will be required to produce thousands of oscillatory zones seem to be physically unreasonable. From petrographic and SEM observations of accessory minerals in the Kwadonkaya complex, it seems more likely that oscillatory zones were produced by different rates of diffusion of components to the surface of a growing accessory phase. One requirement for the development of oscillatory zones, emphasized by Halden et al. (1993), is that two components involved in the diffusion processes have to be coupled. This requirement has been documented in oscillatory zoning and replacement textures in columbite-tantalite grains from Eräjärvi area, southern Finland (Lahti 1987), where zoning is also due mainly to variations in Ta and Nb, and oscillations in Mn follow Nb. This precondition is best illustrated by zoned grains of columbite in the Kwadonkaya complex (Fig. 6.4c, Appendix VI.5), which exhibit a corroded core of early Nb- and Mn-rich columbite overgrown by later zones rich in Ta and Ti. Fe in these examples does not vary noticeably. This is odd, because with increasing fractionation, Mn^{2+} and Ta should both increase outward at the expense of Fe^{2+} and Nb.

6.3.4 *Subsolidus - hydrothermal remobilization and enrichments.*

Patchy zoning (in chevkinite and allanite), superposed on oscillatory zoning (Figs. 6.1a - f, 6.2a - d), is interpreted as a product of replacement (cf. Lahti 1987). Such a pattern of replacement could have occurred at a late magmatic to subsolidus stage, and implies the subsolidus mobilization of the rare earths. The occurrence of allanite along veinlets in some subvolcanic rocks indicates local hydrothermal enrichment of the REE. It seems that a LREE-rich residual volatile phase formed at the late stages of crystallization of some subvolcanic rocks, mostly in the Jubga-Ziem center, in a way comparable to the case of the Skye granites (Exley 1980), and the alkaline granites intruding the Arabian Shield (Drysdall et al. 1984). Alternatively, such LREE-enriched fluid might have been exhaled during the degassing of underlying biotite granite, with the fluid subsequently interacting with overlying subvolcanic rocks.

The trellis type intergrowth (of ilmenite and magnetite) are produced by ilmenite preferentially growing, during oxidation-exsolution (Fig. 6.3a), along {111} planes of the host, because both the {111} planes of spinel and the {0001} planes of ilmenite are similar in packing of oxygen atoms (Haggerty 1991). It is most likely that the intergrowth documents progressive drop in T and increases in $f(\text{O}_2)$ from the QUIF buffer (during which ulvöspinel-magnetite_{ss} was formed) to QFM (Usp-Mag_{ss} → ilmenite + magnetite, i.e., $6\text{Fe}_2\text{TiO}_4 + \text{O}_2 = 6\text{FeTiO}_3 + 2\text{Fe}_3\text{O}_4$) in the subsolidus.

The thorite inclusions in zircon are presumably products of exsolution from the zircon host following the metastable growth of Th- and U-rich zircon (cf. Rubin et al. 1989, Pointer et al. 1988a, b). As above, patchy zoning in zircon and thorite in albitized biotite granite and greisens (Fig. 6.4e) suggests hydrothermal mobilization of Zr and Th. Such mobility was on the micrometer scale (cf. Pointer et al. 1988b).

The patchy zoning and overgrowths in columbite (Fig. 6.4d, Appendix VI.5) are explained

by the generation of a corrosive vapour phase at the solidus, which caused resorption, patchy zoning, and the partial replacement and overgrowth of earlier columbite (cf. Lahti 1987).

6.3.5 Metamictization of chevkinite, allanite, zircon and thorite.

SEM-EDS spectra show chevkinite, allanite and zircon to contain detectable amounts (0.1 to 6%) of U and Th. These elements can undergo radioactive decay and consequently affect the structure of host minerals. The end products of radioactive decay of these radionuclides are isotopes of lead (Pb) produced by α -, β -, and γ -decay events (Hawthorne et al. 1991), which induce metamictization in nearby accessory minerals. Microfractures develop, within accessory phases, in the high α -decay dose zones because of expansion of layers enriched in radiogenic U and Th (Hawthorne et al. 1991). Such systems of microfractures provide pathways for addition or removal of U, Th, and Pb ($\pm$ Zr, $\pm$ REE). It is observed in the Kwandonkaya complex that: i) chevkinite occurring as inclusions in amphibole, and in the groundmass is metamictized along the grain margin, whereas that enclosed in orthoclase microperthite is not (Fig. 6.1c to f), and ii) zircon and thorite in albitized biotite granite and greisens are metamictized to a higher extent than those in the pristine rock units. These observations illustrate well the fact that the presence of trace amounts of water or hydroxide ion can dramatically affect the response of minerals to radiation from external and internal sources (Aines & Rossman 1986). Loss of volatiles to produce drusy facies in rock units, within the Kwandonkaya complex, was invariably accompanied by hydrogen loss and oxidation (cf. chapter 4); thus, the mobility and reactivity of the H species in silicates, as indicated by Aines & Rossman (1986), may explain the variations in degree of metamictization within chevkinite, allanite, and zircon in the Kwandonkaya rock units.

6.4 Summary and conclusions

1) The crystallization of early magmatic accessory minerals like chevkinite, zircon, and ilmenite was controlled by saturation in the melt.

2) The late-magmatic inclusions in amphibole and biotite were produced by build-up of ESCs during crystallization or owing to decrease in solubility of ESCs in a differentiated boundary-layer along the crystal-melt interface.

3) In most instances, the residual F-rich melt, from which ferromagnesian minerals (amphibole and biotite) crystallized, coprecipitated the accessory phases in some subvolcanic rocks and biotite granites.

4) Overall, the accessory mineral suite in both subvolcanic and plutonic rocks is typical of metaluminous-peraluminous granites that, in other localities of NYG and A-type granites, have been interpreted as affected by alkali metasomatism. However, unlike some of the previous interpretations of the accessory mineral suites, those in the Kwandonkaya complex span both the early to late-magmatic and subsolidus (hydrothermal) range.

5) Zoning in most cases is primary and tends to the oscillatory type. Patchy zones are secondary and indicate intra-mineral mobility of some HFS elements (e.g., REE, Ta, Ti, and Nb) at subsolidus-hydrothermal stages. Patchy zoning in chevkinite and allanite as well as columbite-tantalite, and trellis-type intergrowth of ilmenite in magnetite, was produced by subsolidus modifications. On the other hand, allanite along microveins in subvolcanic rocks indicates local hydrothermal enrichments of REE.

6) The hydrothermally affected units are not unusually enriched in most of the trace elements and REE. Therefore, the accessory and ore-mineral suite in these rock types resulted from local dissolution and redeposition.

Chapter 7: Evidence of magma mixing, fractionation, and commingling of magmas - a textural, mineral chemical, and petrochemical study of enclaves.

7.1 Introduction

The Kwandonkaya igneous complex, located on the eastern edge of Jos plateau (Fig. 7.1), in the Nigerian Younger Granite (NYG) province (Fig. 1.1), preserves evidence of magma mixing as an important petrogenetic process. This topic has received concerted attention (Anderson 1976, Whalen & Currie 1984, Cantagrel et al. 1984, Didier & Barbarin 1991, Castro et al. 1991), but has not been properly addressed in the NYG province.

Occurrences of coeval mafic and felsic material are common in volcanic-plutonic terranes (Didier 1973, 1991, Barbarin et al. 1984, Whalen & Currie 1990, Barbarin & Didier 1991,); mafic enclaves could represent residual material (restite) (Chappell & White 1991) produced by partial melting (Tindle 1991), result from the accumulation of fractionated phases (Tindle 1991), or indicate commingling, with or without mixing, of coeval mafic and felsic magmas (Castro et al. 1990, 1991). In most geological settings, the above processes are not exclusive. Tindle (1991) gave the most appropriate description of mafic enclaves: "mafic igneous material frozen in time, offering an insight into processes operating in magma chambers (or conduits tapping such chambers), and providing a natural laboratory for the study of partial melting, fractionation, magma mixing, and diffusion".

Field and petrographic studies in the NYG have previously documented the existence of coeval mafic and felsic lavas (Ike 1979, Turner & Bowden 1979), the occurrence of enclaves in subvolcanic rocks within ring dykes (Ike et al. 1984), and their presence in plutonic units (Didier 1973, Turner & Bowden 1979, Bonin 1991). However, few compositional data were provided, and the role of mafic rocks in the origin and evolution of the predominantly felsic units in the NYG is not well established, possibly owing to the relatively small volume and scarcity of

exposures of mafic and mixed rocks in the province. The only previously documented petrochemical evidence of mixing in the NYG emerged from studies of volcanic rocks in the Tibchi (Ike 1979) and Ningi-Burra complexes (Bowden 1985, Bowden et al. 1987), where mixing between basaltic and rhyolitic magmas was proposed to have produced "andesitic" lava.

In this paper, field-based studies, characterization of mineralogy and textures, and geochemical information on enclaves and their felsic host rocks (Chapter 2) are used to provide an interpretation of the dominant petrogenetic processes. Analytical techniques employed include scanning electron microscopy (SEM) with semiquantitative energy-dispersion spectrometry (EDS) and back-scattered electron imaging. These tools allowed the study of zoning in minerals and of reaction rims developed around enclaves or xenocrysts. Electron-microprobe analyses were undertaken to quantitatively assess chemical variations in the main minerals. Samples of enclaves were analyzed using X-ray fluorescence (XRF) for major and trace elements, and inductively coupled plasma - mass spectrometry (ICP-MS) techniques for the rare-earth elements (REE).

Definitions

To avoid confusion in discussions of the mixing process, the following definitions of terms are used in this paper. **Hybridization** is the process that may produce either mixed or mingled rock. **Magma mixing** involves the blending of two or more magmas and the physical distribution of the respective liquidus phases to produce a homogeneous magma (Castro et al. 1990). **Magma mingling** involves mixing that has not gone to completion, leaving a hybrid rock with identifiable end-members (Castro et al. 1990), and resulting in the formation of enclaves. **Enclaves** are rounded to subrounded magmatic fragments of foreign rocks included within another rock (Didier 1973), and are mostly mafic microgranular enclaves (MME). **Restite** is solid material that is residual, produced during partial melting, and possibly entrained in plutonic or volcanic

rocks (Chappell & White 1991).

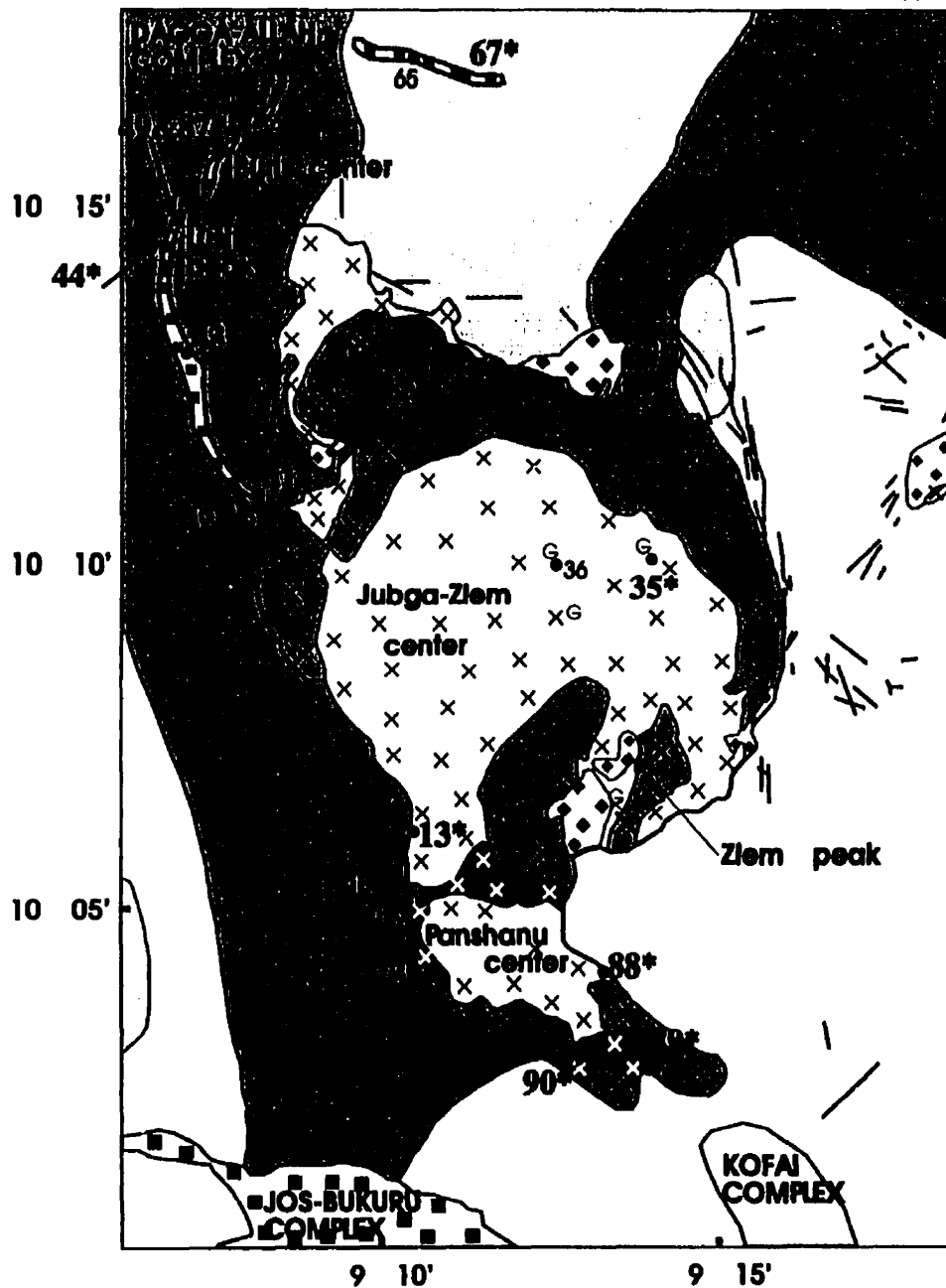
The nomenclature of ferromagnesian minerals follows the recommendations of the IMA subcommittees on pyroxene (Morimoto 1989) and amphibole (Leake 1978, Hawthorne 1983). The nomenclature of micas follows Bailey (1984a, b) and Černý and Burt (1984). Terms used to classify the enclaves pertain to the volcanic or microgranular equivalents, after the recommendations of the IUGS subcommittee on the systematics of igneous rocks (Le Bas et al. 1986).

7.2 Field relations of mixed rocks

The detailed field relations of felsic units in the Kwandonkaya complex have been presented earlier (Chapter 2). The complex is composed of three overlapping intrusive centers that have a NNW-SSE alignment: Buri, Jubga-Ziem, and Panshanu (Fig. 7.1). The Buri center (northernmost) is made up of rhyolite porphyry and hedenbergite fayalite granite porphyry. Of the two, the rhyolite porphyry preserves evidence for pervasive mingling; there is a consistent distribution of enclaves, which comprise 20 to 40% of the rock. Hedenbergite fayalite granite porphyry constitutes about 96% of total exposed area of Buri ring dyke; there are no enclaves in about 90% of the exposures, and up to 20% in the rest.

Whereas evidence for mixing is relatively rare in the subvolcanic and plutonic rocks of the Jubga-Ziem center, the reverse is true for the Panshanu center, a composite intrusive body of about 8 x 4 km, comprised mainly of textural variants of biotite granite. This center preserves the most widespread occurrence of enclaves in the complex. These examples are localized in marginal parts of the Panshanu center and range from less than 20 m² to over 500 m². In most cases, the enclaves do not show preferred orientation or sorting, but seem randomly distributed.

Fig. 7.1: Location of mixed rocks analyzed in this study from the Kwandonkaya complex. The locations where enclaves are plentiful are shown by an asterisk.



• Sample location

MESOZOIC GRANITES

G: Grelsens

▧ Basalt dyke

⊗ Fine- to medium-grained biotite granite

▨ Coarse-grained biotite granite

⊠ Porphyritic biotite microgranite

◻ Hornblende biotite-, biotite-granite porphyry, +/- fayalite syenogranite

◻ Hedenbergite fayalite granite porphyry

◻ Rhyolite porphyry

▧ Felsite porphyry, granite porphyry

PAN-AFRICAN BASEMENT COMPLEX

◻ Biotite hornblende granite

▨ Gneissic granite

◻ Migmatite

1 0 2 4 6 8 km

7.3 Petrography of enclaves

A summary of petrographic features of the enclaves is given in Table 7.1. Enclaves in rhyolite porphyry and granite porphyry in the Buri ring dyke are documented first, followed by enclaves in the biotite granite from the Panshanu center.

7.3.1 Enclaves in the subvolcanic rocks

Trachyandesitic enclaves occur in rhyolite porphyry and range from microscopic to over 10 x 25 cm (Fig. 7.2a, Appendix VII.1a). Many display various degrees of necking (Appendix VII.1b). In other instances, globules of the rhyolite porphyry seem enclosed by the trachyandesite enclaves (Fig. 7.2b, Appendix VII.1b, c). Some of the trachyandesite enclaves are disaggregated into microcrystalline hornblende, biotite, apatite, and Fe-Ti oxides, giving an impression of aerosol-type dispersal (Fig. 7.2c), also described as "emulsified facies" by Gourgard (1991). Such dispersal indicates incipient hybridization. Some examples of enclaves in rhyolite porphyry are zoned, with a fine-grained inner part and a mm-wide darker chilled margin (Fig. 7.2, Appendix VII.1a, d). The zones, in SEM-BSE scans, show the core of the enclaves to be relatively enriched in Ca, Ti, Fe, and Mg, whereas the outer zone is enriched in alkalis, SiO₂, Zn, and Mn. The ocellar quartz xenocrysts that have hornblende $\pm$ biotite surrounding the grains (Figs. 7.2e, f) may further indicate the occurrence of diffusion, or may simply illustrate sudden nucleation of the liquidus minerals on a cool surface.

The textural characteristics of **microsyenodiorite** enclaves (in hedenbergite granite porphyry) are illustrated in Figure 7.3a and Appendix VII.1e. Both ortho- and clinopyroxene occur as subhedral to euhedral microphenocrysts, as inclusions in plagioclase, and as microlites in the matrix (Figs. 7.3a, b). The grains of sodic plagioclase have a clear core and a dusty, brown K-rich mantle (Figs. 7.3c, d), and are replaced to various extents by sericite and epidote. The

Table 7.1: Summary of petrographic characteristics of the enclaves.

Enclave Type	Minerals present	Textures	Other features
Trachyandesite	Mtx: Hbl, Bt, Qtz, Or, Accessory: Ilm, Mag, Ap, Aln, Zrn, Xenocrysts: Hed, Qtz, Or	Microcrystalline, sparsely porphyritic, xenocryst-rich, ocellar rim of Hbl, Bt, and Fe-Ti oxides	Circular, ellipsoidal. Crenulate to sharp boundaries. Zoned enclaves, dispersed enclaves.
Microsyenodiorite	Phen & Mtx: Hd, Aug, Opx, sodic Pl, Or, Amp, Accessory: Ilm, Zrn, Ap, Mag, Fl Xenocrysts: Hd, Or, Qtz, <i>{Src, Epd, Chl, Bt, Amp, Mag}</i>	Microgranular, porphyritic, subophitic with essential minerals in a mesostasis of sodic plagioclase.	Chilled margin, rounded to ellipsoidal with less necking, prominent subsolidus alteration
Trachybasalt	Phen: Pl Mtx: Pl, Hbl, Bt, Qtz, Accessory: Ap, Fe-Ti oxides, <i>{Src, Epd, Chl, Bt}</i>	Porphyritic, sieve- textured plagioclase	Rounded to subrounded; $\mu\text{m-}$ to mm-wide Bt-rich margins, clots of Hbl replaced to various extents by Bt.
Orb-bearing basalt	Orb-nuclei: Pl, Qtz, Fl, Bt, Ilm, Mag; Mtx: Bt, Hbl, Pl, Fe- Ti oxides, Fl <i>{Src, Epd, Ab, Msc, Fl, Bt}</i>	Orbicular texture	Rounded, typically forms double enclaves, highly varied combination of minerals in orb nuclei
Plagioclase xenocryst	Megacrystic Pl (5-10 cm).	Monomineralic, lenticular outline.	Rare; have inclusions and ocellar rind of Hbl, Bt, and Ilm.
Andesite	Phen: Pl, Hbl, Mtx: Hbl, Bt, Qtz Accessory: Ap, Fe-Ti oxides, Zrn Xenocrysts: Pl, Mc, Qtz, <i>{Bt, Hem, Chl, Fl, Ep, Mag}</i>	Porphyritic, contain $\mu\text{m-}$ sized darker MMEs, and quartzofeldspathic clots (of Qtz, Mc, Bt) with ocellar rinds of Hbl + Bt	Globular to ellipsoidal; zoned; alteration halo
Dacite	Phen: Pl, Hbl, Mtx: Pl, Hbl, Bt, Qtz, Ap, <i>{Src, Ab, Bt}</i>	Porphyritic, microcrystalline, or xenocryst-poor	globular, ellipsoidal, elongate, zoned

Alteration assemblages are italicized in curly brackets; symbols of minerals after Kretz (1983), and described in Table 5.1. MME = mafic microgranular enclave.

Fig. 7.2: Field photographs, photomicrographs, and SEM - BSE images of trachyandesite enclaves (in rhyolite porphyry of the Buri ring dyke). **a)** Field photograph of rhyolite porphyry (Locs. 49, 87) showing submicroscopic to mesoscopic enclaves. The enclaves range from xenocryst-poor (T1) to xenocryst- and phenocryst-rich (T2) types. The xenocrysts are of the same size and habit as in the host rock of the enclaves, and thus are considered derived from the rhyolite porphyry. Also, the enclaves do not show any strong sorting or alignment, but rather are dispersed in what might have looked like an emulsion-type mixture of two melts before consolidation. **b)** Photomicrograph of a trachyandesite enclave (Trand.). The dark enclave in this case has a felsic inclusion (F1), which in turn has dispersed ferromagnesian clots ($\pm$ vesicles). These clots indicate the incomplete blending of two melts (trachyandesitic, rhyolitic). *Width of field of view is 3.1 mm.* **c)** Thin section photograph of rhyolite porphyry with trachyandesite enclaves (T3) illustrating various degrees of disaggregation. Some quartz xenocrysts (Qtz) have a rind of ferromagnesian minerals. The enclave disaggregation represents a more advanced stage of hybridization, which produces an emulsified texture similar to what has been described in the Dubre pelean nuées ardentes, in the Mont-Dore composite volcano, Massif Central, France (cf. Gourgaud 1991). *Mag: X2.5.* **d)** Thin section photograph of a zoned trachyandesite enclave (Trand.). The zoning, owing to variations in grain size, is not very obvious, but SEM-BSE scans indicate increases in Ca, Ti, Fe, and Mg toward the core and increases in alkalis, SiO₂, Zn, and Mn at the margin. Xenocrysts of quartz (Qtz^{*}) are similar in size and habit to those in the host rhyolite porphyry. This suggests the provenance of such xenocrysts in enclaves from the host rhyolite porphyry. **e)** SEM-BSE image of parts of Figure 7.2d. The quartz xenocryst (Qtz^{*}) has a rind of hornblende and biotite $\pm$ Fe-Ti oxides, which either formed by diffusion (uphill) or nucleation on a relatively cold surface. **f)** SEM-BSE image of an enlarged area in Figure 7.2e (near the bottom extremity of the quartz xenocryst).

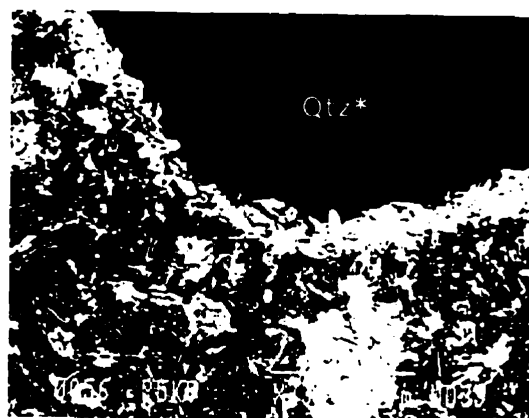
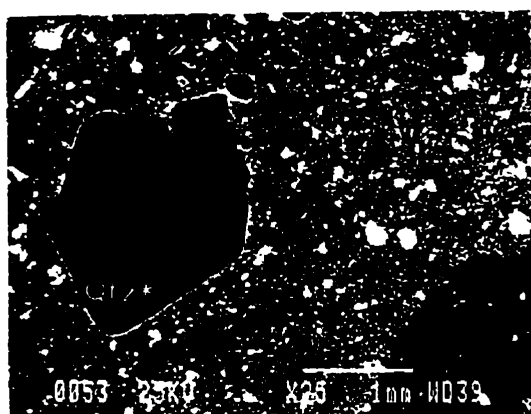
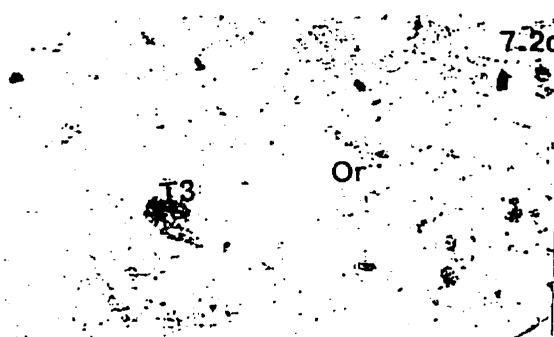
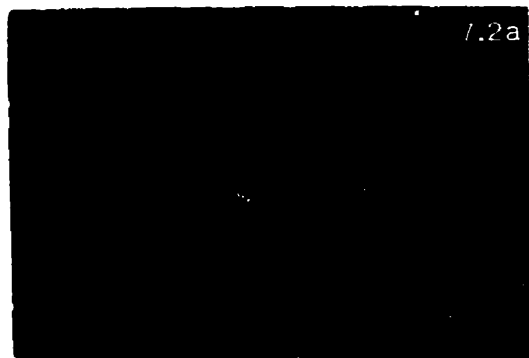
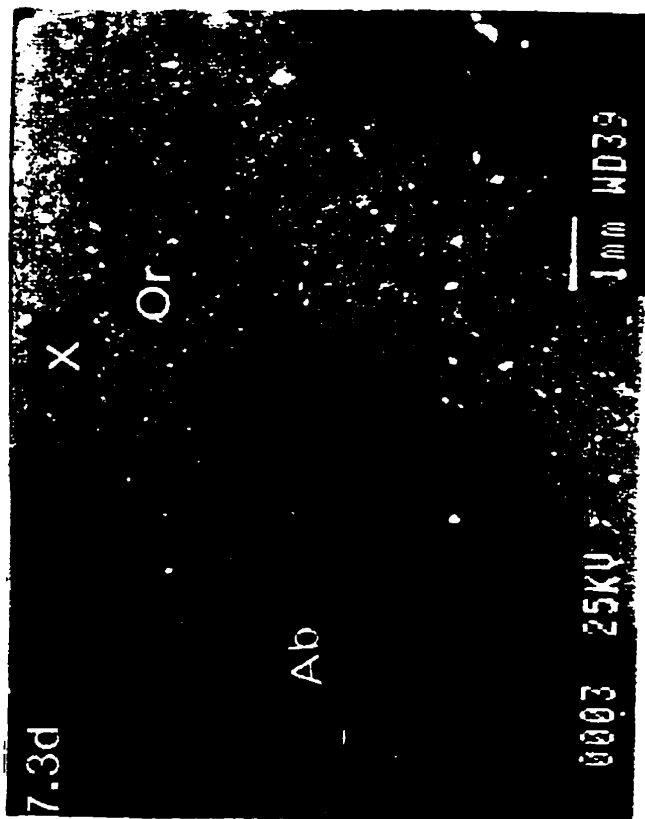
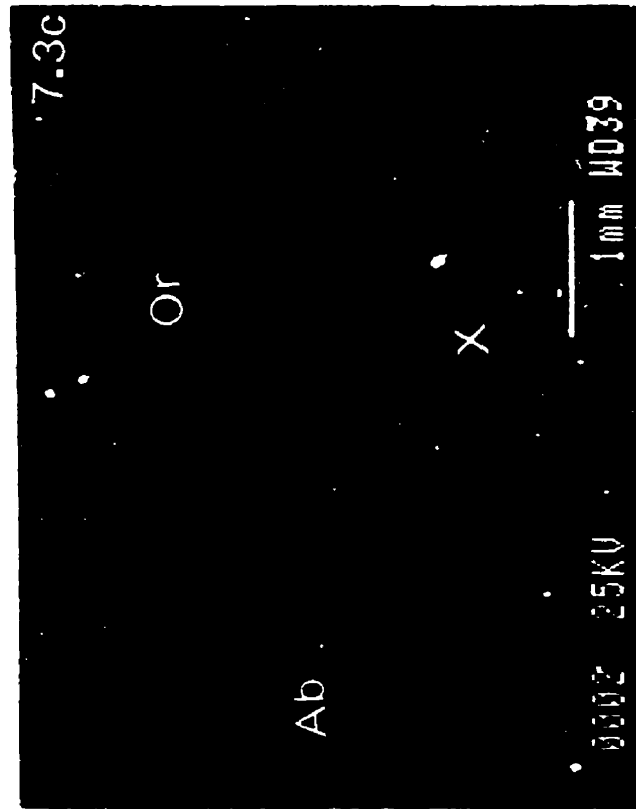
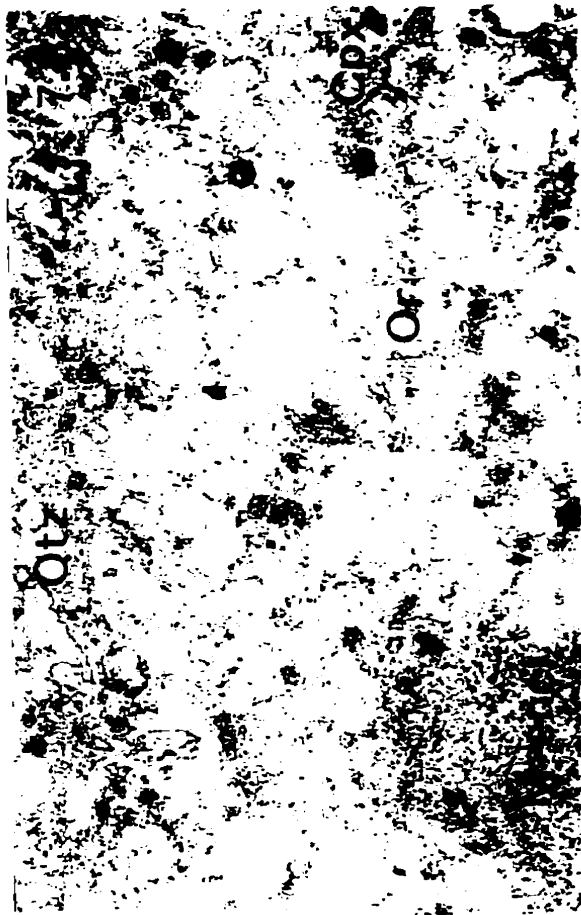


Fig. 7.3: Petrographic characteristics of microsyenodiorite enclaves (in the hedenbergite fayalite granite porphyry of the Buri ring dyke). a) Thin section photograph of a typical sample showing a microsyenodiorite enclave (M-encl) in the lower left corner (sample 67A). In this enclave and host rock, pale to dark green amphibole, chlorite, and Fe-Ti oxides replace the clinopyroxene (Cpx) to various extents. Qtz = quartz, Or = orthoclase microperthite, Cpx = clinopyroxene. *Magnification:* = X2.5. b) SEM-BSE image of microsyenodiorite enclave containing orthopyroxene (Opx) with inclusions of ilmenite (Ilm) and marginal replacement by amphibole (Amp). A separate amphibole grain (Amph) to the upper right corner is replaced along the fractures by chlorite. Annitic biotite (Bt) replaces hedenbergite to the lower left corner. [Brighter specks are other (unidentified) accessory minerals]. c: SEM-BSE image of darker grey sodic plagioclase with inclusions of quartz rimmed by brighter K-rich microperthite within microsyenodiorite enclave. d) SEM-BSE image of darker grey sodic plagioclase (Na-plag) displaying a complex relation with brighter K-rich microperthite within microsyenodiorite enclave. The darkest grey parts are quartz, and point marked X is at the same location as in Fig. 7.3c.

In Figs. 7.3b, c, and d, the dark (medium) grey areas are colourless parts of sodic plagioclase, whereas the brighter parts are turbid, K-rich microperthite in Fig. 7.3a. The influx of K-rich fluids from the rhyolitic host into the microsyenodiorite enclaves resulted in the replacement and mantling of sodic plagioclase (in the enclaves) by K-rich alkali feldspar.



microsyenodiorite enclaves have several petrographic characteristics in common with the Kila-Warji parental syenite (Bowden et al. 1987), and microgranular enclaves in the Tibchi ring dyke (Ike et al. 1984).

7.3.2 Enclaves in biotite granite

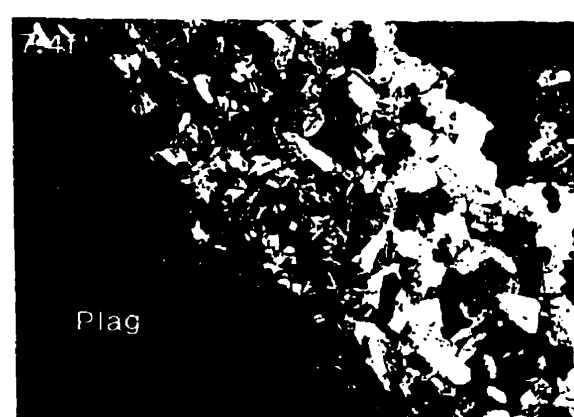
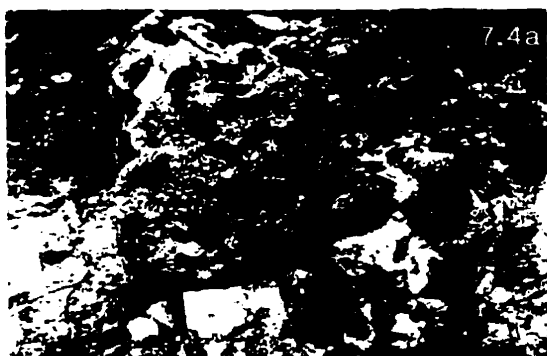
In the plutonic rocks, the mafic enclaves are most commonly encountered along the marginal parts of the Panshanu center (Figs. 7.4a, b), a distribution similar to that described from the adjacent Sara-Fier complex (Turner 1963) to the south (Fig. 1.1).

Inclusions of **trachybasalt** contain plagioclase phenocrysts in the range from 2 x 7 mm to 4 x 4 mm, and constitute 10 to 25% of the enclave population. These phenocrysts are polysynthetically twinned and are unzoned (Fig. 7.4c). The absence of zoning suggests "equilibrium" crystallization prior to mingling. Fine-grained polycrystalline clots of hornblende, biotite, Fe-Ti oxides, and epidote are prominent (Fig. 7.4d), suggesting mingling of a biotite-rich and a hornblende-rich melt, or they may represent replaced pyroxene phenocrysts. Hornblende, biotite, plagioclase, and accessory apatite and Fe-Ti oxides comprise the microcrystalline matrix.

Enclaves of **orb-bearing basalt** are rare, and are hosted by "andesitic" enclaves, forming an example of enclaves within enclaves. The enclaves of orb-bearing basalt contains various types of minerals in the orb-nuclei. These minerals in the orb-nuclei either occur singly or in combination with other minerals (Table 7.1). The plagioclase in most of the orb-nuclei is largely replaced by sericite, epidote, albite, muscovite, fluorite, and biotite. The orb-nuclei are invariably surrounded by a microgranular rim albite + quartz, and groundmass biotite, plagioclase, Fe-Ti oxides, and fluorite (Fig. 7.4e). This assemblage suggests the hybridization of disaggregated clots of basalt with the felsic host.

Plagioclase xenocrysts (5 x 10 cm) are rarely found within trachydacitic enclaves. The

Fig. 7.4: Field photographs and photomicrographs of enclaves in biotite granite. **a)** Field photograph of swarms of enclaves in the Panshanu center (Loc. 73). The enclaves include: andesite (A), trachybasalt (M), anorthosite, and dacite. The fragments of basement (B) may represent disintegrated cauldron-subsided blocks. **b)** Andesitic enclaves (A) containing foliated and rounded fragments of basement (B) (Loc. 73). *Scale: the pen is 14 cm long.* **c)** Photomicrograph of trachybasalt enclave (73B) containing a tabular phenocryst of plagioclase (Plag) in a matrix of hornblende, biotite, Fe-Ti oxides, and plagioclase. *Magnification: x50, X-Nicols.* **d)** Photomicrograph of polycrystalline aggregates of relatively coarse hornblende, plagioclase (plag), biotite, Fe-Ti oxides, and epidote within microgranular matrix of trachybasaltic enclave (73B). *Magnification: x100, X-Nicols.* **e)** Photomicrograph of orb-bearing basalt (89B), with orb nucleus (quartz + Fe-Ti oxides) having a rind of microcrystalline quartz + plagioclase in a biotite-rich matrix. *Magnification: x50, X-Nicols.* **f)** Photomicrograph of plagioclase xenocryst (73B4) with a rim of hornblende, biotite, plagioclase, and Fe-Ti oxides separating the enclave (left) from the felsic host (right). *Magnification: x50, X-Nicols.* **g)** Photomicrograph of an inclusion within plagioclase (Plag) xenocryst (73B4). The inclusions are typically microgranular clots of hornblende, biotite, and Fe-Ti oxides. Fractures in the plagioclase are lined by albite. *Magnification: x50, X-Nicols.* **h)** Photomicrograph of an inclusion mineral bearing clots of hornblende, biotite, and Fe-Ti oxides (cf. Fig. 7.4g). This may be a trapped inclusion of melt that crystallized in situ. *Magnification: x50, X-Nicols.* **i)** Photomicrograph of andesitic enclave (73B2) with laths of zoned prismatic plagioclase in a matrix of quartz, albite, and alkali feldspar. The sericitized cores of plagioclase have sodic overgrowths. *Magnification: x50, X-Nicols.* **j)** Photomicrograph of dacitic enclave (73C) with phenocrysts of zoned plagioclase (Plag) in a matrix of biotite, plagioclase, quartz, and Fe-Ti oxides. *Magnification: x50, X-Nicols.*





xenocrysts have an amphibole-rich rind (Fig. 7.4f) that becomes progressively biotite-rich toward the granitic host. The plagioclase most likely formed as a product of crystallization of a more mafic melt, and was subsequently mixed in with a granitic host (Figs. 7.4f - h). Some of the inclusions in the plagioclase xenocrysts contain multigranular clots of hornblende, biotite and Fe-Ti oxides (Figs. 7.4g, h). Such inclusions may be trapped inclusions of melt that crystallized in situ.

The **andesitic enclaves** have a sharp contact with the felsic host (Appendix VII.1f). Plagioclase phenocrysts range from acicular to stubby prisms, and invariably have a mantle of clear oligoclase (Fig. 7.4i, Appendix VII.1f). Some grains develop a rind of amphibole partly replaced by biotite. This suggests disequilibrium after mixing. The finer-grained plagioclase in the matrix is polysynthetically twinned albite to oligoclase.

Dacitic enclaves are the most abundant type and appear to have incorporated all other types of enclaves in biotite granite, which leads to double enclaves on the scale of a hand specimen. Most of the plagioclase phenocrysts have a relatively clear core, an inclusion-rich outer part, and a mantle of clear albite (Fig. 7.4j). Some of the other plagioclase phenocrysts are zoned, with a dusty brown sericitized core and a clear rim. The microphenocrysts of amphibole bear inclusions of Fe-Ti oxides and biotite along the margin.

Synopsis of petrographic features and inferences

The rounded, ellipsoidal, and elongated forms of the enclaves suggest liquid-liquid contacts between mafic magma and the host magma. The necking of some of the enclaves and the occurrence of felsic inclusions within the mafic enclaves indicate arrested stages of blending during the commingling of the mafic magma and the felsic host-magma. The occurrence of rounded and more mafic polycrystalline inclusions (comprised of hornblende, biotite, Fe-Ti

oxides, and plagioclase, particularly prominent in enclaves within biotite granite), suggests mixing between basaltic magma and a melt less evolved than rhyolite, before mingling of the hybrid with the melt that formed biotite granite. Xenocrysts of hedenbergite, alkali feldspar, and quartz (in trachyandesite and other enclaves), inferred to be derived from the felsic host rock, suggest recycling of liquidus phases between 1) the melt that formed enclaves and 2) the felsic (host rock) magma.

Zoning in plagioclase may reflect the cyclical pulses of mixing that occurred before commingling between the hybrid magma and the felsic rocks. The orbicular texture observed in orb-bearing basalt suggests hybridization with intake of xenocrysts from the felsic host rock or wall rock to form orb nuclei. The presence of a rim on xenocrysts and phenocrysts of ferromagnesian minerals suggests disequilibrium. On the other hand, sieve-textured plagioclase (and clinopyroxene) in the enclaves could have formed as a result of rapid decompression (Nelson & Montana 1992) and change in composition upon mixing, with attendant partial dissolution due to temperature increase (Nixon & Pearce 1987).

To summarize, the textures show that: i) In the subvolcanic rocks, both the enclaves and their felsic host rocks were molten at the time of commingling, ii) mixing was effected to various stages before the hybrid magmas were mingled with their felsic host, and iii) the enclaves were either disrupted and hybridized into their felsic host rock, and in some cases were undercooled, with minimum exchange with the host rock.

7.4 Mineral Chemistry

Chemical data on the plagioclase, pyroxene, amphibole, and mica were collected by wavelength-dispersion spectrometry (WDS) using a Cameca CAMEBAX electron microprobe. An accelerating voltage of 15 kV, beam current of 20 nA, and spot size between 4 and 8 μm were

used. Counting times were 25 seconds for all elements, except for Cl (50 seconds) and F (100 seconds). The series of synthetic and natural standards used include: andradite (Fe), MnTi oxide (Mn, Ti), albite (Na), diopside (Si, Mg, Ca), orthoclase (K, Al), fluorite (F), and vanadinite (Cl). The raw data were ZAF-corrected and reduced using the on-line ONQANT software package. Results of analyses of minerals are presented in Appendix VII.2 - 5.

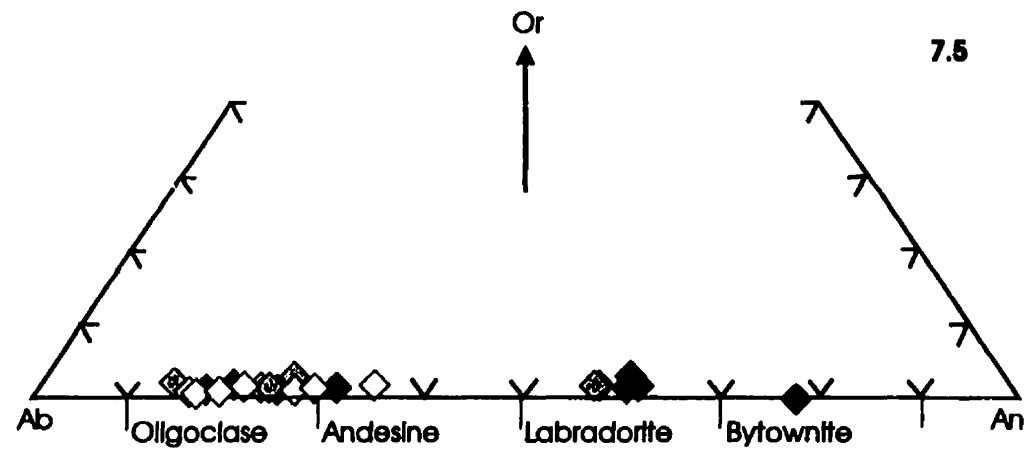
7.4.1 Plagioclase

The microprobe data for plagioclase from enclaves in biotite granite are presented in Table 7.2 and Figure 7.5, whereas the results detailed (spot) analyses are given in Appendix VII.2.

Polymodal plagioclase compositions are documented in the trachybasalt, andesite, and dacitic enclaves. The trachybasalt enclave chosen for detailed study has a mixed population of plagioclase, which includes labradorite ($An_{59.9}$) phenocrysts and oligoclase ($An_{17.3}$) xenocrysts. Within the trachybasalt enclaves, the phenocrysts of labradorite are zoned, but display only minor variations, and therefore likely crystallized within a basaltic host. The phenocrysts of coexisting oligoclase are zoned with slight variations from the core (An_{24}) to the margin ($An_{19.8}$), and most of such oligoclase grains have a mantle of andesine ($An_{30.1}$). The composition of the mantle suggests that the oligoclase is xenocrystic, having initially crystallized in a more evolved melt, presumably corresponding to the granitic host in composition. The oligoclase later was entrained into the trachybasaltic melt, at which point it was overgrown by andesine.

Also, in the andesitic enclaves, the plagioclase xenocrysts are predominantly labradoritic ($An_{56.3-57.1}$), with subsidiary discrete grains of oligoclase ($An_{22.6-13.3}$). Andesine, not labradorite, is expected in "real" andesite; the presence of labradorite confirms that the andesitic compositions are due to mixing. On the other hand, the zoned crystals of plagioclase in the dacitic enclaves vary from andesine (An_{25-34}) to oligoclase (An_{16-28}).

Fig. 7.5: Plagioclase compositions in the enclaves in biotite granite. The bimodality of plagioclase compositions in trachybasalt, andesite, and dacite suggests hybridization of partly crystallized melts of contrasting compositions. The plagioclase xenocrysts in dacitic enclaves have a single composition and might be mantle-derived xenocrysts brought to the surface in alkali basalt. The average composition of these xenocrysts ($An_{77.4}$) is significantly more An-rich than in the anorthosite inclusion from the Rafin-Jaki granite porphyry in the Jos-Bukuru complex (An_{58-65} , Wright 1975), the andesine (An_{50}) in the "parental" Kila-Warji augite syenite (Bowden 1982, 1985 Bowden et al. 1987), and the anorthosite (An_{40-60}) from Aïr in Niger Republic (Perez 1985, Bowden et al. 1987, Moreau et al. 1987).



Enclaves in biotite granite

- ◇ Trachydacite
- ▣ Andesite
- ◆ Trachybasalt
- ◆ Anorthosite

The plagioclase in the dacitic enclaves also is invariably zoned: i) andesine is normally zoned (average core: An_{34.1}, and oligoclase rim: An_{25.8}), with most grains displaying oscillatory zoning; and ii) grains of oligoclase also are normally zoned (core: An₂₈, rim: An_{16.8}). The different plagioclase populations and zoning documented above are similar to what has been described at other localities where mixing occurred (Highwood Mountains, Montana: O'Brien et al. 1988, Iztaccíhuatl volcano, Mexico: Nixon 1988a, b). In the latter suite, Nixon & Pearce (1987) documented the presence of polymodal plagioclase and oscillatory zoning, and inferred recycling of phenocrysts due to the blending (and repeated mixing) of mafic (basaltic) and intermediate (dacite) or hybrid liquids of contrasting compositions.

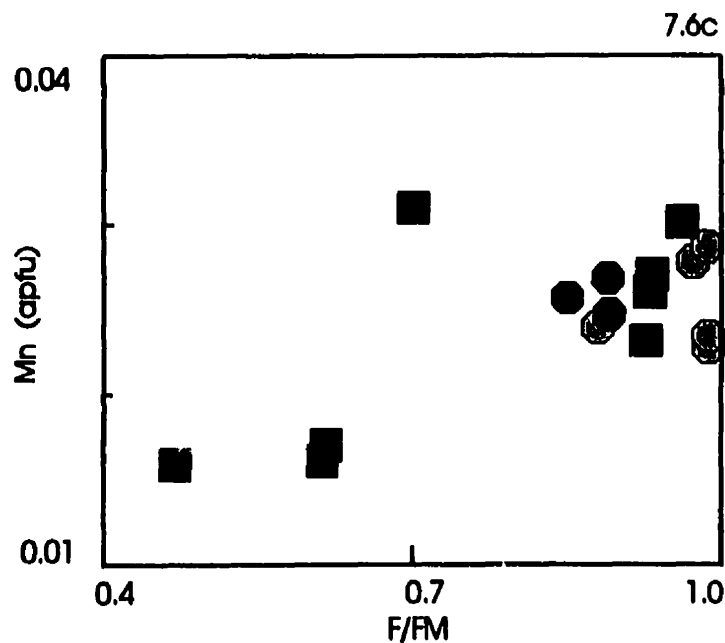
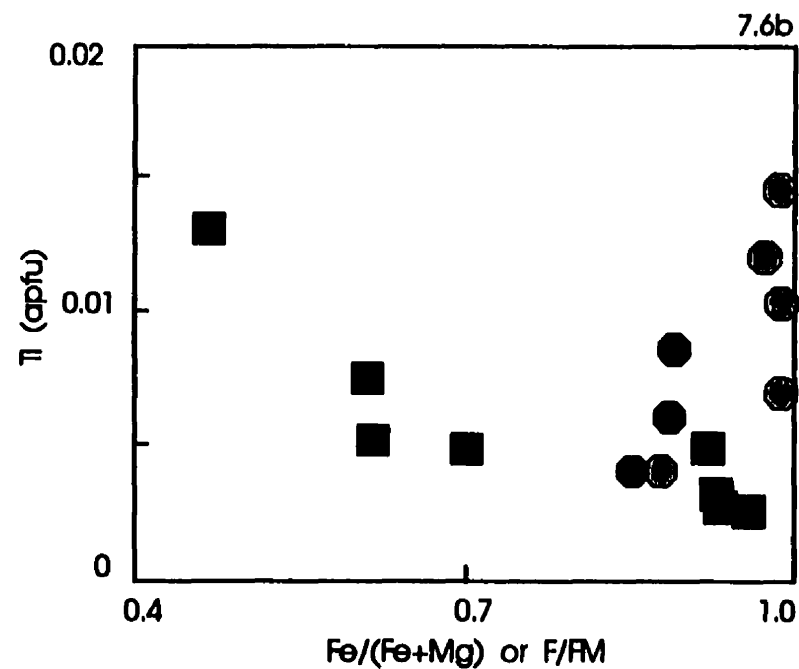
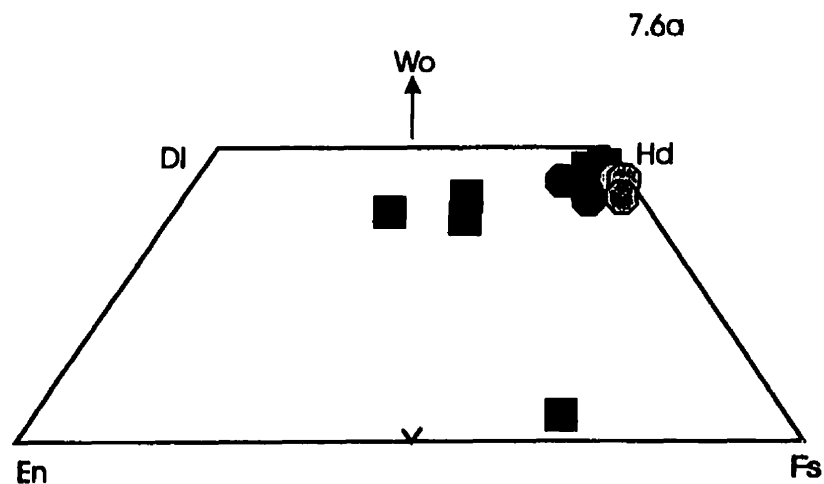
Semiquantitative SEM-EDS analysis shows that the xenocrystic plagioclase (in a dacitic enclave) is bytownite (An₇₇). Such a composition is significantly more An-rich than the labradorite phenocrysts in trachybasalt enclaves described above, and anorthosite inclusions (An₅₈₋₆₅) within the Rafin-Jaki granite porphyry ring dyke of the Jos-Bukuru complex (Wright 1975). This plagioclase composition also is more calcic than the andesine in the "parental" Kila-Warji augite syenite (Bowden 1985, Bowden et al. 1987), as well as the andesine - labradorite range of compositions (An₄₀₋₆₆) found in anorthosite from the Tagueï ring complex, Aïr, Niger Republic (Perez 1985, Bowden et al. 1987, Moreau et al. 1987). Such plagioclase is likely a mantle-derived xenocryst, and shows that relatively primitive basic magma may have attained the shallow crust quickly during the final emplacement of the batches of felsic magma.

7.4.2 Pyroxene

Pyroxene is limited in occurrence to enclaves in subvolcanic rocks of the Buri ring dyke. Average compositions of the ortho- and clinopyroxene in enclaves are presented in Table 7.3, whereas the detailed data are listed in Appendix VII.3.

Fig. 7.6: Compositional variation of pyroxenes in enclaves and their felsic host-rocks. a) Pyroxene quadrilateral (En-Di-Hd-Fs). Whereas the trachyandesite enclaves have a slightly less evolved composition of hedenbergite than the felsic rock-units in the Buri ring dyke, the hedenbergite's composition in the microsyenodiorite is significantly less evolved. b) Plot of Ti vs. $\text{Fe}/(\text{Fe}+\text{Mg})$ or F/FM. Despite the differences in the F/FM, the hedenbergite (in the trachyandesite and microsyenodiorite enclaves, and their felsic host) has low Ti contents, lower than what is found in ferroaugite. c) Mn vs. F/FM. The plot shows the hedenbergite in the microsyenodiorite enclave to have lower Mn contents than the hedenbergite in felsic units, although the clinopyroxene in trachyandesite enclaves have similar Mn content with those in the host rock (rhyolite porphyry).

From these diagrams, it is inferred that: i) the clinopyroxene in trachyandesite enclaves, on the basis of their F/FM values, and Ti and Mn contents, is either xenocrystic, or it (as well as the enclaves) experienced equilibration, upon mingling, with their rhyolitic host-rocks, and ii) the ortho- and clinopyroxene in microsyenodiorite enclaves were least affected by commingling with the felsic host.



SYMBOLS USED IN FIGS. 7.6 - 8

Subvolcanic rocks

- Hedenbergite fayalite granite porphyry
- Rhyolite porphyry

Enclaves in subvolcanic rocks

- Microsyenodiorite
- Trachyandesite

Blotite granites (plutonic)

- + Albitized blotite granite
- × Pristine blotite granite

Enclaves in blotite granite

- ◇ Trachydacite
- ◇ Andesite
- ◆ Trachybasalt
- ◆ Anorthosite

Trachyandesitic enclaves in rhyolite porphyry are characterized by Fe-rich hedenbergite ($\text{Fs}_{50.1}\text{Wo}_{43.5}\text{En}_{6.4}$, $\text{F/FM} = 0.89$). In some instances, the hedenbergite grains are reversely zoned (core $\rightarrow$ rim: $\text{Fs}_{51.6\rightarrow47}$, $\text{Wo}_{41.7\rightarrow44.4}$, and $\text{En}_{6.7\rightarrow8.6}$) (Fig. 7.6a), with F/FM going from 0.89 in the core to 0.85 in the rim, consistent with the presence of a mafic contaminant. However, the hedenbergite in the trachyandesite enclaves is only slightly more magnesian than that in the host rhyolite porphyry (Fig. 7.6a-c). This suggests either that the hedenbergite in the enclave is xenocrystic or that the enclaves were nearly equilibrated with the felsic host rocks (cf. Allen 1991).

In the microsyenodiorite enclaves within the hedenbergite fayalite granite porphyry, both ortho- and clinopyroxene are present (Figs. 7.6a-c). The microphenocrysts of ferrosilite (orthopyroxene) have an average composition of $\text{Fs}_{67.2}\text{Wo}_{3.4}\text{En}_{29.4}$, $\text{F/FM} = 0.62$. This is the first known report of orthopyroxene in a suite of rocks in the NYG. The orthopyroxene coexists with hedenbergite ($\text{Fs}_{36.2}\text{Wo}_{40.6}\text{En}_{23.2}$, $\text{F/FM} = 0.62$), which is more Mg-rich than the hedenbergite in trachyandesite enclaves (Fig. 7.6a). Some (not all) of the coexisting grains of clinopyroxene in the microsyenodiorite enclaves are zoned, with a ferroaugite core ($\text{Fs}_{27.5}\text{Wo}_{39.2}\text{En}_{33.3}$, $\text{F/FM} = 0.47$) and a hedenbergite rim ($\text{Fs}_{36.7}\text{Wo}_{39.6}\text{En}_{23.7}$, $\text{F/FM} = 0.61$). Zoning in this case is normal, i.e., it involves an increase in F/FM , and a decrease in Ti (Fig. 7.6b). Such zoning most likely results from normal fractionation. The hedenbergite in the granite porphyry host of the microsyenodiorite enclave in the Buri ring dyke is much more enriched in Fe ($\text{Fs}_{50.2}$) and Mn, and depleted in Ti (Figs. 7.6a-c). Such pronounced compositional contrast in this example suggests limited exchange between the rhyolitic (granite porphyry) and hybrid (microsyenodiorite) melt, due to quenching of the latter against the former during commingling.

7.4.3 Amphibole

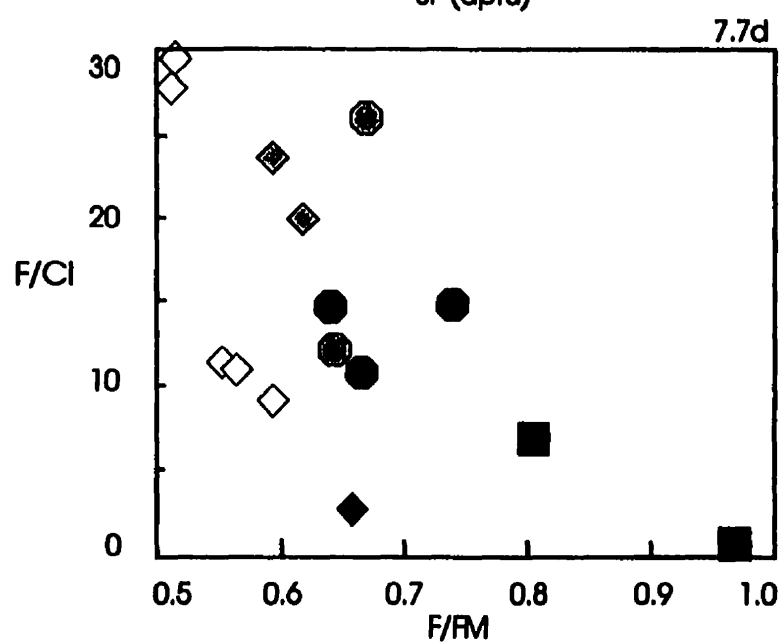
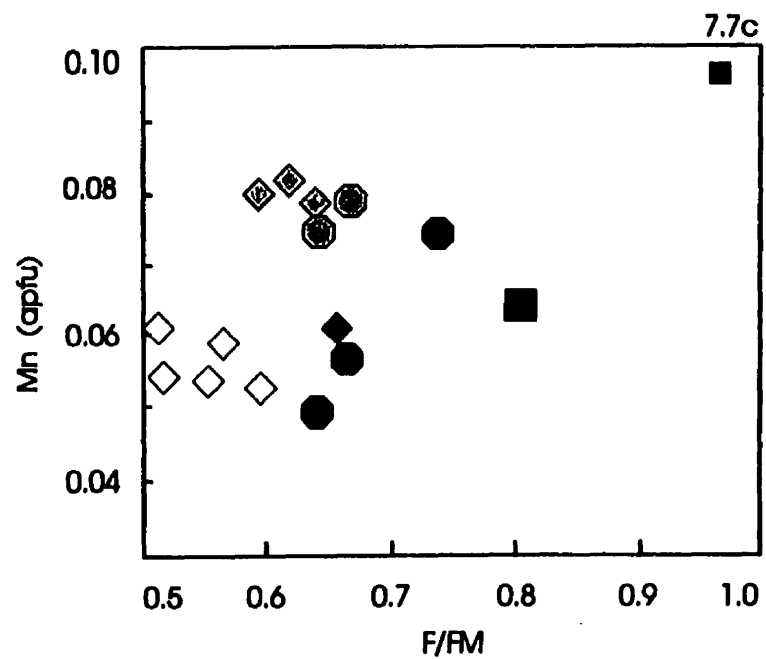
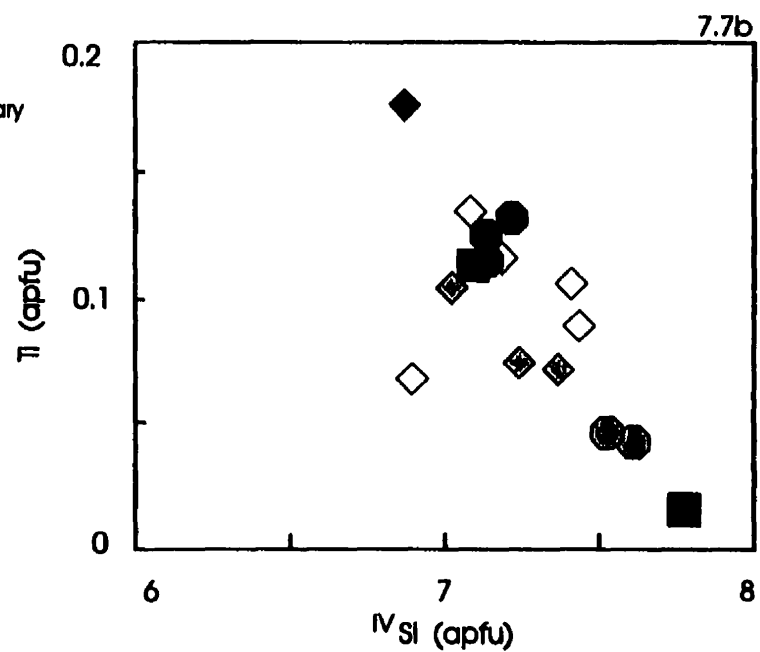
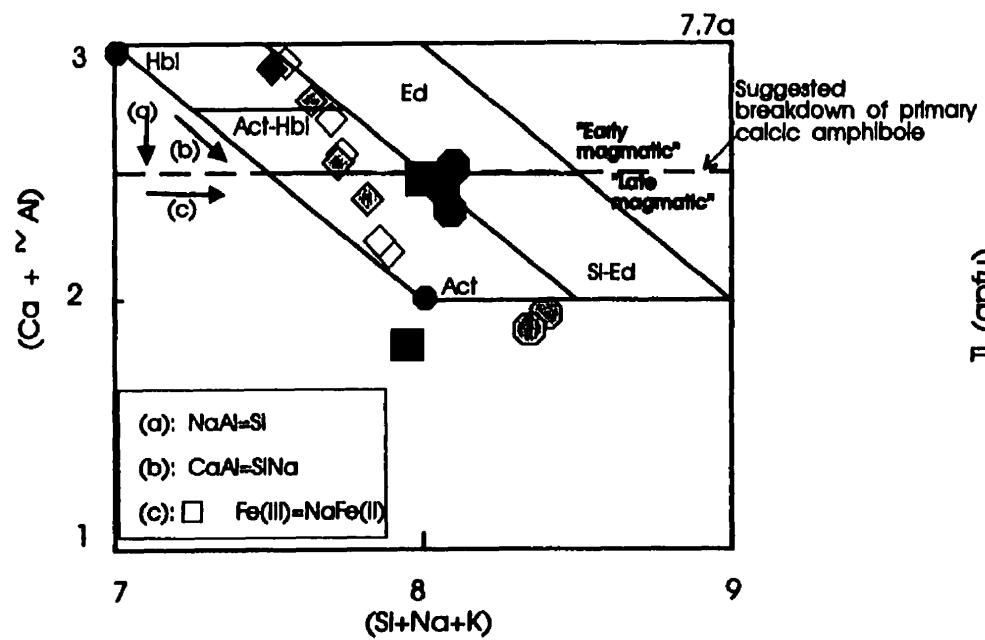
The averaged compositions of amphibole in enclaves and their host rocks are presented in Table 7.4, whereas results of the detailed (spot) analyses are given in Appendix VII.4. The amphibole in trachyandesite and microsyenodiorite enclaves is secondary, replacing clinopyroxene. In the enclaves in biotite granite, the amphiboles seem to be primary, but the amphibole in polycrystalline clots within such enclaves is interpreted as secondary. The compositional variations are displayed in Figures 7.7a - d.

Trachyandesitic enclaves are characterized by ferroedenite ($0.64 \leq F/FM \leq 0.74$). The ferroedenitic composition contrasts with the secondary amphibole in the host rhyolite porphyry, ferroactinolite (Figs. 7.7a - d). Within **microsyenodiorite** enclaves, the limited amphibole analyses obtained indicate a ferrohornblende composition with $F/FM = 0.80$. Typically, the ferrohornblende in the enclave has a higher Ti, ^{IV}Al , and lower Mn than secondary ferroactinolite (after clinopyroxene) in the host granite porphyry (Fig. 7.7a-d). The difference in the composition of calcic amphibole shows that the enclave and the granite porphyry were not in equilibrium.

The **trachybasalt** enclaves in biotite granite contain a homogeneous ferroedenitic amphibole ($F/FM = 0.66$). The amphibole in this type of enclave has higher Ti, ^{IV}Al , and lower Mn than amphibole in other enclaves within biotite granite (Figs. 7.7b-d), which is consistent with the less evolved composition of these enclaves. The "andesitic" enclaves, on the other hand, have polymodal compositions of amphibole (Fig. 7.7a): ferroedenite ($F/FM = 0.64$), ferroactinolitic hornblende ($F/FM = 0.59$), and ferrohornblende ($F/FM = 0.62$); all are present in separate grains. The provenance of each amphibole cannot be discerned, but it is apparent from the F/FM values that these amphiboles had more or less equilibrated with the same hybrid melt. Like the andesitic enclaves, dacitic enclaves also contains amphibole grains with variable composition: ferroedenite ($F/FM = 0.60$), ferrohornblende ($F/FM = 0.56$), and ferroactinolitic hornblende ($0.51 \leq F/FM \leq$

Fig. 7.7: Plots of amphibole composition in enclaves and their host rocks; a) Plot of $(Ca + ^{IV}Al)$ vs. $(Si+Na+K)$. The compositions of amphiboles in subvolcanic enclaves plot close to or below the line of breakdown of magmatic calcic amphiboles, i.e., in the "late magmatic" field of Giret et al. (1980). The amphibole compositions in the enclaves in biotite granite plot in both the "early magmatic" and "late magmatic" fields, although they are considered primary. b) Ti vs. ^{IV}Si (apfu). The Ti content of amphibole in enclaves decreases from trachybasalt through enclaves in subvolcanic rocks and dacitic and andesite enclaves in plutonic rocks to felsic subvolcanic rocks of the Buri ring dyke. c) Mn vs. F/FM. No systematic variations can be discerned between the Mn content and F/FM of amphiboles in the enclaves. However, amphiboles in the felsic granite porphyry has the highest F/FM and Mn. d) F/Cl vs. F/FM. Trachybasalt has the lowest F/Cl, whereas trachydacite has a bimodal distribution of F/Cl of its amphibole. The F/FM and F/Cl values of amphibole in trachyandesite are similar to those of their counterparts in the felsic rhyolite porphyry host (49B); the amphibole in microsyenodiorite enclave has a lower F/FM and higher F/Cl than the granite porphyry (67).

From the above plots, one can infer that the effect of contamination or metasomatic modification of hybrid (enclave-forming) magma cannot be discerned, principally because of absence of primary amphibole in the felsic host rocks of the enclaves (biotite granites).



0.55). The polymodal compositions of amphibole documented in andesitic and dacitic enclaves are typical of hybrid rocks (Nakada 1991).

7.4.4 Mica

The average microprobe data for micas in enclaves within biotite granite are presented in Table 7.5 and Figure 7.8, which includes reference data for selected textural variants of biotite granite (Chapter 4). The results of detailed (spot) analyses of mica are given in Appendix VII.5a.

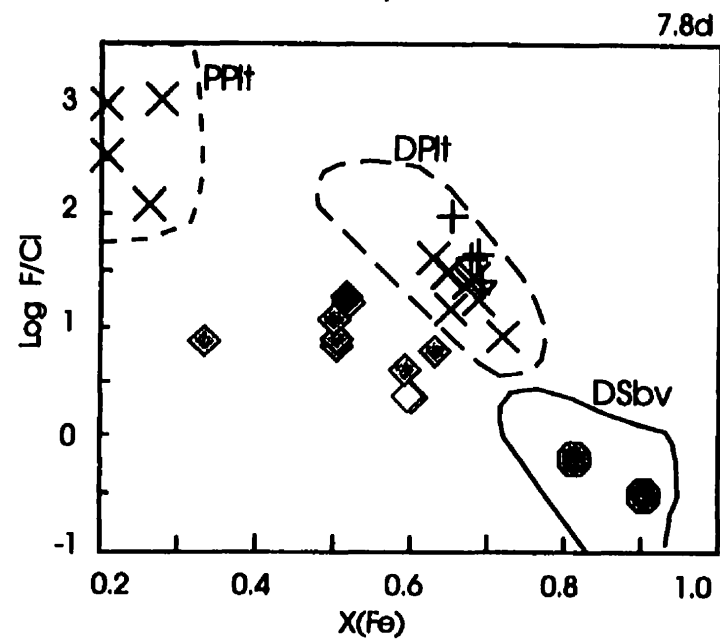
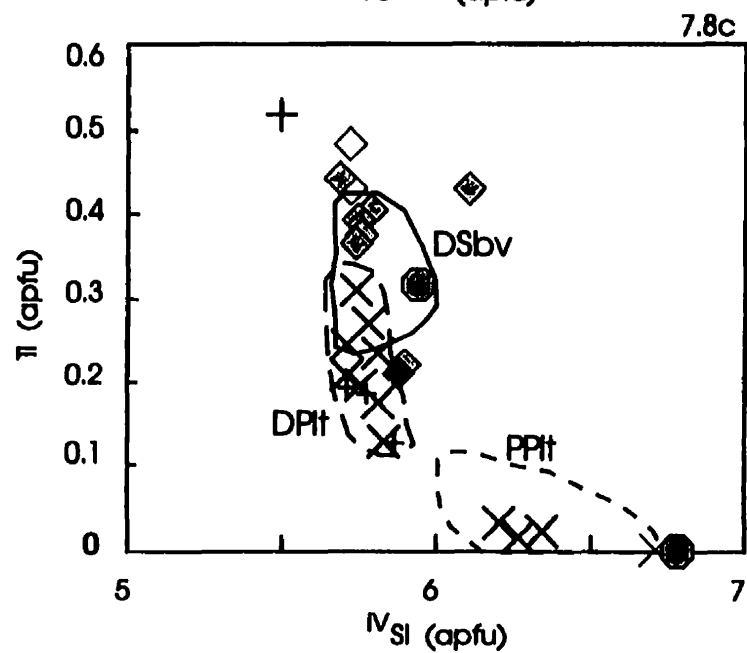
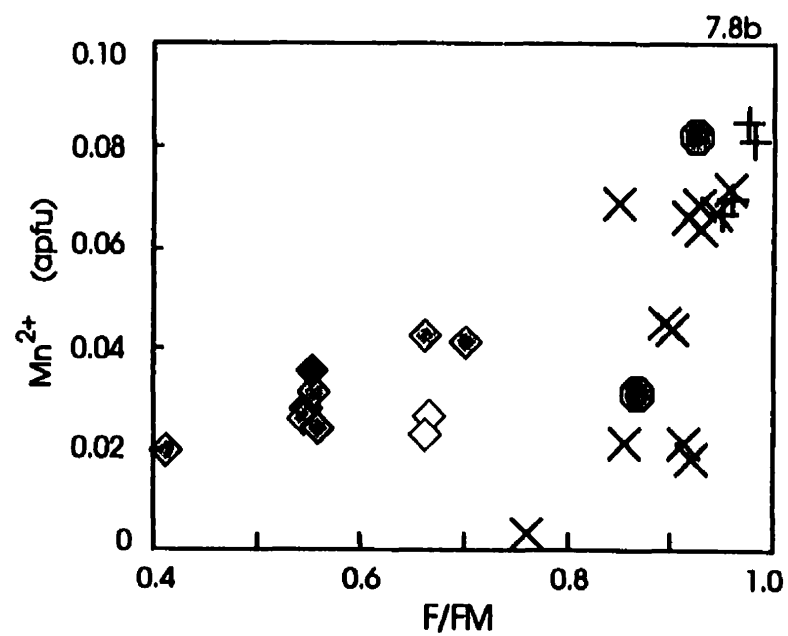
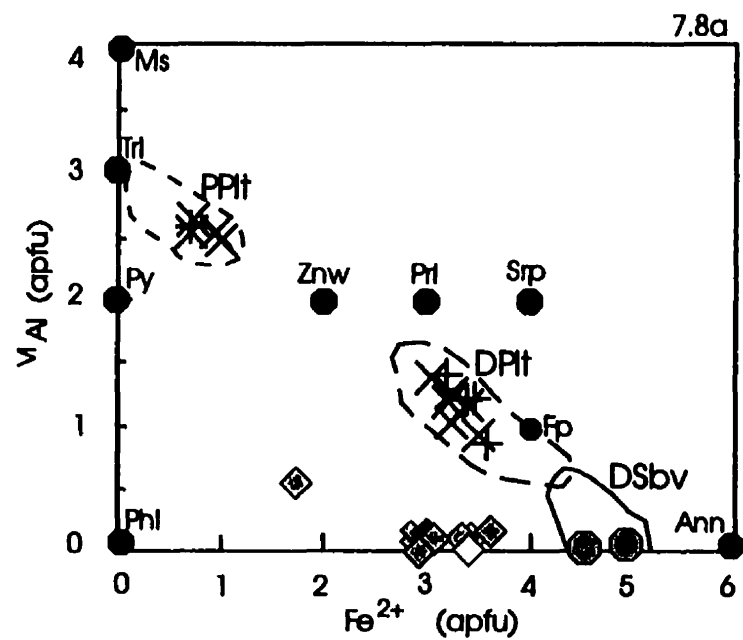
The Fe content of the mica increases in the sequence "andesitic" enclave: ($\text{Ann}_{40-42}\text{Phl}_{50.7-51.3}$), through "trachybasalt" ($\text{Ann}_{50.2}\text{Phl}_{41.8}$) to "dacite" enclaves ($\text{Ann}_{62.4-67.7}\text{Phl}_{30.1-30.5}$). Felsic microgranular enclaves have mica compositions that are uniform and similar to those in dacite ($\text{Ann}_{62.5}\text{Phl}_{28.8}$) (Figs. 7.8a - d). From petrographic studies, the mica is inferred to be secondary, and it seems that a K-bearing fluid in equilibrium with the granitic host most likely replaced amphibole in the enclaves to produce mica. Whereas the F/FM of the mica in andesitic enclaves is similar to that of the mica in the trachybasalt, the F/Cl values of mica in these two types of enclaves differ, the former being more F-rich, and thus possibly more strongly affected by the fluid medium. The dacitic and felsic enclaves have a higher F/FM and lower F/Cl ratios (Fig. 7.8c, d), which suggests a relatively greater extent of interaction with fluids.

Synopsis of mineral chemical variations and inferences

Whereas the similarity in composition between clinopyroxene in trachyandesite to that in the host rhyolite porphyry suggests either phenocryst recycling or the equilibration of phenocrysts in both magmas, the pyroxenes in the microsyenodiorite enclaves are compositionally less evolved than that in their host rocks. Thus in the granite porphyry, the enclave minerals did not equilibrate with the host magmas, and likely were quenched before reaction could take place. This may

Fig. 7.8. Compositional variations of mica in enclaves and biotite granite; **a)** Plot of ^{VI}Al vs. Fe^{2+} . The enclaves in the biotite granite illustrate a different trend of mica compositions (Mg-rich) than the mica in their felsic host. **b)** Mn^{2+} vs. F/FM . Despite the fact that the enclaves have lower F/FM than their felsic host, the biotite in these enclaves has similar Mn contents to the annitic biotite in plutonic and some subvolcanic rocks. **c)** Ti vs. Si. The Ti content of the mica in enclaves is similar to that of the mica in subvolcanic rocks (Chapter 4), and generally higher than that in the felsic plutonic rocks. **d)** $\log (F/Cl)$ vs. X_{Fe} . The $\log F/Cl$ ratio of the mica in enclaves is only slightly lower than that in the biotite granites. This finding suggests incomplete equilibration of the enclaves and their host rocks with respect to the halogens.

Symbols: Ann = annite, Phl = phlogopite, Fp = ferroprotolithionite, Srp = siderophyllite, Prl = protolithionite, Znw = zinnwaldite, Py = polyolithionite, Trl = trilithionite, Ms = muscovite, DSbv = dark mica in subvolcanic rocks, DPlt = dark mica in plutonic rocks, PPlt = pale mica in plutonic rocks.



explain the prominence of subsolidus replacement in the microsyenodiorite enclaves. In the enclaves in the biotite granite, the polymodal composition of plagioclase and mica suggests the mixing of two magmas less evolved than the biotite granite magma, with the hybrid magma later forming enclaves in their present host-rocks. Such a model is further supported by normally zoned oligoclase mantled by andesine. Mica compositions vary within a limited range in each enclave, which likely indicates different extents of late-magmatic equilibration of enclaves with their felsic host rocks.

The pyroxene and plagioclase compositions suggest that the enclaves evolved by hybridization before mingling with their present host rocks, and they subsequently underwent various extents of postmagmatic modifications. The mica chemistry appears to be more useful as an indicator of late-magmatic to subsolidus modification of enclaves than the composition of other minerals because of its mode of formation and occurrence in both the enclaves and their host rocks.

7.5 Geochemical characteristics

Concentrations of major and trace elements were determined using XRF spectrometry on fused beads prepared from ignited samples and on pressed powder pellets, respectively. Analyses for the REE were done by inductively coupled plasma - mass spectrometry (ICP-MS), using the sodium peroxide sintering technique, which involves: i) sintering of an aliquot of 0.2 g sample with sodium peroxide so as to ensure complete digestion of the sample, including particularly resistant REE-bearing accessory phases, ii) dissolution of the sinter cake followed by the separation and dissolution of the REE-hydroxide-bearing precipitate, and iii) analysis by ICP-MS (S. J. Jackson, pers. comm. 1993). Certified geological reference (USGS) standards were analyzed along with these samples to monitor precision.

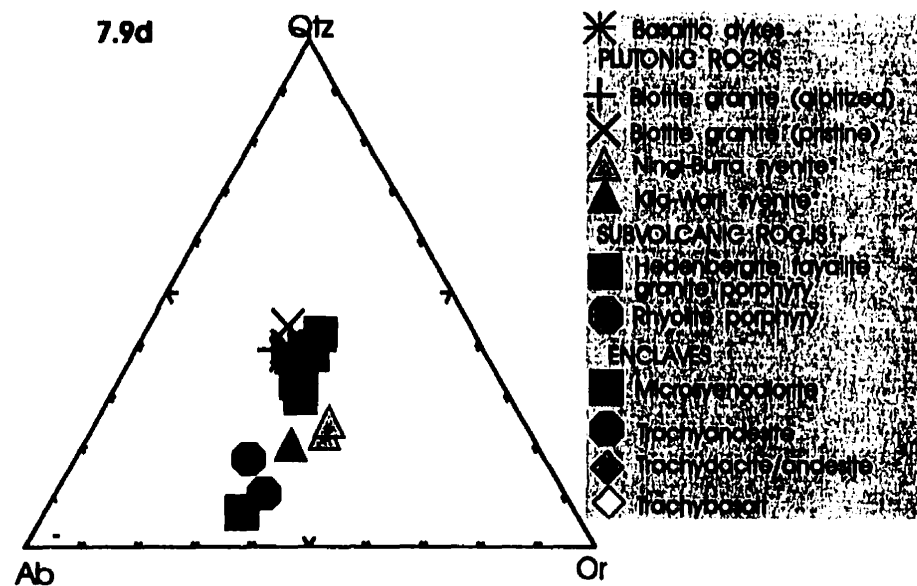
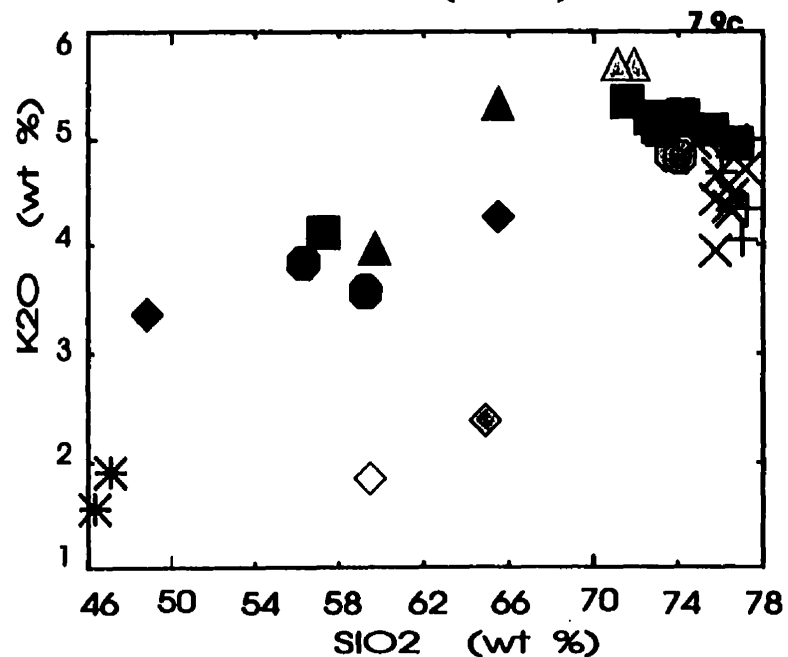
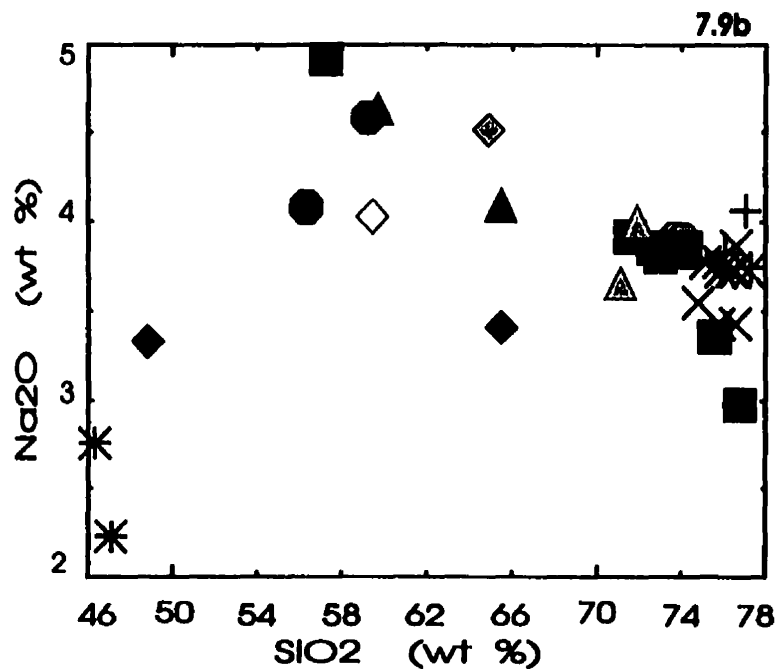
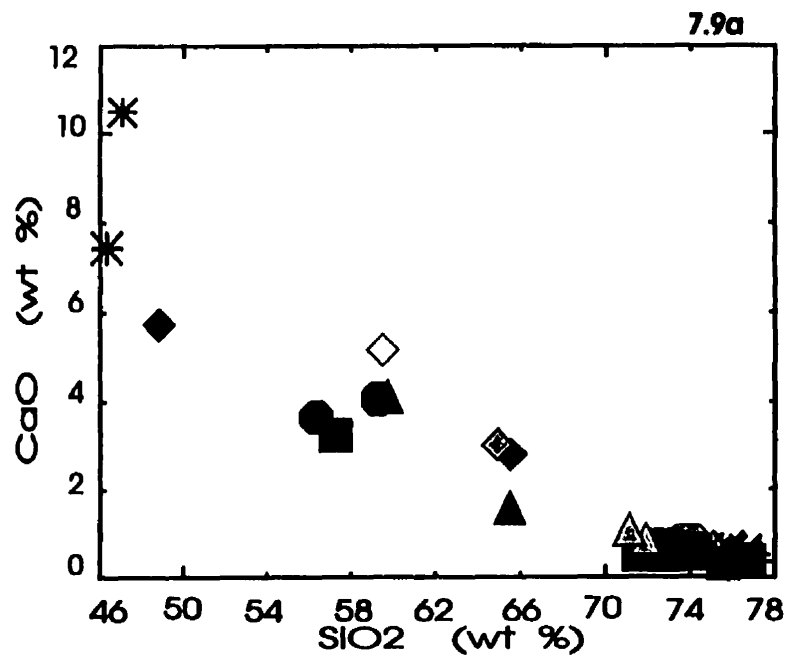
The chemical composition of enclaves and their host rocks, and calculated normative compositions, are given in Table 7.6. Data for selected samples of syenite from the Kila-Warji and Ningi-Burra complexes are included for comparison. An emphasis is placed on the relationship between the enclaves and their felsic host-rocks.

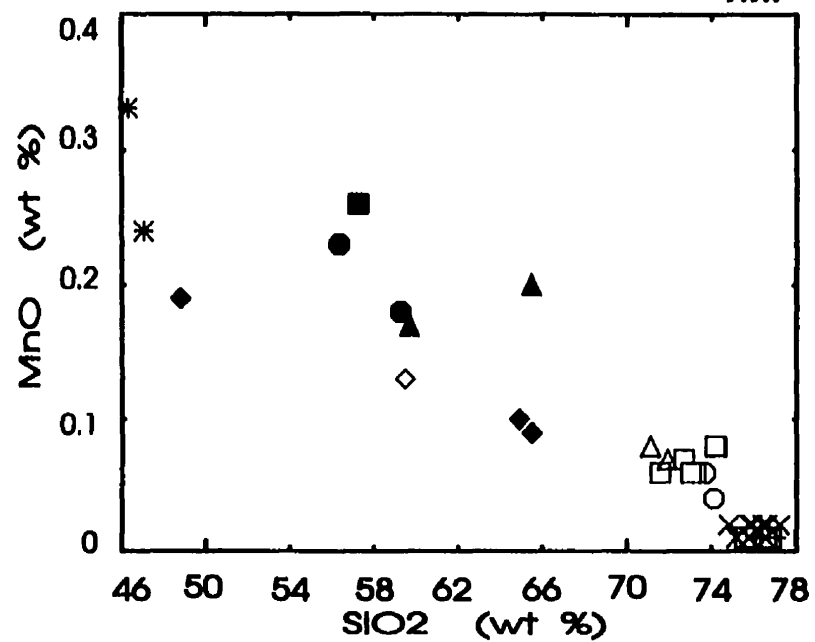
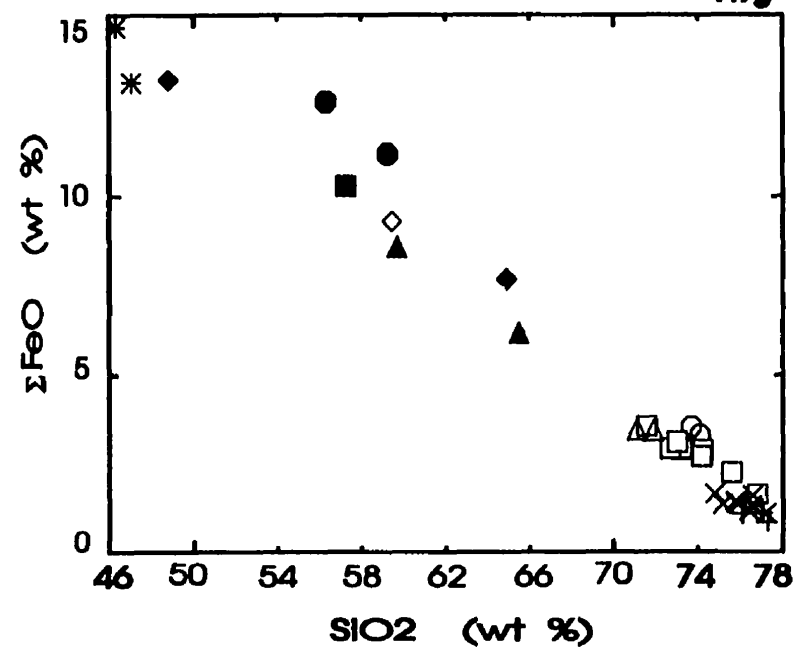
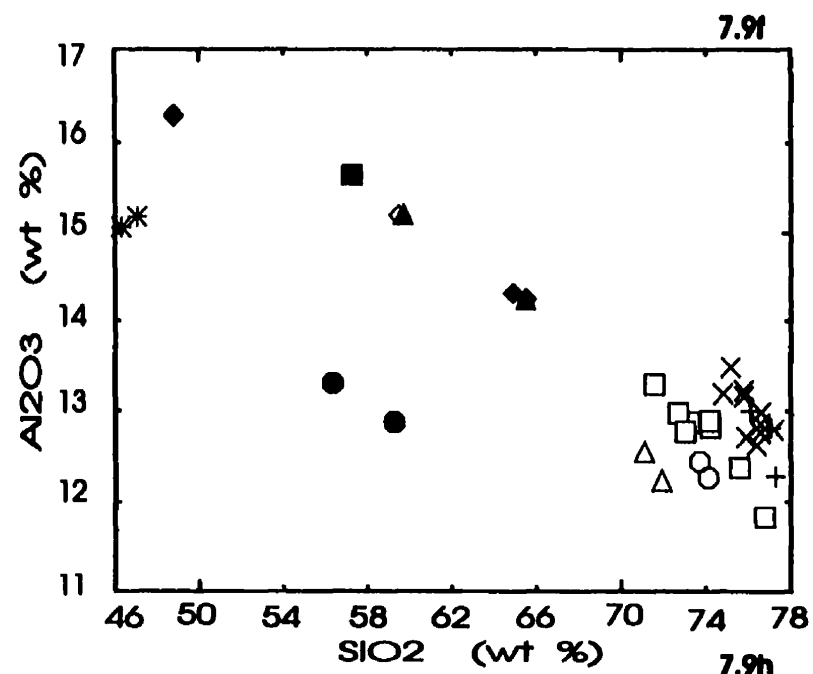
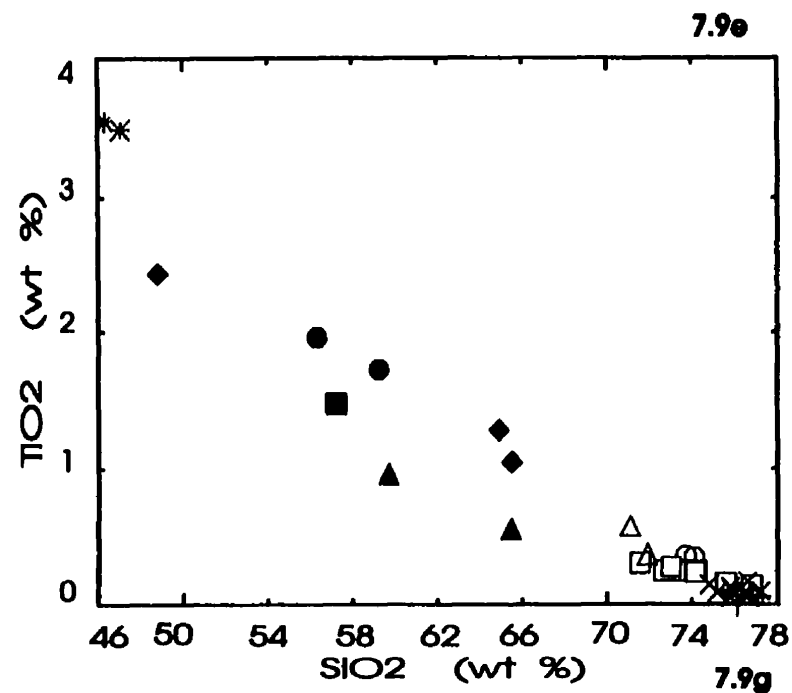
7.5.1 Major elements

The major-element data for the enclaves are plotted in Figures 7.9a - h. There is a minor compositional discontinuity, in terms of SiO_2 , between basaltic dykes and the enclaves, and another between enclaves (and syenites from neighbouring complexes) and the host subvolcanic rocks. In Figure 7.9a, there is a linear decrease in the Ca content from basaltic dykes, via enclaves and syenites, to subvolcanic rocks and biotite granite. The Na content (Fig. 7.9b) increases from typical levels in the basalt to enclaves, and then decreases to the felsic subvolcanic and plutonic rocks. This trend is totally different from what typically is encountered in fractionation-dominated sequences, in which Na build-up is expected at initial stages of crystallization of mafic magma, with a flattened trend toward the end stages. The variation in Na content, unlike that of Ca, cannot be adequately explained by mixing (Fig. 7.9b). In Figure 7.9c, the K_2O content, in a pattern similar to Na_2O , increases from basalts through the enclaves to syenites, and then decreases from the syenite through the subvolcanic to plutonic rocks. Such a pattern of variation, which departs from either fractionation or mixing trends, may be due to diffusive exchange between commingled magmas and later metasomatic modification of the enclaves. From petrographic observations, the entrained orthoclase microperthite xenocrysts in trachyandesite enclaves (Fig. 7.2a), the mantling of sodic plagioclase by K-rich microperthite (microsyenodiorite: Figs. 7.3a, c, d), and the replacement of clinopyroxene and amphibole in most enclaves by mica all provide indications of K replacing Na, i.e., by ion exchange. It is therefore inferred that, to

Fig. 7.9: Plots of major element variations in enclaves and their host rocks: **a)** CaO, **b)** Na₂O, and **c)** K₂O versus SiO₂; and **d)** triangular plot of normative albite - orthoclase - quartz, **e)** TiO₂, **f)** Al₂O₃, **g)** ΣFeO, and **h)** MnO versus SiO₂.

(Symbols: **filled diamonds** = trachybasalt enclaves, **open diamonds** = andesitic enclaves, **filled circles** = trachyandesite enclaves, **filled squares** = microsyenodiorite enclaves, ***** = basaltic dykes, **open circles** = rhyolite porphyry, **open boxes** = granite porphyry, **x** = pristine plutonic rocks, **+** = altered plutonic rocks, **closed triangles** = Kila-Warji syenite, **open-triangles** = Ningi-Burra syenite).





a first approximation, the composition of the enclaves can be explained by hybridization followed by disturbances in the alkalis, which are expected to be rather mobile.

In Figure 7.9d, a plot of normative albite - quartz - orthoclase, most of the enclaves plot in the field of NYG syenites, which have been suggested as parental to the rhyolitic subvolcanic rocks. Some of the enclaves in biotite granites (trachybasalt and andesite) do not lie close to such a trachyte-rhyolite fractionation trend (further examined in chapter 8). Plots of TiO_2 , Al_2O_3 , ΣFeO , and MnO against SiO_2 (Figs. 7.9e to h, respectively) show decreasing contents of these constituents from basalts via the enclaves to subvolcanic and plutonic rocks. The linear array could be explained by mixing of basalt and trachyte - rhyolite to produce the enclaves. However, the Al_2O_3 content of some of the enclaves lies above a single mixing line, proving that one or more of the simplifying assumptions is violated.

7.5.2 Trace elements

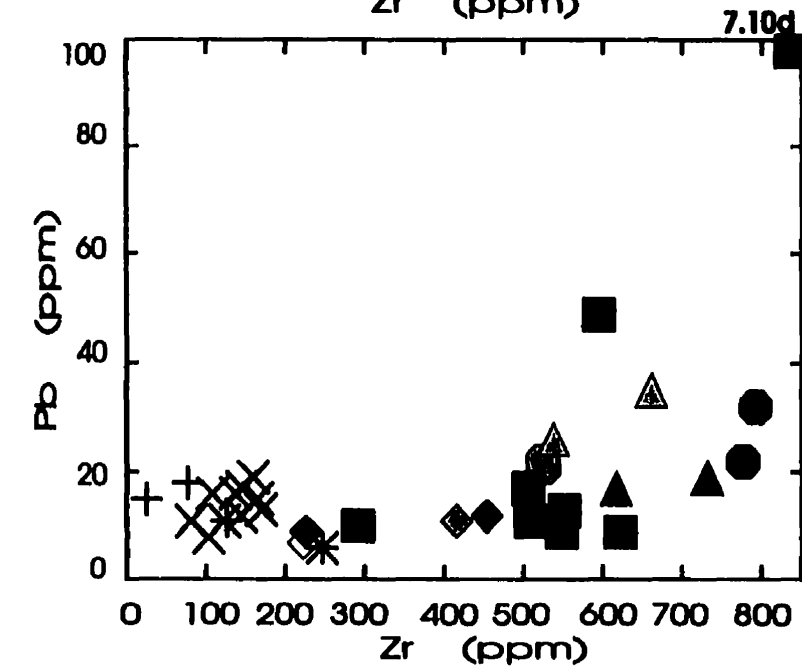
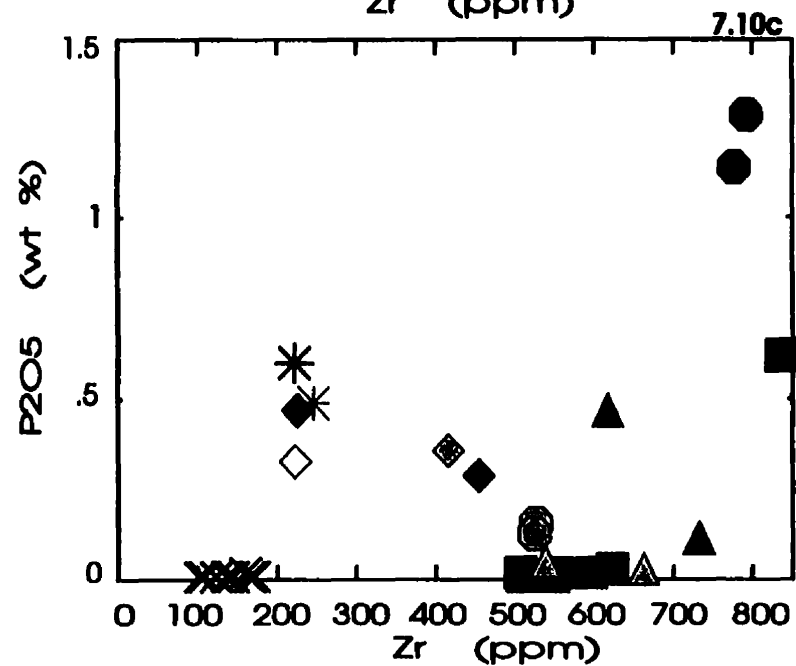
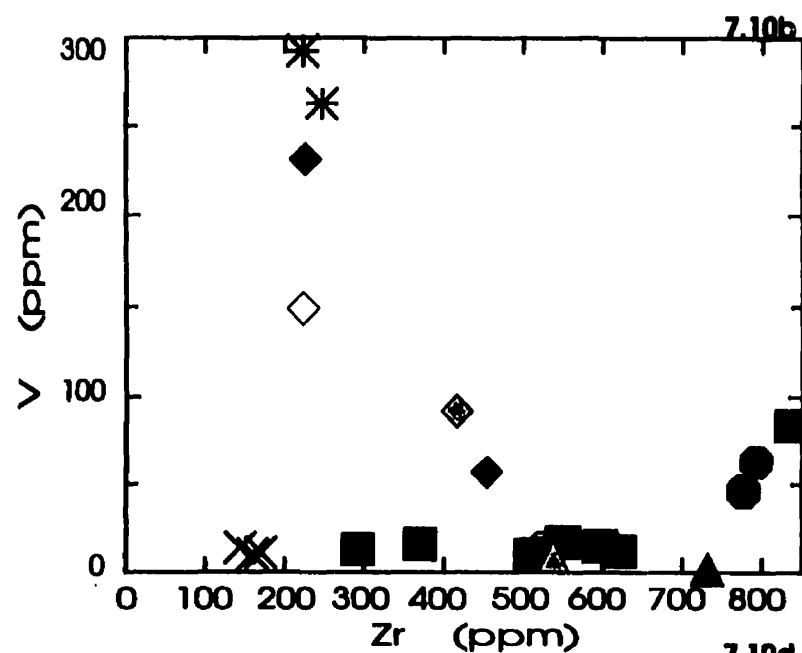
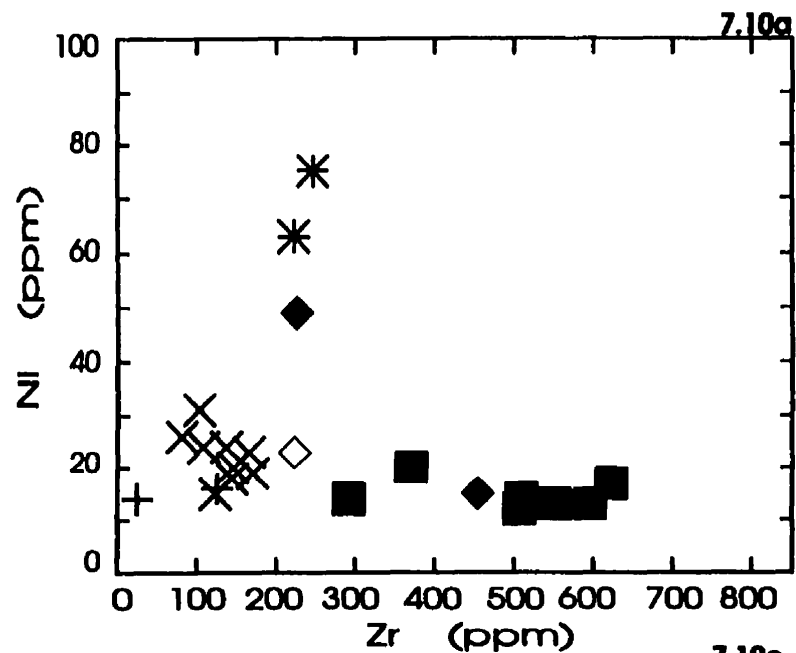
The trace element contents of the enclaves and their host rocks are presented in Figures 7.10a to l as a function of Zr content. In the Kwandonkaya suite, Zr is known to become progressively depleted as SiO_2 increases.

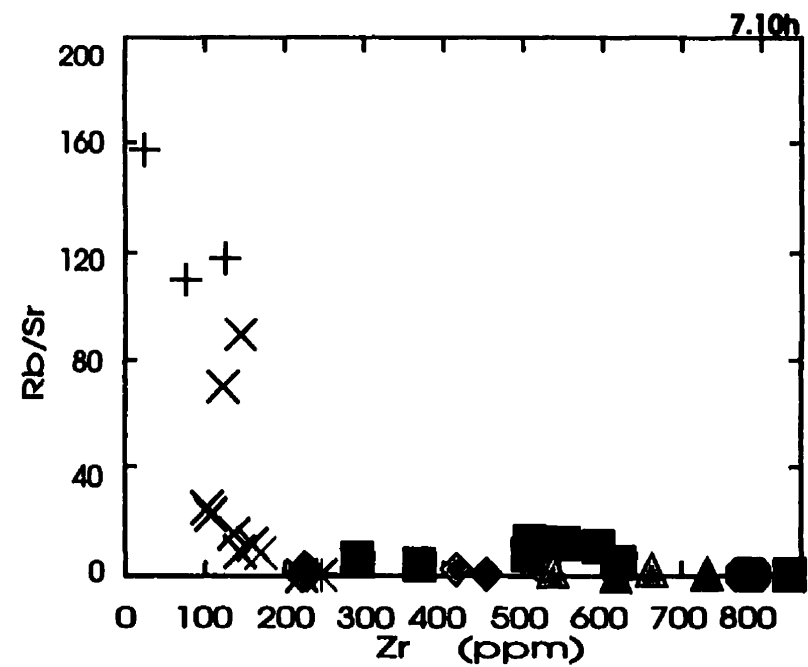
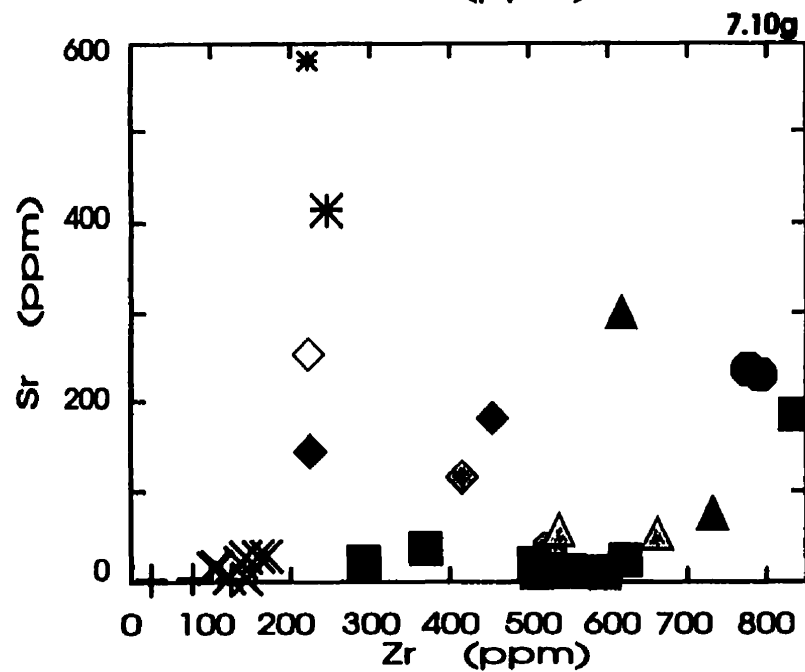
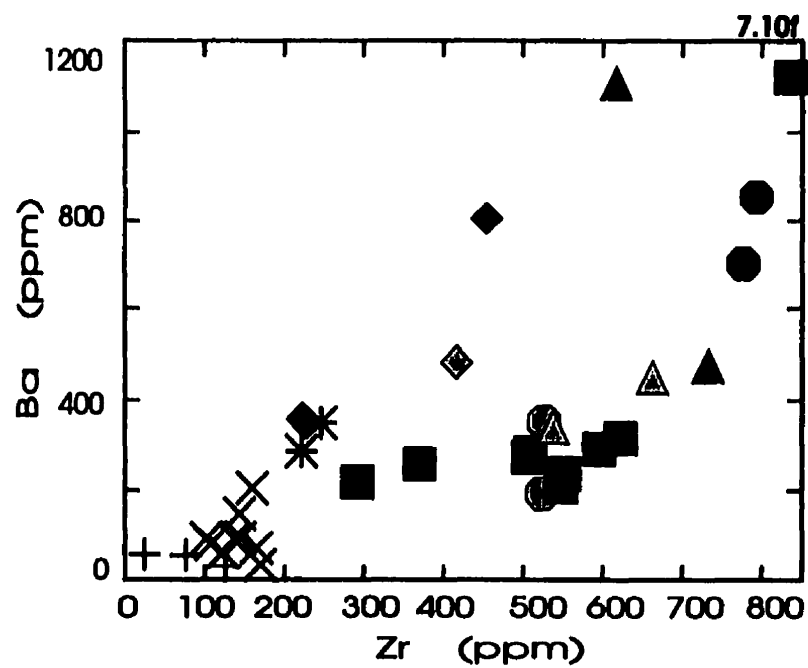
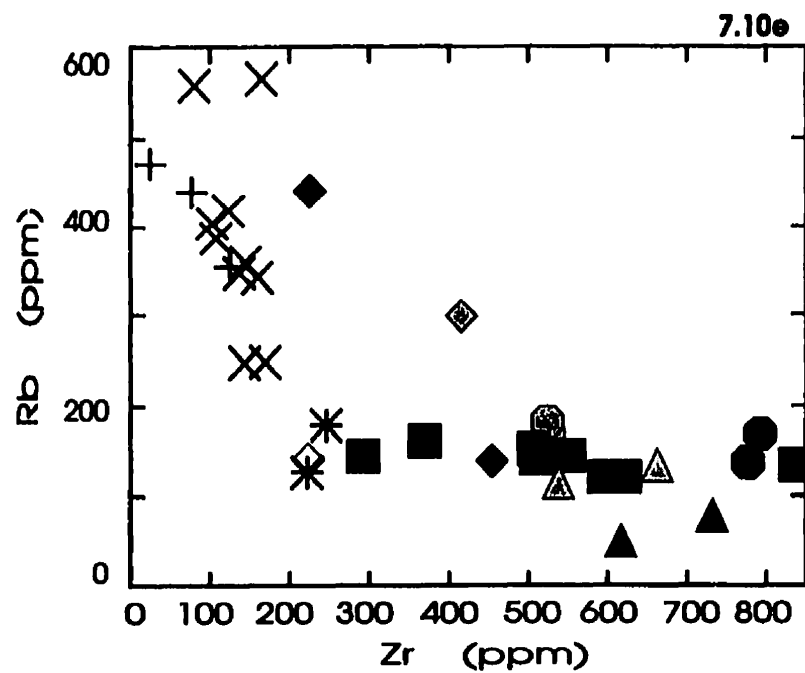
Unlike plots based on SiO_2 contents, in which the enclaves consistently plot between basaltic dyke and felsic subvolcanic rocks (Figs. 7.9a - h), the Zr content is quite variable (Figs. 7.10a - l). Whereas the enclaves in the Buri ring dyke do have a higher Zr content (750 - 850 ppm) than their felsic hosts (290 - 630 ppm), those in the plutonic rocks have lower Zr contents, intermediate between the felsic subvolcanic and plutonic rocks (210 - 460 ppm). The variations in Zr content do not suggest mixing of a magma represented by their host rocks with a relatively mafic melt to produce the enclaves.

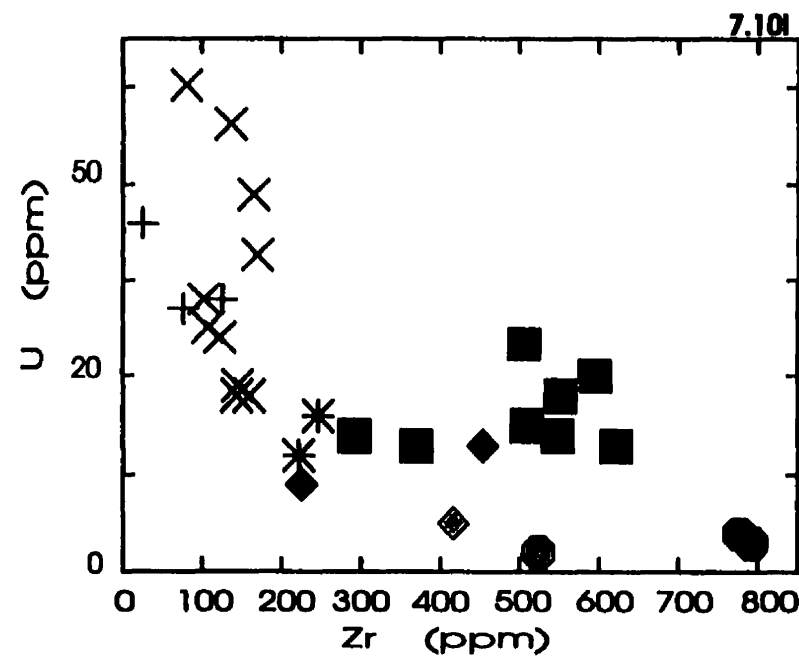
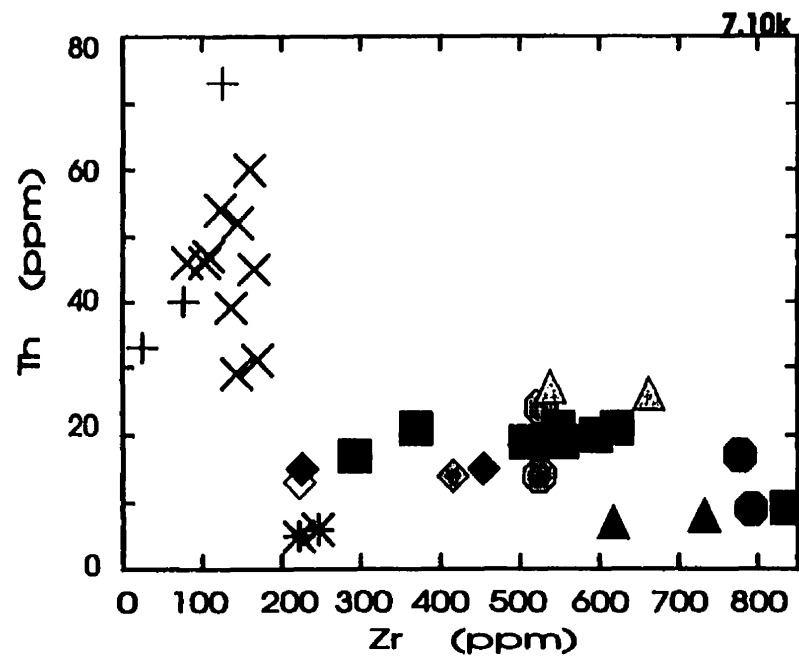
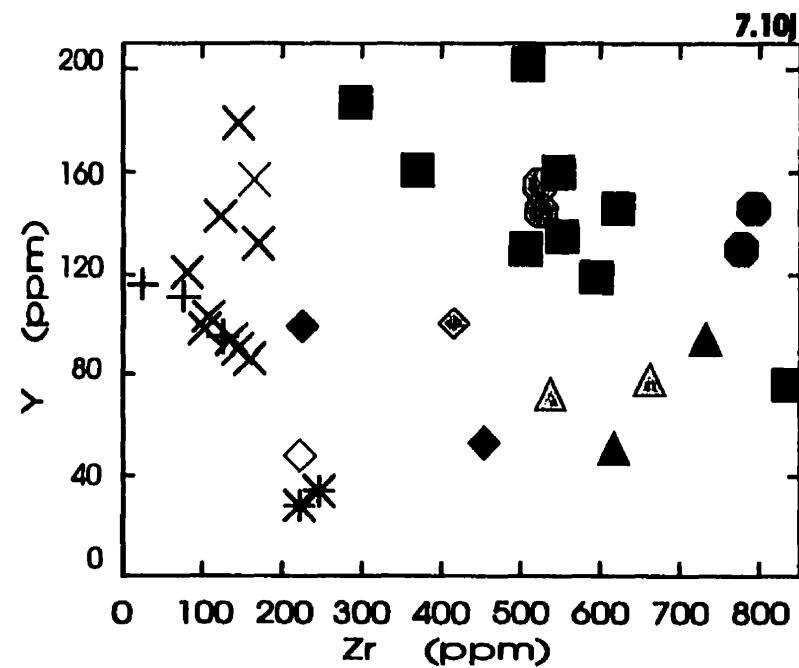
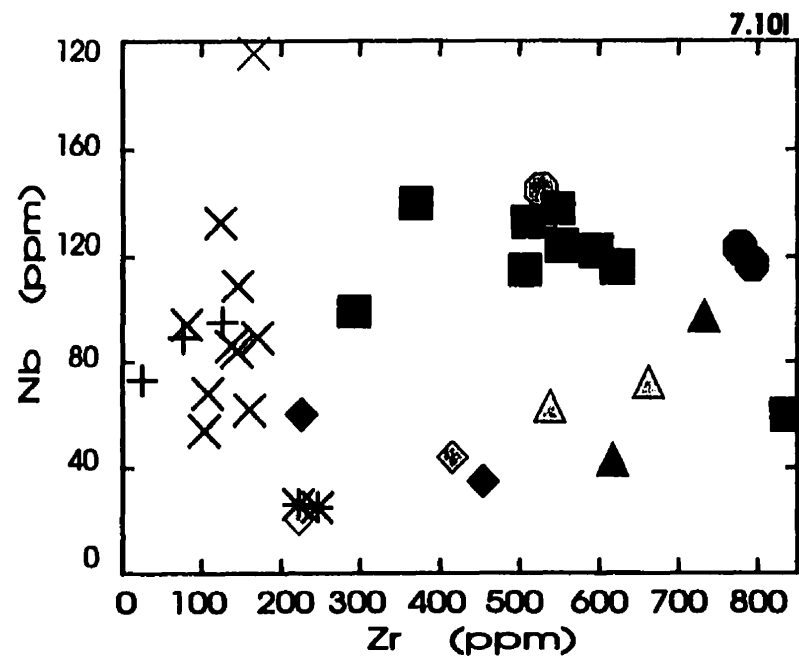
Plots of Ni, V, P_2O_5 , and Pb versus Zr (Figs. 7.10 a - d) cannot be reasonably explained

Fig. 7.10: Plots of trace element variations for enclaves and their host rocks: **a)** Ni (ppm), **b)** V (ppm), **c)** P_2O_5 (wt. %), **d)** Pb (ppm), **e)** Rb (ppm), **f)** Ba (ppm), **g)** Sr (ppm), **h)** Rb/Sr ratio, **i)** Nb (ppm), **j)** Y (ppm), **k)** Th (ppm), and **l)** U (ppm) versus Zr.

Symbols are the same as in Fig. 7.9.







by any model of mixing. Whereas the Ni content of the enclaves might indicate contamination of felsic rocks by mafic magma, the V and P contents of the enclaves are consistently at or below detection limits. The concentration of Pb, on the other hand (Fig. 7.10d), varies according to the degree of contamination of mafic magma by crustal material. It is, however, surprising that the microsyenodiorite enclave, inferred to have more limited interaction with the host rock, has abnormally high Pb content (ca. 100 ppm).

In plots of Rb, Ba, Sr, and Rb/Sr versus Zr (Figs. 7.10e - h, respectively), mixing cannot be inferred from the trends. Rb is relatively constant in subvolcanic rocks, enclaves and basaltic dykes (Fig. 7.10e), despite their differences in SiO_2 , Zr contents, and differentiation index (DI). The Ba content decreases progressively, with Zr, from the enclaves in the Buri ring dyke (and Kila-Warji augite syenite), via quartz syenites to the felsic subvolcanic and plutonic rocks. The Sr content (Fig. 7.10g) also decreases from enclaves in the Buri ring dyke, via syenites to the felsic subvolcanic and plutonic rocks. Once again, the basaltic dyke and enclaves in plutonic rocks do not give any easily interpretable trend. However, note that the Rb/Sr ratio is consistently low in the basaltic dykes, enclaves, and felsic subvolcanic rocks, and rises sharply in felsic plutonic rocks (cf. Chapter 2).

In the plots of Nb (Fig. 7.10i), and Y (Fig. 7.10j) versus Zr, both Nb and Y increase smoothly from basalts via enclaves in biotite granite and syenites to enclaves in subvolcanic rocks. The Nb and Y contents of felsic subvolcanic rocks, enclaves in the subvolcanic rocks, and biotite granite are quite similar (Figs. 7.10i, j). The observation that the enclaves do not plot at an intermediate location between mafic (basaltic) and felsic (rhyolitic) compositions indicates that the variation pattern of these elements are not amenable to simple mixing models.

The Th contents of the basalt and various enclaves are quite similar (Fig. 7.10k), despite increases in Zr, in the sequence, basalt via trachytes and enclaves in subvolcanic rocks. Figure

7.10l shows that the enclaves and basaltic dykes generally have lower U than the felsic host rocks, as expected in general terms.

7.6 Petrochemical model(s) of enclave evolution

A variety of criteria have been proposed, and extensively reviewed, to distinguish which of the processes best explains the composition of enclaves in plutonic suites (Langmuir et al. 1977, Allen 1991, Didier & Barbarin 1991, Liankun & Kuirong 1991, Tindle 1991). These criteria are applied so as to identify the best-fit model for the enclaves in the Kwandonkaya complex.

7.6.1 The restite model

This model implies that the enclaves are fragments of restite related to the production of the felsic magma that crystallized as the host rocks (cf. Chappell & White 1991).

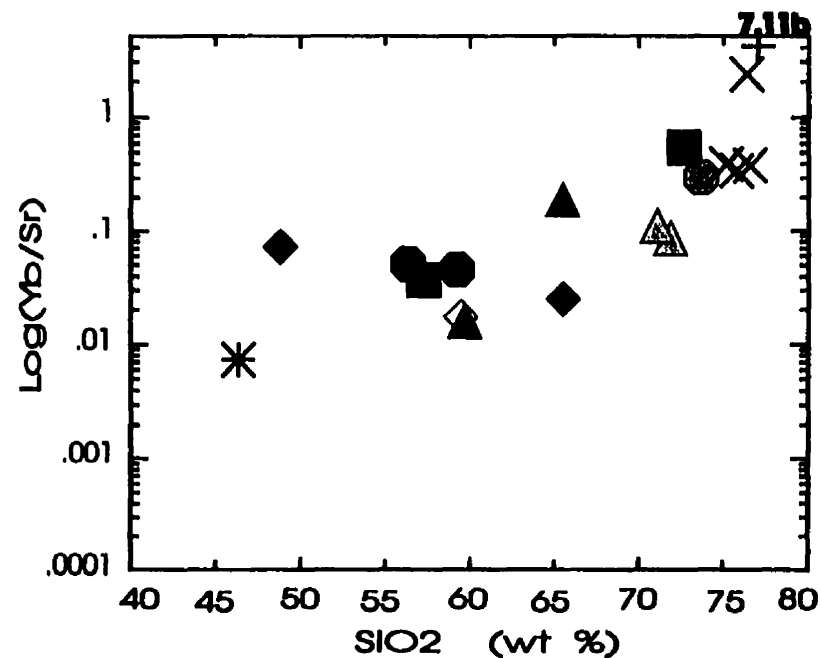
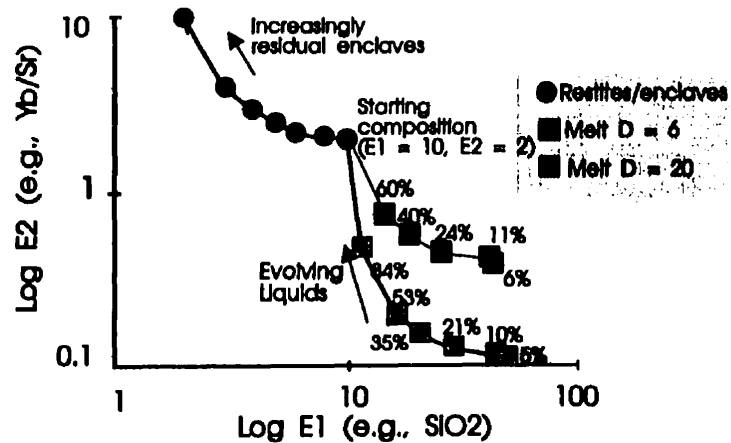
The equation that allows one to model partial melting (Tindle 1991) is: $C_i^o = FC_i^m + (1-F)C_i^R$, and $C_i^m = C_i^R/D$, where C_i^o = concentration of element i in the starting material (parent), C_i^R = concentration of element in the residue, C_i^m = concentration of element in the melt, F = weight fraction of melt produced, D = bulk distribution coefficient between residue and melt. If this model applies to the enclaves and their host rocks, then the melt and restite compositions should vary as shown in Figure 7.11a. It is clear in Figure 7.11b that the enclaves do not fit the restite hypothesis. Moreover, the occurrence of crenulate contacts, chilled margins, and alteration haloes are all indicative of the fact that the enclaves were molten (cf. Cantagrel et al. 1984, Vernon et al. 1988, Didier 1991), and not residual solids. Also, volcanic rocks with compositions similar to those of the enclaves are well documented in the NYG (Bennett et al. 1984, Bowden et al 1987). These include i) occurrences of trachyandesite flows with trachyte, basalt, and trachybasalt flows in Kila-Warji, Ningi-Burra, Tibchi, and Zaranda complexes, and ii) intrusive

Fig. 7.11: Geochemical plots to qualitatively appraise the restite (a - b), and mixing (c - d) models: **a)** Theoretical illustration of the restite or partial melting model (cf. Fig. 1d of Tindle 1991). The residual enclaves (or restites) are shown as filled circles. The starting composition is at element $E2 = 2$, and element $E1 = 10$. Element $E2$ favours the melt more strongly than element $E1$. The compositions of liquids derived from various degrees of partial melting are shown for two assumed values of bulk melt D . **b)** A plot of Log Yb/Sr ratio against SiO_2 (wt.%) for enclaves and their felsic host rocks. The trend of the suite of enclaves and their felsic hosts is not consistent with residue-partial melt relationship illustrated in Figure 7.11a. **c)** Ratio - ratio plot of Zr/Y vs. K/Rb . Elemental ratio plots of mixed magmas and the end members should fall on a hyperbola whose curvature is controlled solely by the end members, for a mixing model to be valid. Thus a plot of Zr/Y versus K/Rb (i.e., Fig. 7.11c) should give a hyperbolic distribution if mixing had occurred, which is not the case. **d)** A companion plot of K/Rb vs. Y/Rb . In a companion plot, such as Figure 7.11d, mixed rocks should fall on a straight line; again the Kwadonkaya enclaves and their felsic hosts fail this test. This plot (Fig. 7.11d) rather indicates that the trachyandesite enclaves and the enclaves in biotite granite have similar ratios to that of their host rocks.

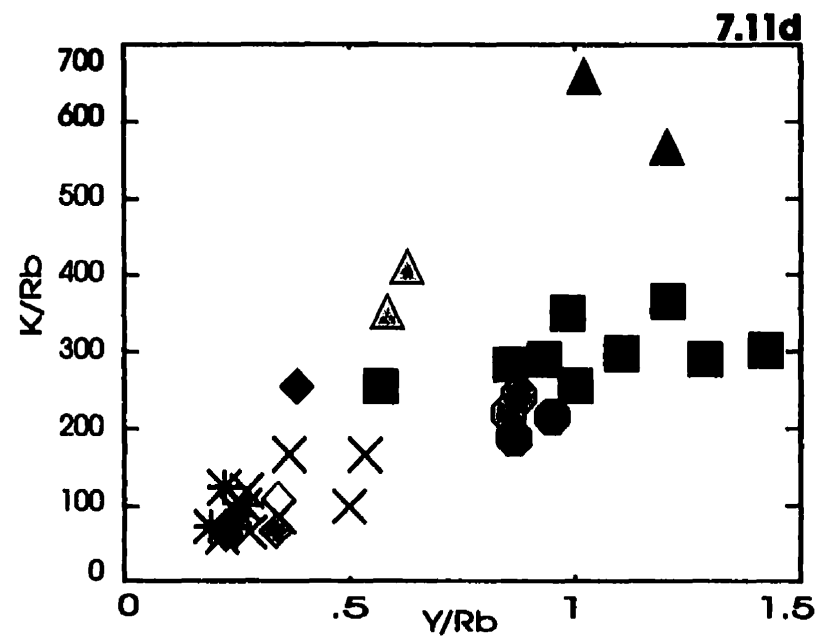
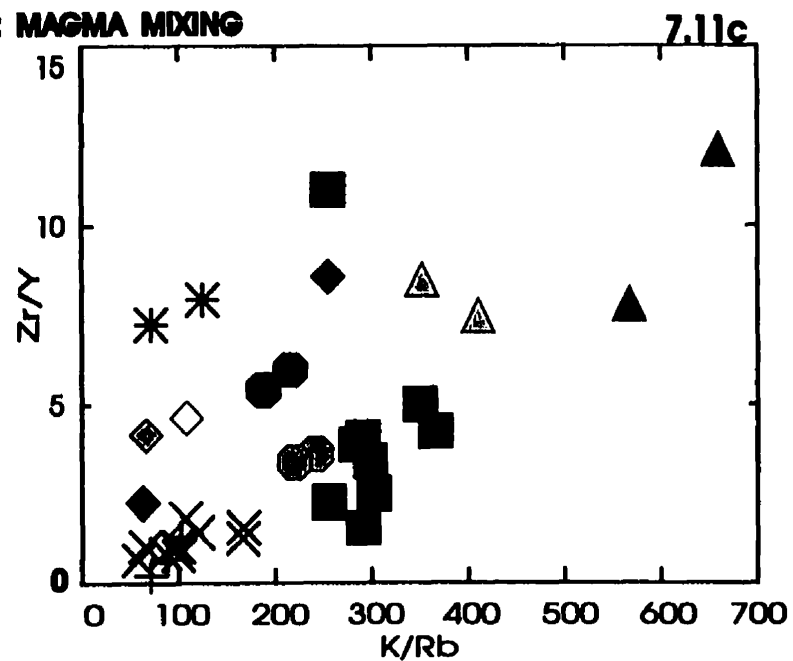
Plot symbols are the same as in Fig. 7.9.

7.11a

I: PARTIAL MELTING



II: MAGMA MIXING



equivalents, including gabbro and monzogabbro in net-veined acid-basic rocks of Kofai, Sha-Kaleri, Shere, Sara-Fier, and Jere-Sanga complexes. The enclaves thus cannot be considered fragments of restite.

7.6.2 Magma-mixing model

Mixing can be established qualitatively by using ratio-ratio plots (Langmuir et al. 1977). Elemental ratio plots of mixed magmas and the end members should fall on a hyperbola whose curvature is controlled solely by the end members, for a mixing model to be valid. Thus a plot of Zr/Y versus K/Rb should give a hyperbolic distribution if mixing had occurred. This is not the case, as seen in Figure 7.11c. Furthermore, a plot of one of the ratios against the ratio of the denominators, known as a companion plot, can further test for mixing (Langmuir et al. 1978). In such a companion plot, the mixed rocks should fall on a straight line, which is not the case in the companion plot K/Rb vs Y/Rb (Fig. 7.11d). Furthermore, Figure 7.11d indicates that the trachyandesite enclaves and the enclaves in biotite granite have similar ratios to those of their host rocks. These enclaves were thus equilibrated by diffusive exchange with respect to Rb and Y. Also, the fact that the enclaves and their host rocks do not conform to these mixing criteria indicates that fractionation, after mixing, was a more important process in controlling the present chemical composition of the enclaves.

Despite the above qualitative observations, it is assumed that an initial mixing occurred in some of the enclaves before mingling with their present host rocks. According to Langmuir et al. (1977) and Cantagrel et al. (1984), the extent of mixing can be calculated using the equation: $C_i^m = aC_i^A + (1-a)C_i^B$, which can be rewritten as: $(C_i^m - C_i^B) = a(C_i^A - C_i^B)$, which is an equation of a straight line with a slope of a (i.e., the proportion of magma A in the mixture: $0 < a < 1$). A and B are the magmas involved in mixing, $C_i^{m, A, \text{ or } B}$ is the amount of element i in

the mixture (m), and the mixing magmas A and B, respectively. The extent of mixing can be quantitatively assessed by plotting $(C_i^A - C_i^B)$ against $(C_i^m - C_i^B)$ and then, from the plot, determining the slope (a) to give the proportion of mixing. Alternatively, a can be derived by direct calculations using the relation: $a = (C_i^m - C_i^B)/(C_i^A - C_i^B)$. This latter approach was taken, because it is more straightforward, and the results of calculation for the different enclaves are given in Appendix VII.6. Assuming that the silica content of the enclaves was least affected by diffusive exchange and phenocryst recycling during mixing, the following proportions of trachytic to basaltic melts were calculated: 1) trachyandesite (sample 49M = 50:50, sample 87M = 66:34), 2) microsyenodiorite (55:45), and 3) trachybasalt (13:87). For the andesitic enclaves, the estimated proportion of mixing of dacitic and basaltic magmas is between 68:32 and 71:29, whereas using rhyolitic and basaltic magmas as end-members yielded a mixing ratio of 45:55. Very substantial differences are observed in the proportions on the basis of different elements (Appendix VII.6). The identity of the end-members during such mixing cannot be established with certainty. Other processes during subsequent mingling with the felsic magmas, such as the recycling (or removal) of phenocrysts from crystal-rich felsic magma (i.e., AFC), diffusive exchange and, possibly, metasomatic modifications are all problems that affect any quantitative model proposed. Overall, these other processes evidently contributed to various extents to enclave modification and their deviation from mixing lines. These factors are considered more important than the possible misidentification of the end-members involved in mixing.

7.6.3 Fractionation or cumulate model

The cumulate model implies a cogenetic relation between the enclaves and their felsic hosts. Assuming perfect fractional crystallization of the dominant felsic magma, the melt composition will vary according to Rayleigh's distillation law (Tindle 1991): $C_i/C_o = F^{(D-1)}$, where

C_1 = concentration of an element in the residual melt, C_0 = concentration of the same element in the initial melt, F = weight fraction of melt remaining, and $D = \sum X_i K_{di}$).

In a log-log element plot of a compatible versus an incompatible element, the trend of the enclaves (= possible cumulates) and their felsic host rocks (= melts) should be subparallel in a situation where Rayleigh fractionation had occurred (Fig. 7.12a). As seen in Figure 7.12b, the enclaves and their host rock do not produce such subparallel trends. Rather both the enclaves and the felsic rocks plot along a roughly collinear trend (Fig. 7.12b). This does not agree with the cumulate model or the expected trend during Rayleigh fractionation (Tindle 1991). Furthermore, the chondrite-normalized REE plots (Figs. 7.12c, d) further support the inference that the enclaves were blobs of melt that underwent some extent of crystal fractionation before mingling (with their felsic hosts) and solidification. The LREE content is similar in trachyandesite, rhyolite, and microsyenodiorite enclave, with the granite porphyry being slightly LREE-enriched (Fig. 7.12c). The HREE pattern is similar in the trachyandesite enclave, rhyolite porphyry, and granite porphyry, whereas microsyenodiorite is slightly HREE-enriched. Generally, the magnitude of the negative Eu-anomaly in the enclaves is less than that in their host-rocks. This is consistent with a lower degree of fractionation of enclaves than in the felsic host rocks.

In contrast, the enclaves in biotite granite are depleted in both LREE and HREE relative to their plutonic host rocks, with the exception of one sample that has similar HREE concentrations to the biotite granite. Generally, the enclaves have a Eu anomaly intermediate in magnitude with respect to that in basaltic dykes and their felsic host rocks. From the above observations, it is inferred that the enclaves were incorporated as coeval melts that underwent limited fractionation before mingling with their felsic host rocks.

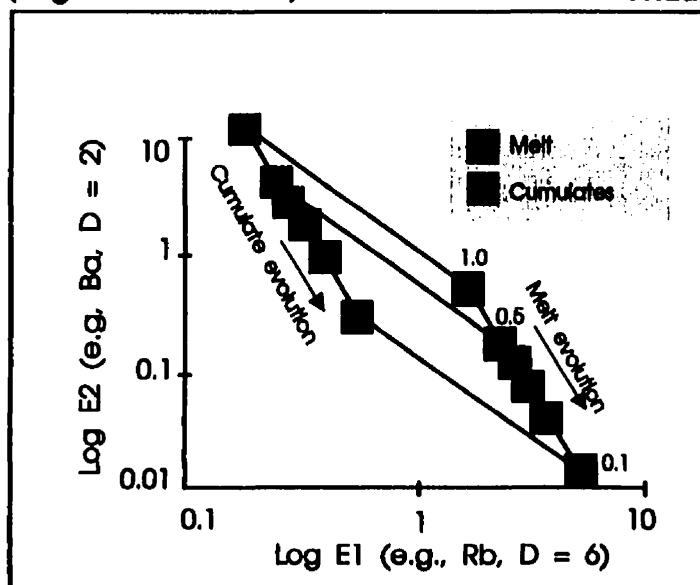
Fig. 7.12. Plots of trace elements to qualitatively appraise the fractionation model. Also chondrite-normalized REE plots for enclaves and their host rocks are shown: a) A schematic plot of theoretical trends of composition of enclaves and their host rocks expected in a cumulate model (cf. Fig. 1a of Tindle 1991, p. 315). Mafic cumulates are shown in filled squares, whereas the cogenetic melt is shown in open squares. In the log-log element plot of concentrations of compatible vs incompatible element (e.g., Ba vs. Rb), the evolutionary trend of the cumulate and cogenetic melts are expected to be subparallel during Rayleigh fractional crystallization. For a given starting composition, the data will fall in a compositional parallelogram bounded by "all melt" and "all cumulate" curves, which are straight lines on logarithmic plots. Crystal mushes could be produced, which vary in the same direction or along mixing lines. b) Plot of Log Ba vs. Log Rb. The plot symbols in Figure 7.12b are the same as in Figure 7.9. The composition of the enclaves and that of their coeval felsic hosts are inconsistent with the "melt" and "cumulate" trends illustrated in Figure 7.12a. c) chondrite-normalized REE plot for enclaves in Buri ring dyke and their host rocks. d) chondrite-normalized REE plot for enclaves in biotite granites and their host rocks.

The Eu anomalies in Figures 7.12c and d provide evidence that fractionation did occur in the enclaves, when they were still molten magmas before mingling with their felsic host rocks. The similarity in magnitude of REE pattern of trachyandesite and their rhyolite porphyry host rocks further indicates equilibration by diffusive exchange, but the enclaves are invariably characterized by lower Eu anomalies compared to their felsic subvolcanic and plutonic host rocks.

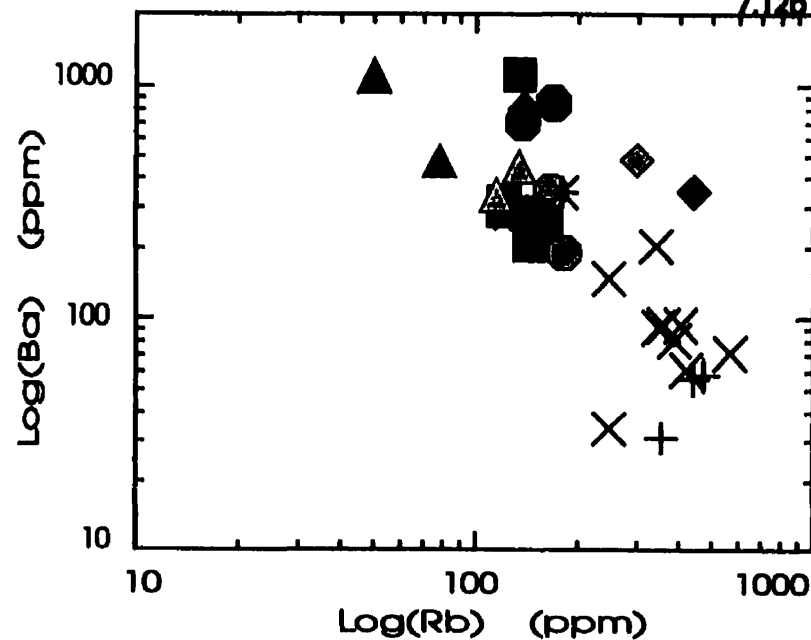
In Figure 7.12c: sample 43 = hedenbergite fayalite granite porphyry, 67M = microsyenodiorite enclave, 49F = rhyolite porphyry, 49M/87M = trachyandesite enclaves. In Figure 7.12d: sample 1B = porphyritic biotite granite, 73 = trachybasalt, 73C = trachydacite, 73B2 = andesite. In both Figures 7.12c and 7.12d sample 20 is basaltic dyke.

III: CRYSTAL FRACTIONATION
(cogenetic enclaves)

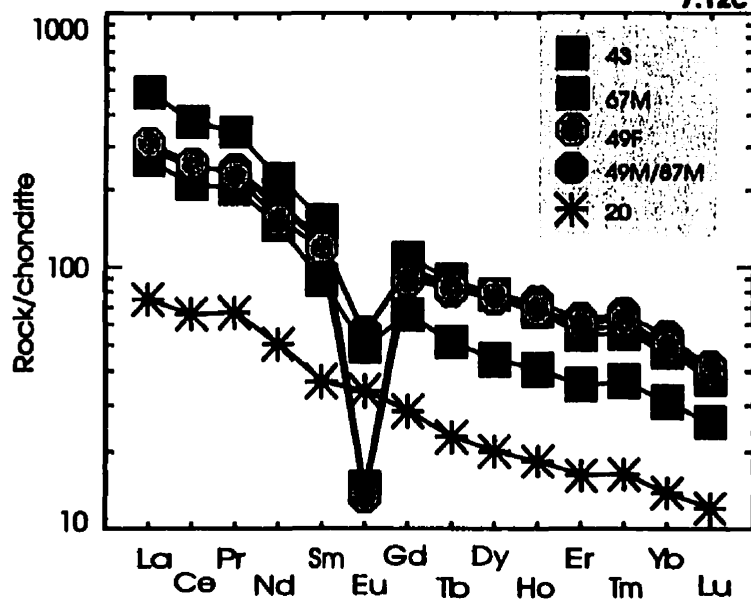
7.12a



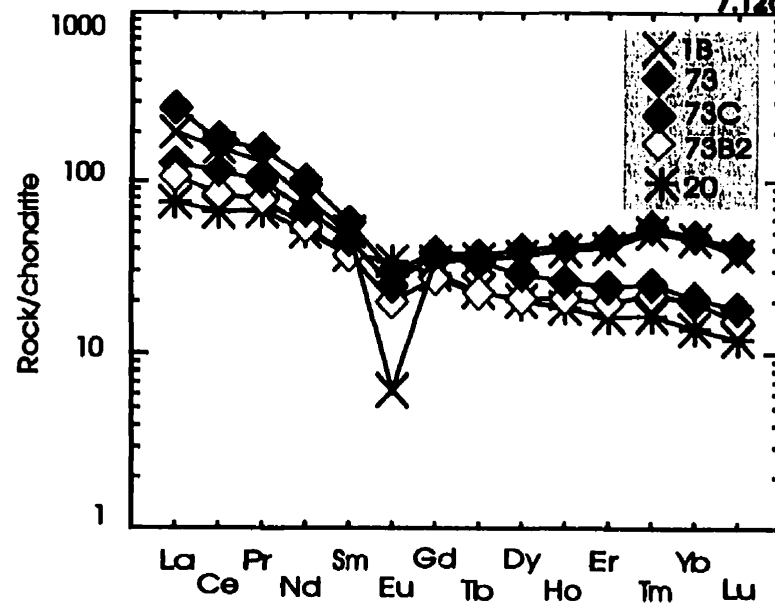
7.12b



7.12c



7.12d



7.6.4 Other modes of enclave modification

Chemical diffusion can be expected with time to degrade the chemical gradient between enclaves and their host magma, and the relics of such a process may be preserved as zoned enclaves. One striking example of a zoned enclave depicting modification due to diffusion is a trachyandesite enclave in sample 49 (Fig. 7.2d, Appendix VII.1a). The core-to-rim variations are summarized in footnotes in Table 7.7, which shows increases from core to rim in SiO_2 (3 to 4%), MnO (0.01 to 0.05%), ZnO (0.10%), Na_2O (0.03 to 0.06%) and K_2O (0.35 to 0.45%), and also decreases in TiO_2 (0.09 to 0.14%), Al_2O_3 (0.39 to 1.19%), FeO (2.18 to 2.42%), MgO (0.49 to 0.53%), and CaO (1.01%).

Generally, diffusion in a melt is affected by a plethora of factors, which include: state of enclaves and host rocks, field strength of the elements, temperature, pressure, viscosity, alkali/alumina ratio, vapour content, and the size of the enclaves (Baker 1990, 1991, Baker & Bossanyi 1994). It is, therefore, possible to produce a spectrum of compositions by diffusive exchange between enclave magmas and their host rock magmas. In fact, Debon (1991) has shown that diffusion between enclave and host can produce compositional modifications completely different from those due to simple two-end-member mixing. It is beyond the scope of this study to either constrain the above factors or quantitatively characterize the nature and effect of diffusive exchange. However, petrographic, mineral chemical evidence, and observations of similarities in contents of Pb (Fig. 7.10d), Rb (Fig. 7.10e), Y (Fig. 7.10j), K/Rb and Y/Rb (7.11d), and REE (7.12c, d) in some enclaves and their host rocks clearly suggest that limited diffusive exchange accompanied the mingling of phenocryst-bearing enclave-forming magma with the host magma.

Evidence for metasomatic modification is limited in microsyenodiorite, trachybasalt and orb-bearing basalt enclaves. This evidence is mainly in the form of secondary biotite, chlorite,

epidote, fluorite, sericite, and Fe-Ti oxides replacing a primary amphibole (or pyroxene), and plagioclase. As these minerals are all of secondary origin, this assemblage indicates the preponderance of potassic alteration of the enclaves, which can mask the effects of diffusion between coexisting magmas. The lack of chemical data on the altered enclaves precludes any quantitative modelling.

7.7 Discussion

The petrographic studies reveal a variety of textural features which, along with mineral chemical studies, can help in the interpretation of processes responsible for the formation of enclaves. Petrochemical models are proposed to explain the present characteristics of enclaves.

7.7.1 Petrographic, mineral chemical and petrochemical features

Textural features observed in enclaves (Table 7.1) include ocellar, orbicular textures, pillow-like shapes, crenulate margins, chilled margins, and reaction rims of enclaves (Figs. 7.2 - 4). All indicate liquid-liquid contacts between enclaves and their host and also preclude the possibility that the enclaves were restites. Ocellar textures (Figs. 7.2e, f, and 7.4c, f) are interpreted as originating by diffusion of alkalis in a boundary layer around xenocrysts (Sato 1975, Nakada 1991), and consequently producing a rim of amphibole or mica on such xenocrysts. The orbicular texture documented in some basaltic enclaves (Figs 7.4e) probably resulted from superheating followed by supercooling (Vernon 1991, Ort 1992). According to these authors, superheating occurs as a result of the intrusion of a hotter magma, which consequently resorbs the framework of silicate nuclei. Subsequently, nucleation occurs along an available substrate like wall-rock fragments, refractory crystals, or pre-existing phenocrysts in the melt. Usually, in such cases, the pre-existing phenocryst phase is partly resorbed, and heterogeneous nucleation proceeds

after mixing (Ort 1992).

Mineral-chemical evidence of mixing includes the polymodal composition of minerals, in which two or more populations of a single phenocryst phase within an enclave provides evidence for disequilibrium, e.g., plagioclase (Fig. 7.5). Features suggesting complex changes in chemical composition and temperature during mixing include: a) resorption textures cutting across concentric zoning in plagioclase within trachybasalt and trachydacite, along with an oligoclase mantle on more An-rich (labradoritic) plagioclase grains in enclaves within biotite granite (Figs. 7.4i, j), b) the occurrence of dusty brown zones on margins of plagioclase or pervading some grains (Figs. 7.3a, c), c) the presence of reversely zoned hedenbergite and ferroaugite, sieve-textured hedenbergite in enclaves, and reverse zoning of hedenbergite phenocrysts in the felsic subvolcanic rocks, and d) polymodal amphibole compositions in andesitic and dacitic enclaves. The occurrence of plagioclase with truncated growth-zones or repeated resorption-surfaces, and reversely zoned phenocrysts of oligoclase, further indicate that the minerals were not in equilibrium with their current hosts (cf. Cantagrel et al. 1984, Tsuchiyama 1985, Nixon & Pearce 1987, Nixon 1988a, b, Castro et al. 1990, Vernon 1991). Reverse zoning is typical of pyroxene and plagioclase derived from low-temperature felsic magma and subsequently mixed with a more mafic magma. The presence of disequilibrium features also indicates incomplete homogenization by diffusion in the time available.

On the basis of petrochemical evidence, I propose that most of the enclaves formed by the mixing of mafic and intermediate to felsic magmas with different crystal(phenocryst):melt ratios, and hence different degrees of fractionation, in one or more chambers. The hybrid magmas were then mingled with batches of rhyolitic magma (their present host rocks) in the subvolcanic magma chamber. Fractionation in the hybrid magma, diffusive exchange between the hybrid magma and felsic host during commingling, followed by metasomatic alterations at a subsolidus

stage, have modified the enclave compositions from the initial mixing lines. Similar interpretations, involving various combinations of fractionation, mixing, and diffusive exchange, have been proffered for enclaves with petrographic and mineral chemical features depicting mixing but having petrochemical features inconsistent with such observations (Gourgaud & Maury 1984, Crisci et al. 1989, Liankun & Kuirong 1991, Tindle 1991, Orsini et al. 1991).

7.7.2 Mechanism of hybridization, and structural implications

The factors controlling the nature and degree of felsic melt - mafic melt interaction have been extensively reviewed (Huppert et al. 1984, Baker 1990, 1991, Fernandez & Barbarin 1991, Tindle 1991). These include the initial temperature, relative volumes, composition, degree of crystallization, absolute viscosities, and relative rheology. The assessment of these factors is beyond the scope this work, as evidence for magmatic interactions is clouded by the hydrothermal overprint. In this study, the mechanism of hybridization is first appraised, and this is followed by a discussion of structural implications.

Mechanism of magma hybridization (mixing and mingling)

Two contiguous magmas will mix very effectively if their viscosities match, whereas high contrasts in viscosity impede hybridization (cf. Fernandez & Barbarin 1991, Mungall 1993). Usually, partially crystallized felsic magmas behave as viscoplastic bodies during their intrusion history; depending on their rheological properties, they will undergo various extents of mixing with intermediate and mafic magmas.

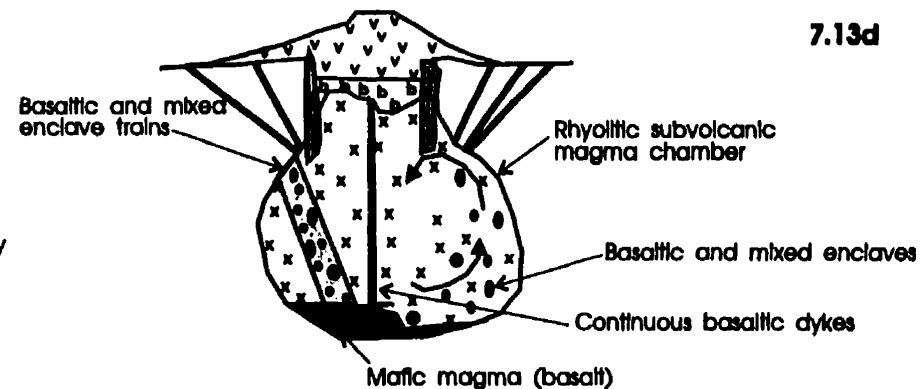
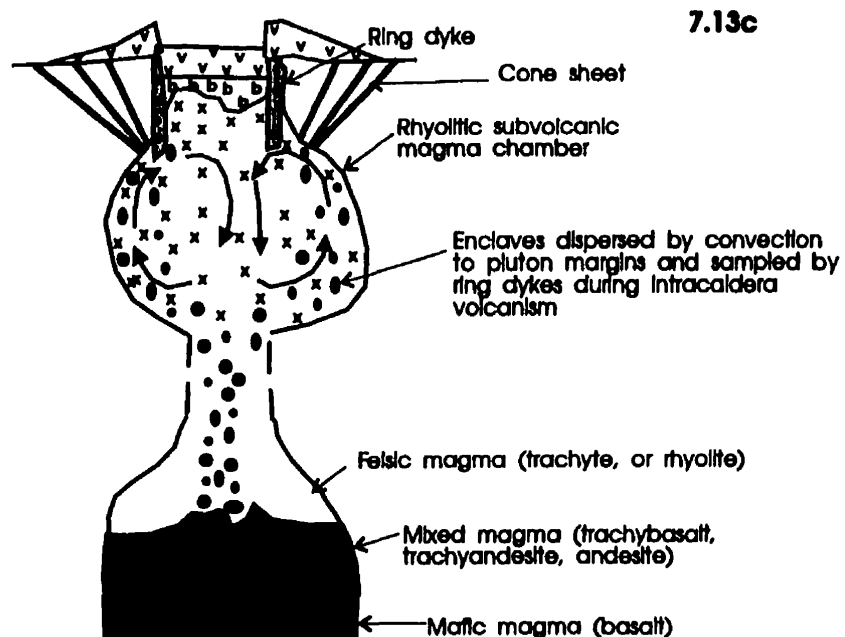
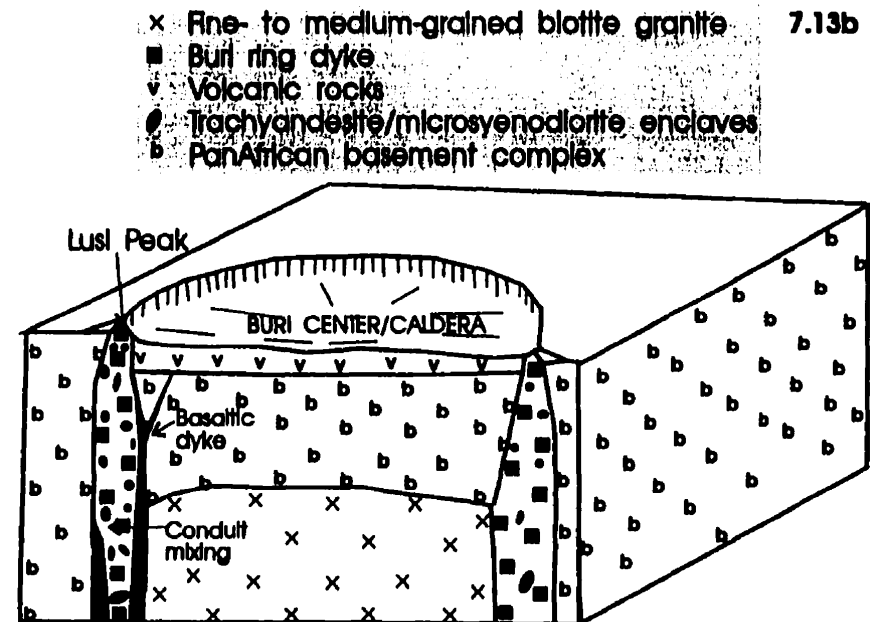
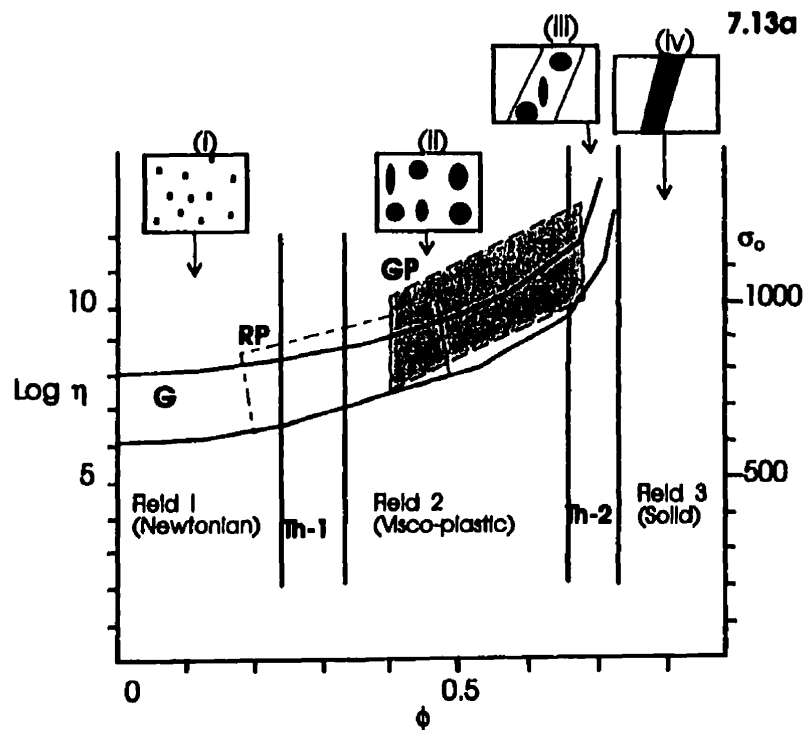
The rheological behaviour of crystal-bearing magmas can be assessed using the equation (Soo 1967): $\sigma_o = k \cdot \phi^3$, where k is a constant (3×10^4 dyne cm^{-2} , Spera 1980), σ_o represents the yield strength of a viscoplastic body, and ϕ , volume fraction of crystals. The different fields of

behaviour of crystallizing magmas are schematically illustrated in Figure 7.13a (cf. Fernandez & Barbarin 1991). Within Field 1, a rhyolitic magma, with moderate viscosities (10^6 to 10^8 poises), is expected to behave as a newtonian liquid. Mafic or intermediate magma will be more effectively mixed with a felsic melt within this field (case i in Fig. 7.13a). Since there is no evidence for high degrees of mixing between the enclave magma and their felsic host, it is inferred that most enclaves were formed within the vicinity of Th-1 (threshold 1).

The crystal-rich rhyolitic magma that formed rhyolite porphyry ($0.20 < \phi < 0.50$) falls essentially within the viscoplastic field (Fig. 7.13a), and thus solidified in a range typical of Bingham bodies (cf. Gourgard 1991). Therefore, the commingling of trachyandesite and crystal-rich rhyolitic magmas was favoured, and resulted in the formation of enclaves with different degrees of mixing between the two (case ii in Fig. 7.13a). Also, crystal-rich magmas that produced granite porphyries ($0.40 < \phi < 0.65$) underwent little or no mixing with trachyandesite (microsyenodioritic) magmas. The enclaves in the Buri ring dyke were entrained along conduits (during intracaldera volcanism) (Figs. 7.13b, c).

In the case of porphyritic biotite granite of the Panshanu center, three types of mingling occurred. In the first instance, hybridized mafic magma mingled within the range of the second threshold (Th-2 of Fig. 7.13a) with the crystal-rich magma that formed porphyritic biotite granite ($0.65 < \phi < 0.75$), producing disrupted dykes and trains of enclaves (case iii in Fig. 7.13a). At this stage, granites can undergo fracturing, thereby creating passageways for synmagmatic dykes, which produce pillow-like enclaves with interaction with the felsic host (cf. Fernandez & Barbarin 1991, Gourgard 1991). In some cases, the mafic dykes were emplaced once the crystal-rich felsic magmas ($\phi > 75\%$) were mostly crystalline, and within rheological Field 3. Felsic magmas in this field behave as brittle solids that can be affected by fracturing (Gourgard 1991, Fernandez & Barbarin 1991). The size of the body of mafic or intermediate magma has an important bearing

Fig. 7.13: Illustration of mechanisms of mixing and structural implications. **a)** Plot of ϕ (volume fraction of crystals) versus σ_0 (yield strength in dynes cm⁻²) and $\log \eta$ (where η = viscosity in poises) to illustrate the main stages of evolution of crystallizing felsic magmas (cf. Fernandez & Barbarin 1991). Th-1: first rheological threshold; Th-2: second rheological threshold. **G** = trend of the rheological properties of granitic bodies (Fernandez & Barbarin 1991), **RP** = Inferred range of rheological properties of rhyolite porphyry estimated from ϕ , **GP** = Inferred range of rheological properties for granite porphyry, and **i, ii, iii, iv** = different mixing scenarios depending on degree of crystallization. **b)** Schematic illustration of mingling of hybridized enclaves in the subvolcanic Buri ring dyke of the Kwandonkaya complex. **c)** Formation of hybridized enclaves by mixing before mingling in the Buri ring dyke, and parts of Panshanu center. This is similar to the case of enclaves from the Mukaiyama rhyolitic dome in Japan (Koyaguchi 1991). **d)** Formation of basaltic enclaves in the biotite granite of the Panshanu center. This is similar to the case of enclaves in the Achiyama rhyodacitic dome in Japan (Koyaguchi 1991).



on the nature of its interaction with near-completely crystallized felsic bodies (granites). If the body of mafic or intermediate magma is less than 5 m (in width), it likely will form continuous dykes (case iii in fig. 7.13a), but if the width is greater than 5 m, the mafic or intermediate body remains molten for a sufficiently long period of time to cause overheating of the host granite, induce local convection, and produce trains of enclaves (cf. Pitcher 1991, Gourgaud 1991).

Structural implications

Enclaves of hybridized magma originated where blending of magma reached completion before mingling with their felsic host (Fig. 7.13c). In fact, the majority of the Kwandonkaya enclaves, including the trachyandesite, microsyenodiorite, trachybasalt, and andesite, are interpreted as mixed enclaves, similar to the "hybridized enclave" model of Koyaguchi (1991), described in studies of the Mukaiyama rhyolitic dome in Japan. In this model, the basaltic magma is ponded below a felsic one, and mixing occurs at depth before the mixed magma is tapped and mingled with the felsic rock at the subvolcanic level (Fig. 7.13c). The process of mingling and mixing can occur for any number of reasons (Sparks et al. 1977, 1984, Huppert et al. 1984, Blake & Campbell 1986, Koyaguchi & Blake 1991): a) the injection of hotter mafic magma into a felsic body of magma, b) overturn of a layered reservoir of magma due to crystallization of the lower mafic magma and the production of a residual less dense magma underneath the main felsic magma, c) forced convection during chamber evacuation and magma ascent through conduits such as ring dykes and volcanic pipes, and d) forcible intrusion of a rising batch of magma into an upper layer resulting into turbulence and, possibly, explosive eruption. The preponderance of enclaves in rhyolite porphyry of the Buri ring dyke suggests that the mingling of trachyandesite and the rhyolitic melt was assisted by the intrusion of coeval felsic and mafic melts, cauldron subsidence, and possibly explosive cycles of eruption.

Enclaves of the second type vary from trachybasaltic via dacitic to "andesitic" enclaves (Fig. 7.13d). These usually occur as trains of enclaves along pluton margins, and mostly are a result of thermal disequilibrium between a swarm of basaltic dykes and the felsic host magma. Some of the enclaves dispersed in the Panshanu center most likely indicate the disruption of the basalt-rhyolite interface resulting in mingling (Figure 7.13d).

7.8 Summary and conclusions

1) From petrographic and petrochemical studies, the enclaves in the Kwandonkaya subvolcanic and plutonic rocks do not correspond to restite produced by melting that yielded the felsic rock units. The petrographic evidence rather suggest liquid-liquid contacts; mineral chemical evidence indicates various degrees of mixing.

2) The hybridization in most cases occurred initially in mesozonal magma chambers, to produce mixed enclaves, before mingling with their rhyolitic host-rocks.

3) The petrochemical data for the enclaves cannot be modelled as resulting from simple two end-member mixing, because a variety of processes followed the initial mixing and the subsequent stages of mingling: fractionation, phenocryst recycling (i.e., the multistage incorporation of phenocrysts from the felsic host rocks by enclaves, and vice versa), and diffusive exchange between the enclaves and their host rocks.

4) Furthermore, some of the enclaves either assimilated basement rocks, particularly the andesitic enclaves within plutonic rocks, or were in addition affected by metasomatic modifications, which are not well constrained. These processes contributed to different extents in shifting the enclave compositions from trends expected by mixing or fractionation.

5) The microsyenodiorite enclaves in the Buri ring dyke have petrographic, mineral chemical, and petrochemical characteristics analogous to the least-evolved syenite (parental Kila-

Warji augite syenite) yet documented in the Nigerian Younger Granite province. The hypothesis that these enclaves could fractionate (like their syenitic analogues in Kila-Warji complex) to yield the rhyolitic subvolcanic magma is further explored in the petrogenetic modelling of the felsic rocks.

6) The enclaves in the biotite granite, on the other hand, do not show characteristics consistent with possible compositions of parental melt.

7) Overall, the presence of dark enclaves indicates that mafic melt was available throughout the evolution of the Kwandonkaya complex.

Table 7.2: Results of electron-microprobe analyses of plagioclase in enclaves within biotite granite: I (Trachybasalt: 73B); II (Andesite: 73B1, 73B2); III (trachydacite: 73C); IV (Anorthosite: 73B4). c = core, i = intermediate zone, m = margin, t = mantle. *Ab* = albite, *Ol* = oligoclase, *And* = andesine, *Lab* = labradorite, *Byt* = bytownite, *An* = anorthite, and *Or* = orthoclase.

# n	I							II					III				
	73B.c1	73B.m1	73B.t1	73B.c2	73B.m2	73B.6	73B.7	73B1.1	73B1.2	73B2.1	73B2.2	73B2.3	73C.c1	73C.m1	73C.c2	73C.m2	73C.c3
	1	1	1	1	1	6	1	9	1	3	2	1	3	3	2	2	2
SiO2	61.81	62.91	61.53	52.83	53.00	64.68	53.20	53.73	65.59	53.98	63.18	63.94	59.11	61.40	60.89	63.63	62.80
Al2O3	23.71	22.62	23.52	29.08	29.36	22.26	29.36	28.97	21.57	28.17	23.20	22.88	25.19	23.78	24.16	22.31	22.63
Fe2O3	0.09	0.17	0.09	0.51	0.22	0.09	0.13	0.25	0.14	0.23	0.09	0.12	0.10	0.10	0.13	0.12	0.08
CaO	5.07	4.11	5.20	12.02	11.92	3.57	12.07	11.61	2.76	11.43	4.63	4.29	7.06	5.36	5.83	3.52	3.96
Na2O	8.63	9.01	6.52	4.32	4.34	9.38	4.36	4.70	9.67	4.77	8.64	8.96	7.39	8.45	8.10	9.50	9.06
K2O	0.26	0.26	0.23	0.32	0.22	0.16	0.15	0.19	0.38	0.20	0.20	0.17	0.20	0.16	0.23	0.16	0.22
BaO	0.00	-	-	0.01	0.04	0.05	-	0.01	0.02	0.04	0.02	-	0.06	0.02	0.07	0.02	0.02
SrO	-	-	-	-	0.04	-	-	-	-	-	-	-	-	-	-	-	-
Total	99.57	99.08	97.09	99.10	99.12	100.20	99.27	99.46	100.14	98.82	99.96	100.35	99.10	99.27	99.43	99.25	98.77

Structural formula on the basis of 8 oxygens

Si4+	2.75	2.81	2.78	2.42	2.42	2.85	2.42	2.44	2.88	2.47	2.79	2.81	2.66	2.74	2.72	2.83	2.81
Al3+	1.24	1.19	1.25	1.57	1.58	1.15	1.58	1.55	1.12	1.52	1.21	1.19	1.33	1.25	1.27	1.17	1.19
Fe3+	0.00	0.01	0.00	0.02	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca2+	0.24	0.20	0.25	0.59	0.58	0.17	0.59	0.57	0.13	0.56	0.22	0.20	0.34	0.26	0.28	0.17	0.19
Na+	0.75	0.78	0.57	0.38	0.38	0.80	0.39	0.41	0.82	0.42	0.74	0.76	0.65	0.73	0.70	0.82	0.79
K+	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Ba2+	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
Sr2+	-	-	-	-	0.00	-	-	-	-	-	-	-	-	-	-	-	-
Ab	74.39	78.68	68.32	38.68	39.18	81.86	39.20	41.82	84.47	42.49	76.26	78.34	64.68	73.34	70.58	82.29	79.53
An	24.13	19.84	30.09	59.45	59.50	17.25	59.93	57.08	13.33	56.32	22.60	20.71	34.15	25.75	28.07	16.82	19.20
Or	1.48	1.48	1.59	1.87	1.32	0.89	0.87	1.09	2.20	1.19	1.13	0.95	1.18	0.91	1.34	0.89	1.28
	Ol	Ol	And	Lab	Lab	Ol	Lab	Lab	Ol	Lab	Ol	Ol	And	Ol	Ol	Ol	Ol

# n	III (cont'd)			IV
	73C.m3	73C.c4	73C.m4	73B4*
	3	1	1	2
SiO2	61.18	64.28	63.19	52.62
Al2O3	23.83	21.96	22.57	27.92
Fe2O3	0.14	0.17	0.07	-
CaO	5.24	3.19	4.02	16.76
Na2O	8.29	9.61	9.11	2.70
K2O	0.21	0.18	0.13	-
BaO	0.03	0.05	0.03	-
SrO	-	-	-	-
Total	98.91	99.43	99.14	100.00

Structural formula on the basis of 8 oxygens

Si4+	2.74	2.85	2.81	2.40
Al3+	1.26	1.15	1.18	1.50
Fe3+	0.00	0.01	0.00	-
Ca2+	0.25	0.15	0.19	0.82
Na+	0.72	0.83	0.79	0.24
K+	0.01	0.01	0.01	-
Ba2+	0.00	0.00	0.00	-
Sr2+	-	-	-	-
Ab	73.21	83.63	79.77	22.61
An	25.56	15.34	19.46	77.39
Or	1.23	1.03	0.77	-
	Ol	Ol	Ol	Byt

Table 7.3: Results of electron-microprobe analyses of pyroxene in enclaves and their host rocks:

Rhyolite porphyry:- a) greenish rhyolite porphyry:- 87.1+ (unzoned grain), 87.2c+, 87.2i+, & 87.2r+ (core, intermediate zone, and rim of zoned) hedenbergite, b) grey rhyolite porphyry:- 49B+: microphenocrysts. **Trachyandesite enclave in rhyolite porphyry:-** 49B.1 (sieve-textured grain), 49B.2c, 49B.2r (core and rim of zoned grain) of hedenbergite. **Hedenbergite fayalite granite porphyry:-** 67.1+: microphenocryst, 67.2c+, 67.2i+, 67.2r+ = core, intermediate zone and rim of zoned hedenbergite. **Microsyenodiorite enclave in hedenbergite fayalite granite porphyry:-** 67.1 (orthopyroxene), 67.2 (Mg-rich hedenbergite), 67.3c (Fe-rich augite core), and 67.3r (Mg-rich hedenbergite rim). In this and subsequent Tables:- n.d. = not determined, dash = below limit of detection, and zero indicates values less than 0.005. # = analyses number, n = number of analyses averaged, with individual spot analyses having an n of one. Analyses with plus sign at the end (e.g., 87.1+) are for felsic host rocks, others analyses are for enclaves.

Table 7.3

	Rhyolite porphyry					Trachyandesite enclave			Hedenbergite fayalite granite porphyry				Microsyenodiorite enclave			
#	87.1+	87.2c+	87.2i+	87.2r+	49B+	49B.1	49B.2c	49B.2r	67.1+	67.2c+	67.2i+	67.2r+	67.1	67.2	67.3c	67.3r
n	2	2	1	2	2	2	2	3	2	2	1	2	2	2	1	1
SiO ₂	46.97	47.61	47.69	47.58	49.07	49.37	47.68	48.60	47.94	48.41	48.17	48.36	48.47	49.98	50.65	50.22
TiO ₂	0.46	0.33	0.38	0.22	0.13	0.20	0.28	0.13	0.08	0.10	0.16	0.09	0.16	0.17	0.45	0.22
Al ₂ O ₃	0.57	0.44	0.39	0.35	0.38	0.28	0.90	0.45	0.12	0.17	0.26	0.15	0.29	0.66	1.10	0.62
Cr ₂ O ₃	n.d	n.d	n.d	n.d	-	0.03	-	0.00	n.d	n.d	n.d	n.d	n.d	n.d	n.d	n.d
Fe ₂ O ₃	1.15	1.03	0.84	1.65	1.19	0.19	1.23	1.07	0.84	1.45	1.78	1.10	1.06	0.60	1.21	0.27
FeO	29.48	29.58	30.49	29.56	26.81	27.96	27.55	25.75	27.52	26.85	26.99	27.06	38.39	20.89	16.26	21.45
MnO	0.81	0.66	0.79	0.65	0.69	0.78	0.70	0.74	0.85	0.77	0.66	0.74	0.89	0.50	0.48	0.47
NiO	n.d	n.d	n.d	n.d	0.04	0.03	-	-	n.d	n.d	n.d	n.d	n.d	n.d	n.d	n.d
MgO	0.26	0.24	0.52	0.25	2.16	2.01	2.02	2.65	0.66	1.18	1.31	1.18	9.43	7.50	11.05	7.74
CaO	18.91	19.08	18.40	18.93	19.08	18.92	17.39	18.97	20.42	20.07	19.79	20.33	1.52	18.31	18.14	18.06
Na ₂ O	0.34	0.45	0.34	0.46	0.62	0.56	0.47	0.48	0.28	0.54	0.52	0.41	0.01	0.26	0.27	0.30
K ₂ O	0.00	0.00	-	0.01	0.02	0.00	0.01	0.05	0.05	0.00	-	0.01	-	-	0.01	-
Total	98.95	99.43	99.85	99.66	100.20	100.33	98.24	98.89	98.77	99.55	99.64	99.44	100.24	98.87	99.62	99.35
ΣFeO	30.51	30.51	31.25	31.05	27.89	28.13	28.66	26.72	28.27	28.16	28.59	28.05	39.35	21.42	17.35	21.69
Structural formula on the basis of 6 oxygens																
IVSi	1.97	1.99	1.98	1.99	2.00	2.01	1.99	2.00	2.00	2.00	1.99	2.00	1.98	1.99	1.96	1.99
IVAl	0.03	0.02	0.02	0.01	0.00	-	0.01	0.01	0.00	0.00	0.01	0.00	0.01	0.01	0.04	0.01
T site	2.00	2.00	2.00	2.00	2.00	2.01	2.00	2.00	2.00	2.00	2.00	2.00	1.99	2.00	2.00	2.00
VIAl	0.00	0.01	0.00	0.00	0.02	0.01	0.03	0.02	0.00	0.01	0.00	0.01	-	0.02	0.01	0.02
Ti	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01
Cr	-	-	-	-	-	0.00	-	0.00	-	-	-	-	-	-	-	-
Fe ³⁺	0.04	0.03	0.03	0.05	0.04	0.01	0.04	0.03	0.03	0.04	0.06	0.03	0.03	0.02	0.04	0.01
Fe ²⁺	1.03	1.03	1.06	1.03	0.91	0.95	0.96	0.88	0.96	0.92	0.93	0.93	1.31	0.69	0.52	0.71
Mn ²⁺	0.03	0.02	0.03	0.02	0.02	0.03	0.02	0.03	0.03	0.03	0.02	0.03	0.03	0.02	0.02	0.02
Ni	-	-	-	-	0.00	0.00	-	-	-	-	-	-	-	-	-	-
Mg	0.02	0.01	0.03	0.02	0.13	0.12	0.12	0.16	0.04	0.07	0.08	0.07	0.57	0.44	0.64	0.46
Ca	0.85	0.85	0.82	0.84	0.83	0.82	0.77	0.83	0.91	0.89	0.87	0.90	0.07	0.78	0.75	0.77
Na	0.03	0.04	0.03	0.04	0.05	0.04	0.04	0.04	0.02	0.04	0.04	0.03	0.00	0.02	0.02	0.02
K	0.00	0.00	-	0.00	0.00	0.00	0.00	0.00	0.00	0.00	-	0.00	-	-	0.00	-
M1,M2	2.01	2.00	2.00	2.01	2.01	1.99	2.00	2.00	2.00	2.00	2.01	2.01	2.01	2.00	2.01	2.00
F/FM	0.99	0.99	0.97	0.99	0.88	0.89	0.89	0.85	0.96	0.93	0.92	0.93	0.70	0.62	0.47	0.61
Fs	54.41	54.32	55.44	54.47	48.66	50.14	51.57	47.00	50.16	49.11	49.36	49.01	67.17	36.20	27.47	36.74
Wo	44.72	44.90	42.87	44.70	44.35	43.44	41.69	44.39	47.69	47.04	46.37	47.16	3.42	40.65	39.26	39.63
En	0.87	0.78	1.69	0.83	6.98	6.43	6.74	8.61	2.15	3.86	4.27	3.82	29.41	23.15	33.27	23.63

Table 7.4: Results of electron-microprobe analyses of amphibole in enclaves and their host rocks: I (rhyolite porphyry, Buri ring dyke): 49A.1+ = ferroactinolite, 49A.2+ = silicic ferroedenite; **II (trachyandesite enclaves within rhyolite porphyry):** 49A.1 = ferroedenite, 49A.2 = ferroedenite, and 49B = ferroedenite; **III (granite porphyry, Buri ring dyke):** 67+ = ferroactinolite replacing clinopyroxene; **IV (microsyenodiorite enclave in granite porphyry):** 67 = ferrohornblende replacing clinopyroxene; **Enclaves in biotite granite: V (trachybasalt):** 73B: ferroedenite, **VI (andesite):** 73B1.1 = ferroedenite, 73B1.2 = ferroactinolitic hornblende, & 73B1.3 = ferrohornblende; **VII (trachydacite):-** 73C.1 = ferroedenite, 73C.2 = ferrohornblende, 73C.3 = ferroactinolitic hornblende, 73C.4c & 73C.4r (ferroactinolitic hornblende core and rim, respectively of a zoned grain).

Table 7.4:

	I		II			III	IV	V	VI			VII				
#	49A.1+	49A.2+	49A.1	49A.2	49B	67+	67	73B	73B1.1	73B1.2	73B1.3	73C.1	73C.2	73C.3	73C.4c	73C.4r
n	2	1	1	1	5	3	1	7	3	3	1	3	2	1	1	2
SiO2	49.59	48.72	45.54	46.36	44.39	47.46	43.55	43.72	44.63	47.57	47.02	43.58	46.28	49.24	49.02	45.95
TiO2	0.37	0.39	1.26	1.14	1.21	0.13	1.12	1.49	0.89	0.61	0.64	0.57	1.00	0.78	0.94	1.16
Al2O3	3.38	3.44	4.59	4.32	4.43	1.45	4.63	7.31	6.28	4.20	5.10	7.62	5.67	3.72	3.88	5.99
ΣFeO	24.23	25.48	24.73	25.78	27.04	34.01	30.21	23.56	23.49	22.47	23.36	21.28	20.72	20.41	20.44	20.55
MnO	0.57	0.60	0.37	0.43	0.54	0.70	0.47	0.46	0.59	0.61	0.62	0.39	0.45	0.48	0.42	0.41
MgO	7.54	7.09	7.81	7.28	5.40	1.09	4.11	6.90	7.42	8.64	8.08	8.12	8.96	10.90	10.72	9.34
ZnO	0.20	0.15	-	-	-	-	-	-	-	-	-	-	-	-	-	-
CaO	9.02	8.85	9.52	9.42	9.77	10.65	9.11	10.51	10.61	10.58	10.81	10.75	10.53	10.05	10.20	10.77
Na2O	1.93	2.38	2.47	2.30	2.41	0.36	2.30	1.53	1.35	1.06	1.08	1.41	1.33	1.27	1.30	1.55
K2O	0.76	0.80	0.93	0.91	0.99	0.25	0.89	1.02	1.04	0.71	0.83	1.15	0.77	0.41	0.39	0.78
H2O	1.54	1.26	0.89	1.25	1.13	1.78	1.14	1.60	1.22	1.23	1.32	1.24	1.32	1.27	1.30	1.30
F	0.80	1.39	2.02	1.30	1.46	0.64	1.30	0.48	1.35	1.43	1.25	1.24	1.19	1.45	1.40	1.25
Cl	0.13	0.10	0.26	0.23	0.18	0.09	0.35	0.33	0.17	0.11	0.12	0.26	0.20	0.10	0.09	0.21
O=F+Cl	-0.40	-0.60	-0.90	-0.60	-0.70	-0.40	-0.60	-0.20	-0.60	-0.60	-0.60	-0.60	-0.50	-0.60	-0.60	-0.60
Total	99.70	100.03	99.47	100.11	98.30	98.56	98.54	98.65	98.43	98.60	99.69	97.02	97.86	99.45	99.48	98.69
Structural formula on the basis of 24(O,OH,F,Cl)																
IVSi	7.61	7.53	7.13	7.22	7.14	7.79	7.09	6.87	7.02	7.37	7.24	6.90	7.18	7.44	7.41	7.08
IVAl	0.39	0.47	0.85	0.78	0.83	0.21	0.89	1.13	0.98	0.63	0.76	1.10	0.82	0.56	0.59	0.92
IVTi	-	-	0.02	-	0.03	-	0.02	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	0.22	0.15	-	0.01	0.01	0.06	-	0.22	0.18	0.14	0.17	0.32	0.22	0.10	0.10	0.17
Ti	0.04	0.04	0.13	0.13	0.11	0.02	0.11	0.18	0.10	0.07	0.07	0.07	0.12	0.09	0.11	0.14
Mg	1.73	1.63	1.82	1.69	1.29	0.27	1.00	1.61	1.74	2.00	1.86	1.91	2.07	2.45	2.42	2.14
Fe2+	3.01	3.17	3.05	3.17	3.57	4.65	3.89	2.97	2.97	2.80	2.90	2.70	2.59	2.35	2.38	2.55
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	0.02	-	-	-	-	-	-	-	-	-	-	-
Ca	-	-	-	-	0.00	-	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Fe2+	0.10	0.12	0.19	0.19	0.07	0.01	0.22	0.11	0.12	0.11	0.11	0.12	0.10	0.23	0.21	0.10
Zn	0.02	0.02	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	0.07	0.08	0.05	0.06	0.06	0.10	0.06	0.06	0.08	0.08	0.08	0.05	0.06	0.06	0.05	0.05
Ca	1.48	1.46	1.60	1.57	1.68	1.87	1.59	1.77	1.78	1.76	1.78	1.81	1.75	1.63	1.65	1.78
Na	0.32	0.32	0.17	0.18	0.18	0.03	0.12	0.06	0.03	0.05	0.02	0.02	0.09	0.09	0.09	0.07
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	-	-	0.00	-	-	0.01	-	-	0.02	-	-	-	-
Na	0.25	0.39	0.58	0.51	0.57	0.09	0.60	0.41	0.38	0.27	0.30	0.41	0.31	0.29	0.29	0.39
K	0.15	0.16	0.19	0.18	0.20	0.05	0.18	0.20	0.21	0.14	0.16	0.23	0.15	0.08	0.08	0.15
A site	0.40	0.55	0.77	0.69	0.77	0.14	0.78	0.61	0.60	0.41	0.46	0.66	0.46	0.36	0.37	0.55
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.58	1.29	0.93	1.30	1.21	1.74	1.23	1.67	1.28	1.27	1.36	1.31	1.36	1.28	1.31	1.34
F	0.39	0.68	1.00	0.64	0.74	0.12	0.67	0.24	0.67	0.70	0.61	0.62	0.58	0.69	0.67	0.61
Cl	0.03	0.03	0.07	0.06	0.05	0.14	0.10	0.09	0.05	0.03	0.03	0.07	0.05	0.02	0.02	0.05
F/FM	0.64	0.67	0.64	0.67	0.74	0.97	0.80	0.66	0.64	0.59	0.62	0.60	0.56	0.51	0.52	0.55

Table 7.5: Results of electron-microprobe analyses of mica in enclaves within biotite granite of the Panshanu center: I (trachybasalt): 73B; II (andesitic enclave): 73B1.1 and 73B1.2, 73B2.1 & 73B2.2; 73B1.3+ & 73B2.3 are mica in felsic enclaves within andesitic enclaves; III (trachydacite enclaves): 73C.1 & 2; IV (porphyritic biotite granite, Panshanu center): 1A1a+, 1A2a, 1B2a+, & 1B2b+ = dark mica, 1A1b+ & 1A2b+: pale coloured mica; V (medium-grained biotite granite, Panshanu center): 2A1a+ & 2B1a+ = dark mica, 2A1b+ & 2B1b+ = pale coloured mica; VI (albitized biotite granite, Panshanu center): 1C+, 74C+, 76A1+, & 76B+ = dark mica, 76A2 = pale coloured mica.

Table 7.5:

#	I		II				III				IV				V	
	73B	73B1.1	73B1.2	73B1.3+	73B2.1	73B2.2	73B2.3+	73C.1	73C.2	1A1a+	1A1b+	1A2a+	1A2b+	1B2a+	1B2b+	2A1a+
	6	5	1	2	5	1	2	3	1	10	1	13	1	8	6	9
SiO ₂	38.39	37.36	38.87	36.57	38.08	37.88	36.95	36.93	36.92	37.26	52.00	38.17	47.47	37.19	37.21	37.25
TiO ₂	1.84	3.23	1.91	3.11	3.56	3.46	3.80	3.71	4.14	1.79	0.01	1.05	0.34	2.30	2.12	1.66
Al ₂ O ₃	12.46	12.79	12.46	12.97	12.98	12.65	13.18	12.99	12.65	18.95	25.08	19.67	27.51	17.69	17.69	19.09
ΣFeO	23.36	22.17	23.59	27.49	22.42	22.91	26.15	26.21	26.28	26.56	6.72	23.89	8.56	24.93	27.34	25.07
MnO	0.28	0.22	0.24	0.31	0.18	0.20	0.32	0.18	0.20	0.52	0.03	0.49	0.16	0.34	0.34	0.52
MgO	10.59	10.30	10.60	6.63	10.01	10.67	7.56	7.49	7.35	0.88	1.18	1.00	0.40	1.64	1.66	1.10
CaO	0.06	0.01	0.07	0.00	0.02	0.03	0.00	0.03	0.02	0.02	0.34	0.01	-	0.01	0.01	0.01
Na ₂ O	0.09	0.10	0.11	0.05	0.06	0.09	0.08	0.06	0.10	0.12	0.02	0.17	0.14	0.10	0.10	0.15
K ₂ O	9.87	10.04	10.02	9.75	10.10	9.86	9.88	9.73	9.85	9.83	5.40	9.85	11.10	9.90	9.86	9.46
F	2.83	1.86	2.63	1.45	1.30	1.43	1.02	0.52	0.55	2.43	3.32	3.04	2.35	1.95	2.06	2.54
Cl	0.28	0.30	0.30	0.47	0.35	0.35	0.48	0.43	0.47	0.18	0.02	0.14	0.04	0.26	0.22	0.16
O=F+Cl*	-1.00	-0.90	-1.00	-0.70	-0.60	-0.70	-0.50	-0.30	-0.30	-1.00	-1.00	-1.00	-1.00	-0.90	-0.90	-1.00
Total	98.80	97.53	99.63	98.10	97.44	98.85	98.90	97.95	97.19	97.48	92.72	96.17	96.07	95.45	97.67	95.91
H ₂ O*	2.44	2.90	2.54	2.96	3.17	3.10	3.20	3.45	3.42	2.65	2.99	2.41	3.24	2.85	2.80	2.61
Structural formula on the basis of 24 (O,OH,F,Cl)																
IVSi	5.88	5.76	5.90	5.74	5.78	5.75	5.70	5.73	5.72	5.72	6.71	5.84	6.20	5.79	5.72	5.76
IVAl	2.12	2.24	2.10	2.26	2.01	2.25	2.30	2.27	2.28	2.28	1.29	2.16	1.80	2.21	2.28	2.24
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
IVAl	0.13	0.08	0.13	0.14	0.11	0.01	0.09	0.10	0.03	1.14	2.53	1.38	2.53	1.03	0.92	1.24
VITi	0.21	0.37	0.22	0.37	0.42	0.39	0.44	0.43	0.48	0.21	0.00	0.12	0.03	0.27	0.24	0.19
Fe ²⁺	2.99	2.86	2.99	3.61	2.85	2.91	3.37	3.40	3.41	3.41	0.73	3.06	0.96	3.24	3.51	3.24
Mn ²⁺	0.04	0.03	0.03	0.04	0.02	0.03	0.04	0.02	0.03	0.07	0.00	0.06	0.02	0.05	0.04	0.07
Mg	2.42	2.37	2.40	1.55	2.27	2.41	1.74	1.73	1.70	0.20	0.23	0.23	0.08	0.38	0.38	0.25
O site	5.79	5.70	5.77	5.71	5.16	5.75	5.68	5.68	5.65	5.03	3.49	4.85	3.62	4.97	5.10	4.99
Ca	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.00	0.00	0.00	0.00	0.00
Na	0.03	0.03	0.03	0.02	0.02	0.03	0.02	0.02	0.03	0.04	0.00	0.05	0.04	0.03	0.03	0.04
K	1.93	1.97	1.94	1.95	1.97	1.91	1.94	1.92	1.95	1.92	0.89	1.92	1.89	1.97	1.93	1.87
A site	1.97	2.00	1.98	1.97	1.98	1.94	1.97	1.95	1.98	1.96	0.94	1.97	1.93	2.00	1.96	1.91
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.55	3.02	2.66	3.15	3.28	3.22	3.38	3.63	3.61	2.77	2.64	2.49	3.00	2.97	2.94	2.71
F	1.37	0.90	1.26	0.72	0.62	0.69	0.50	0.26	0.27	1.18	1.36	1.47	0.99	0.96	1.00	1.24
Cl	0.07	0.08	0.08	0.13	0.56	0.09	0.13	0.11	0.12	0.05	0.00	0.04	0.01	0.07	0.06	0.04
F/FM	0.55	0.55	0.56	0.70	0.56	0.55	0.66	0.66	0.67	0.94	0.76	0.93	0.92	0.89	0.90	0.93
X(F)	0.34	0.23	0.32	0.18	0.16	0.17	0.12	0.06	0.07	0.30	0.34	0.37	0.25	0.24	0.25	0.31
X(Cl)	0.02	0.02	0.02	0.03	0.02	0.02	0.03	0.03	0.03	0.01	0.00	0.01	0.00	0.02	0.01	0.01
F/Cl	19.07	11.40	16.59	5.79	6.86	7.59	3.88	2.28	2.17	25.99	326.47	42.04	125.03	14.27	17.95	31.23
Phl	41.76	41.46	41.56	27.18	40.17	41.96	30.53	30.45	30.07	4.00	4.00	4.00	1.00	7.00	7.00	4.00
Srp	8.02	4.88	7.79	10.64	7.01	0.71	6.68	7.20	2.52	18.00	41.00	22.00	42.00	16.00	14.00	20.00
Ann	50.22	53.66	50.65	62.18	52.52	57.33	62.79	62.35	67.40	78.00	56.00	74.00	57.00	77.00	79.00	76.00

V: cont'd

VI

	2A1b+ 2B1a+		2B1b+ 1C+		74C+ 76A1+		76A2+ 76B+	
n	1	15	1	11	4	7	1	6
SiO2	46.84	37.50	47.63	37.14	37.64	35.99	35.44	36.90
TiO2	0.14	2.00	0.23	1.60	1.26	2.36	2.27	1.67
Al2O3	26.71	18.30	27.21	18.76	19.22	18.37	18.12	19.30
ΣFeO	9.10	24.77	6.61	26.12	24.76	27.87	28.46	26.75
MnO	0.18	0.50	0.19	0.53	0.69	0.52	0.57	0.62
MgO	0.47	1.25	0.63	0.63	1.07	0.82	0.74	0.30
CaO	0.00	0.01	-	0.01	0.00	0.00	-	0.01
Na2O	0.12	0.13	0.20	0.16	0.20	0.12	0.08	0.13
K2O	10.67	9.21	9.95	9.73	9.56	9.85	10.11	9.89
F	2.35	2.59	2.42	3.13	3.69	2.27	2.35	2.54
Cl	0.01	0.18	0.00	0.14	0.08	0.17	0.13	0.11
O=(F+Cl)	-1.00	-1.00	-1.00	-1.00	-2.00	-1.00	-1.00	-1.00
Total	95.64	95.32	94.05	96.60	96.61	97.35	97.24	97.12
H2O*	3.24	2.59	3.27	2.31	2.21	2.67	2.84	2.60

Structural formula on the basis of 24 (O,OH,F,Cl)

Si	6.27	5.82	6.34	5.77	5.80	5.62	4.58	5.70
IVAl	1.73	2.18	1.66	2.23	2.20	2.38	2.76	2.30
IVTi	-	-	-	-	0.00	-	0.66	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	2.49	1.16	2.62	1.20	1.05	1.00	-	1.22
VITi	0.01	0.23	0.02	0.19	0.10	0.28	1.55	0.19
Fe2+	1.02	3.21	0.74	3.39	3.23	3.64	3.08	3.46
Mn2+	0.02	0.07	0.02	0.07	0.19	0.07	0.06	0.08
Mg	0.09	0.29	0.12	0.15	0.26	0.19	0.14	0.07
O site	3.64	4.97	5.52	4.99	4.83	5.17	4.83	5.03
Ca	0.00	0.00	-	0.00	0.00	0.00	-	0.00
Na	0.03	0.04	0.05	0.05	0.05	0.04	0.02	0.04
K	1.82	1.82	1.69	1.93	1.46	1.96	1.67	1.95
A site	1.85	1.86	1.74	1.98	1.51	2.00	1.69	1.99
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.00	2.68	2.98	2.43	2.55	2.83	3.01	2.73
F	1.00	1.27	1.02	1.54	1.43	1.12	0.96	1.24
Cl	0.00	0.05	0.00	0.04	0.02	0.04	0.03	0.03
F/YM	0.92	0.92	0.86	0.96	0.93	0.95	0.96	0.98
X(F)	0.25	0.32	0.25	0.38	0.36	0.28	0.24	0.31
X(Cl)	0.00	0.01	0.00	0.01	0.00	0.01	0.01	0.01
F/Cl	1038.3	30.03	902.82	45.00	90.15	27.15	33.25	45.61
Phl	2.00	5.00	2.00	3.00	11.00	3.00	3.00	1.00
Srp	41.00	18.00	43.00	20.00	16.00	16.00	-	20.00
Ann	58.00	76.00	55.00	78.00	73.00	81.00	97.00	79.00

Table 7.6: Whole-rock chemical data for enclaves and their felsic host rocks

	<u>49M</u>	<u>87M</u>	<u>67M</u>	<u>73</u>	<u>73B2</u>	<u>73C</u>	<u>73M</u>	<u>20</u>	<u>44A</u>	<u>KW6</u>
Major elements										
SiO ₂	56.35	59.25	57.26	48.83	59.47	65.53	64.91	46.33	47.08	59.70
TiO ₂	1.96	1.72	1.47	2.44	1.70	1.05	1.28	3.55	3.49	0.96
Al ₂ O ₃	13.30	12.88	15.64	16.29	15.19	14.25	14.31	15.06	15.18	15.20
Fe ₂ O ₃	12.57	11.17	10.29	13.19	9.29	2.15	7.67	4.02	4.63	1.83
FeO	n.d.	n.d.	n.d.	n.d.	n.d.	4.10	n.d.	10.64	8.48	6.74
MnO	0.23	0.18	0.26	0.19	0.13	0.09	0.10	0.33	0.24	0.17
MgO	2.06	1.77	1.35	4.35	3.13	1.06	1.44	5.32	5.23	1.26
CaO	3.65	4.05	3.24	5.73	5.16	2.80	3.02	7.45	10.48	4.11
Na ₂ O	4.08	4.58	4.92	3.33	4.03	3.41	4.51	2.76	2.23	4.63
K ₂ O	3.83	3.56	4.12	3.36	1.85	4.27	2.39	1.56	1.91	3.97
P ₂ O ₅	1.29	1.14	0.62	0.47	0.33	0.29	0.36	0.49	0.60	0.47
LOI	1.29	0.00	1.01	2.06	0.46	0.37	0.58	2.26	1.06	0.89
Trace elements										
Cr	0	0	0	51	17	10	0	38	118	n.d.
Ni	0	0	0	49	23	15	0	75	63	n.d.
V	63	46	84	231	149	57	92	263	293	15
Pb	32	22	98	9	7	12	11	6	0	17
S	0	0	0	0	0	400	n.d.	44720	800	n.d.
Rb	169	137	134	441	141	139	298	179	127	50
Ba	854	701	1122	351	358	803	483	349	286	1108
Sr	230	236	184	144	254	181	115	414	580	300
Ta	2.45	4.58	2.19	1.97	1.29	2.8	n.d.	1.25	n.d.	n.d.
Nb	117	123	60	60	21	35	44	25	26	43
Hf	18.93	18.95	18.11	5.62	5.76	10.69	n.d.	6.11	n.d.	11
Zr	792	777	836	226	223	454	416	246	222	617
Y	146	130	76	99	48	53	100	34	28	51
Th	9	17	9.25	15	13	15	14	6	5	7
U	3	4	0	9	9	13	5	16	12	n.d.
Rare earth elements (REE)										
La	96.92	95.68	86.49	42.60	32.59	90.92	n.d.	24.68	n.d.	131
Ce	212.39	214.49	182.17	100.26	73.02	153.89	n.d.	57.03	n.d.	202
Pr	27.16	26.52	22.58	11.26	8.70	17.82	n.d.	7.48	n.d.	n.d.
Nd	111.32	106.70	92	42.34	33.81	64.24	n.d.	31.81	n.d.	63.80
Sm	26.05	25.25	18.02	9.16	7.45	11.61	n.d.	7.45	n.d.	18.90
Eu	4.31	4.25	3.80	1.88	1.53	2.27	n.d.	2.61	n.d.	4.84
Gd	27.29	26.28	18.08	10.50	7.46	10.54	n.d.	7.88	n.d.	13.10
Tb	4.00	3.81	2.41	1.73	1.05	1.58	n.d.	1.07	n.d.	n.d.
Dy	26.83	25.99	15.13	13.63	7.11	9.80	n.d.	6.88	n.d.	2
Ho	5.11	4.68	2.84	2.90	1.45	1.82	n.d.	1.28	n.d.	2
Er	14.15	13.04	8	9.94	4.39	5.42	n.d.	3.64	n.d.	n.d.
Tm	1.98	1.76	1.10	1.61	0.65	0.75	n.d.	0.49	n.d.	1
Yb	11.84	10.71	6.79	10.41	4.37	4.55	n.d.	3.04	n.d.	4.79
Lu	1.61	1.48	1.01	1.58	0.60	0.71	n.d.	0.47	n.d.	0.86
Normative composition										
Qtz	11.22	12.61	6.30	0.29	15.28	20.84	22.01	2.98	0.11	5.53
Cor	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Zrn	0.16	0.16	0.17	0.05	0.04	0.09	0.08	0.05	0.04	0.12
Or	22.70	21.09	24.40	20.03	10.99	25.29	14.24	9.29	11.34	23.48
Ab	34.52	38.75	41.63	28.17	34.10	28.85	38.16	23.35	18.87	39.17
An	6.67	4.07	8.45	19.53	17.90	11.00	11.73	24.10	25.78	9.00
Ac	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Dp	0.00	2.75	0.00	0.00	0.25	1.07	0.00	7.82	18.27	7.48
Hyp	5.13	3.14	3.36	10.83	7.68	6.21	3.59	11.36	10.46	8.91
Mag	0.00	0.00	0.00	0.00	0.00	3.12	0.00	5.83	6.71	2.65
Chr	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.03	0.00
Hem	12.57	11.17	10.29	13.19	9.29	0.00	7.67	0.00	0.00	0.00
Ilm	0.49	0.38	0.55	0.41	0.28	1.99	0.21	6.74	6.63	1.82
Ttn	2.33	3.74	2.77	4.21	3.82	0.00	0.75	0.00	0.00	0.00
Rut	0.76	0.00	0.06	0.51	0.00	0.00	0.86	0.00	0.00	0.00
Apat	3.09	2.72	1.49	1.12	0.78	0.70	0.86	1.16	1.43	1.13
Pyr	0.00	0.00	0.00	0.00	0.00	0.07	0.00	8.36	0.15	0.00

N.B.: Semiquantitative SEM - EDS analyses of zoned trachyandesite enclave shows the following compositional range: Core (Rim in brackets) - SiO₂ = 54.56 (57.66-58.36), TiO₂ = 2.47 (2.33-2.38), Al₂O₃ = 13.46 (12.27-13.17), FeO = 15.64 (13.22-13.46), MnO = 0.29 (0.30-0.34), ZnO = 0.07 (0.06-0.17), MgO = 1.36 (0.83-0.87), CaO = 5.14 (4.13-5.48), Na₂O = 4.05 (4.08-4.11), and K₂O = 2.98 (3.33 - 3.43).

	KM15	NG9	NG18	49F	87F	43	65	1A	1B	2A
Major elements										
SiO ₂	65.50	71.90	71.10	73.70	74.10	76.29	71.58	77.20	75.79	75.89
TiO ₂	0.56	0.37	0.58	0.36	0.35	0.25	0.31	0.10	0.13	0.10
Al ₂ O ₃	14.24	12.24	12.56	12.45	12.28	12.98	13.29	12.80	13.22	12.72
Fe ₂ O ₃	2.47	1.83	2.56	3.56	3.32	1.33	3.16	0.64	1.40	0.99
FeO	3.71	1.61	0.92	n.d.	n.d.	1.62	0.42	0.46	0.00	0.38
MnO	0.20	0.07	0.08	0.06	0.04	0.07	0.06	0.02	0.01	0.02
MgO	0.47	0.12	0.08	0.21	0.12	0.00	0.07	0.00	0.00	0.00
CaO	1.58	0.87	1.09	0.85	0.84	0.72	0.49	0.61	0.68	0.59
Na ₂ O	4.09	4.00	3.66	3.91	3.90	3.86	3.92	3.73	3.78	3.73
K ₂ O	5.33	5.69	5.69	4.85	4.83	5.11	5.34	4.72	4.43	4.67
P ₂ O ₅	0.12	0.05	0.03	0.16	0.13	0.02	0.03	0.01	0.02	0.01
LOI	0.00	0.00	0.00	0.46	0.24	0.38	0.74	0.45	0.56	0.52
Trace elements										
Cr	n.d.	n.d.	n.d.	25	25	0	0	0	50	29
Ni	n.d.	n.d.	n.d.	0	0	12	17	31	170	24
V	2	9	0	13	0	0	12	0	10	0
Pb	19	26	35	21	22	17	9	8	19	16
S	n.d.	n.d.	n.d.	0	0	0	8720	0	400	400
Rb	78	115	134	166	182	152	121	404	342	387
Ba	476	339	450	351	192	273	315	94	205	81
Sr	76	58	54	36	26	19	24	16	29	17
Ta	n.d.	n.d.	n.d.	9.09	n.d.	6.01	n.d.	n.d.	10.18	n.d.
Nb	97	63	72	145	133	115	116	54	61.18	68
Hf	16	13	15	11	n.d.	11.23	n.d.	n.d.	3.91	n.d.
Zr	732	538	662	526	524	506	622	103	160	108
Y	94	72	78	145	155	130	146	98	86	103
Th	8	27	26	24	14	19	21	46	60.02	47
U	n.d.	n.d.	n.d.	0	2	23.31	13	28	18.02	25
Rare earth elements										
La	257	133	128	102.37	n.d.	161.42	n.d.	n.d.	65.52	n.d.
Ce	450	245	231	221.16	n.d.	325.95	n.d.	n.d.	138.71	n.d.
Pr	n.d.	0	n.d.	25.66	n.d.	38.54	n.d.	n.d.	14.82	n.d.
Nd	304	124	143	96.15	n.d.	140.96	n.d.	n.d.	48.58	n.d.
Sm	66.50	22	24	24.16	n.d.	30.94	n.d.	n.d.	10.17	n.d.
Eu	7.10	1.20	2.2	1.03	n.d.	1.12	n.d.	n.d.	0.48	n.d.
Gd	54.80	12	19	24.55	n.d.	30.01	n.d.	n.d.	9.31	n.d.
Tb	n.d.	n.d.	n.d.	3.87	n.d.	4.19	n.d.	n.d.	1.64	n.d.
Dy	6	4	4	26.52	n.d.	26.86	n.d.	n.d.	12.48	n.d.
Ho	6	4	4	4.81	n.d.	4.71	n.d.	n.d.	2.77	n.d.
Er	n.d.	n.d.	n.d.	13.49	n.d.	12.28	n.d.	n.d.	9.10	n.d.
Tm	2	1	1	1.88	n.d.	1.66	n.d.	n.d.	1.50	n.d.
Yb	14.40	5	6	10.93	n.d.	10.41	n.d.	n.d.	10.05	n.d.
Lu	2.55	1	1	1.57	n.d.	1.47	n.d.	n.d.	1.45	n.d.
Normative composition										
Qtz	16.05	25.59	26.47	30.86	31.58	28.23	27.17	35.95	35.34	34.98
Cor	0.00	0.00	0.00	0.00	0.00	0.00	0.21	0.44	0.98	0.45
Zrn	0.15	0.11	0.13	0.11	0.11	0.10	0.13	0.02	0.03	0.02
Or	31.53	33.67	33.68	28.73	28.62	30.73	31.61	28.05	26.31	27.75
Ab	34.61	31.25	30.97	33.08	33.00	32.66	33.17	31.56	31.98	31.56
An	4.77	0.00	1.02	2.09	1.72	2.76	2.33	2.99	3.31	2.89
Ac	0.00	2.29	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Dp	2.10	3.57	0.43	0.12	0.46	0.68	0.00	0.00	0.00	0.00
Hyp	4.31	0.07	1.59	0.47	0.09	1.23	0.17	0.19	0.00	0.00
Mag	3.58	1.51	1.54	0.00	0.00	1.93	0.00	0.93	0.00	0.86
Chr	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.01
Hem	0.00	0.00	1.49	3.56	3.32	0.00	3.16	0.00	1.40	0.40
Ilm	1.06	0.70	1.10	0.12	0.08	0.47	0.00	0.19	0.00	0.19
Ttn	0.00	0.00	0.00	0.73	0.76	0.00	0.00	0.00	0.00	0.00
Rut	0.00	0.00	0.00	0.00	0.00	0.00	0.86	0.00	0.15	0.00
Apat	0.29	0.12	0.07	0.39	0.31	0.05	0.07	0.02	0.05	0.02
Pyr	0.00	0.00	0.00	0.00	0.00	0.00	1.63	0.00	0.08	0.07

The detection limits are: 0.01% for major elements, 15 ppm for Cr₂O₃, 10 ppm for V, Ni, and BaO, 4 ppm for Pb, and 2 ppm for Nb, Zr, Y, Sr, Rb, Th, and U. Total iron was analyzed as Fe₂O₃.

Chapter 8: Petrogenetic model of evolution of the subvolcanic felsic rocks of the Kwandonkaya complex.

8.1 Introduction

The first part of this presentation includes a review of the critical petrographic, mineral chemical and petrochemical features that have been described in earlier chapters, followed by the presentation of a model for the magmatic evolution for the felsic subvolcanic rocks of the Kwandonkaya complex. The subvolcanic rocks are chosen for modelling studies because: a) they preserve assemblages of the pristine magmatic minerals involved in crystal fractionation, and a devitrified microcrystalline groundmass representative of residual melt; b) these units have experienced the least extensive subsolidus modification, and hence minimal mobilization of elements, i.e., they can be assumed to have been closed systems, and c) they cover a range of bulk compositions that is more extensive than the plutonic suite (Chapter 2). In the next part of this presentation, previously proposed models for the origin of the felsic rocks in A-type granite provinces are evaluated for their applicability to the Kwandonkaya complex and the origin of the least evolved felsic unit.

The results of this study indicate that fractional crystallization can account for the observed variation in trace element concentrations in the subvolcanic rocks. Furthermore, partial melting of global average compositions of the lower crust, granodiorite, and granulite are incapable of producing the trace element patterns of the inferred parental magma for the felsic rock units. It is shown that selective enrichment of trace elements in realistic compositions of granulite and average lower crust is required to yield source-rock compositions capable of producing the parental melt for the Kwandonkaya complex and, by implication, other centers of A-type magmatism.

8.2 Constraints on the evolution of subvolcanic rocks at Kwandonkaya

The petrographic, mineralogical, and petrochemical observations and inferences, presented in the previous chapters, indicate that the subvolcanic rocks of the Kwandonkaya complex constitute a comagmatic suite related by crystal fractionation. The metaluminous to peraluminous rhyolite (+ biotite granites) and peralkaline granites of the NYG province plot near the "granite minimum" along the quartz - alkali feldspar cotectic (Fig. 8.1a) (cf. Bennett 1981, Bennett et al. 1984). Also, the subvolcanic rocks of the Kwandonkaya complex (Fig. 8.1b) all plot along the low-temperature trough, along with quartz-bearing syenites of the NYG province (Fig. 8.1a), extending towards the alkali feldspar join. These features indicate that the subvolcanic rocks evolved by low-pressure crystal fractionation [e.g., from (1) to (2) in Fig. 8.1b].

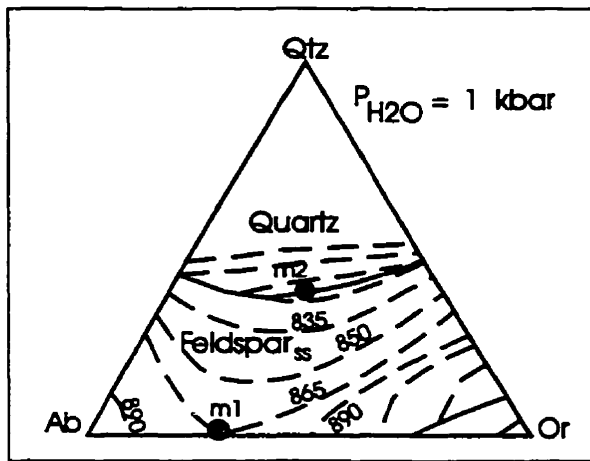
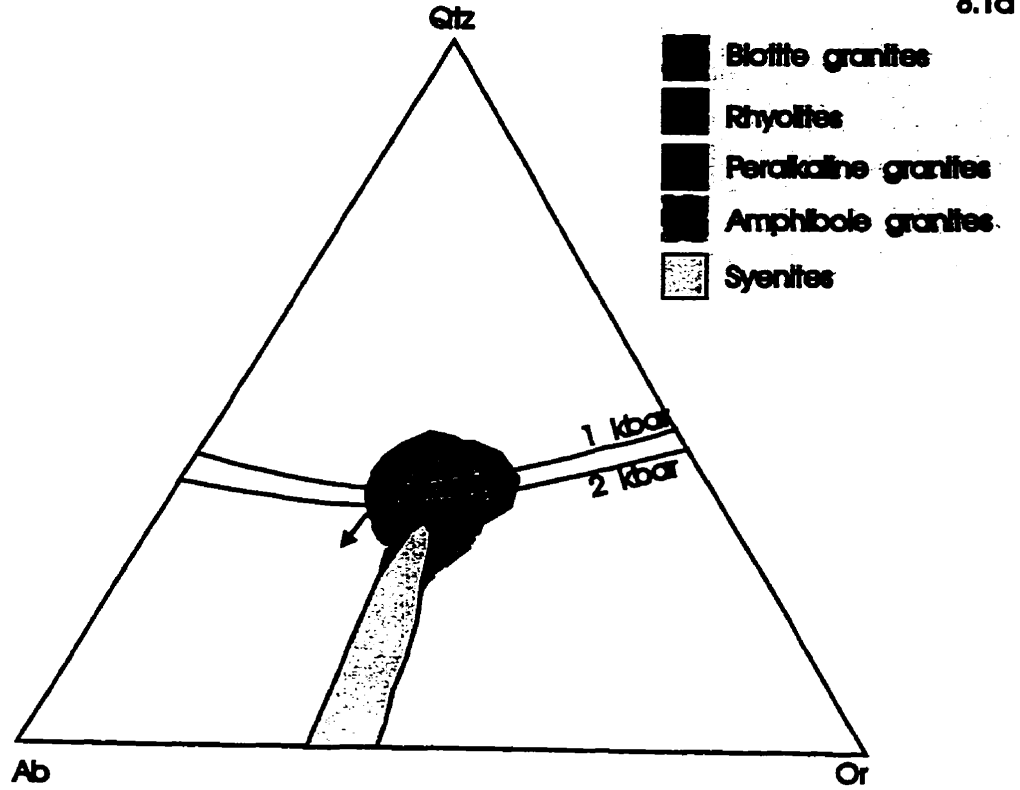
An overlap in some trace element ratios (Figs. 8.2a - h, and Table 8.2) further suggests that the subvolcanic rocks (Buri ring dyke, cone sheets, and Jubga-Ziem ring dyke) of the Kwandonkaya complex are comagmatic. The increase in magnitude of the negative Eu anomaly with concentrations of Rb (Fig. 8.2i, l) is consistent with fractionation of sodic plagioclase, and the HREE enrichment in the more evolved samples (documented in Chapter 2) requires the fractionation of LREE-enriched accessory phases (e.g., chevkinite and allanite).

A systematic decrease in CaO, TiO_2 , ΣFeO , Ba, Sr, Zr, ΣREE (Fig. 8.2j,.) and $(\text{La/Lu})_N$ (Fig. 8.2k) with increasing proportions of SiO_2 , Th, Rb, and U (Chapter 2) suggests the fractional crystallization of alkali feldspars, clinopyroxene, fayalite $\pm$ (zircon, Fe-Ti oxides, chevkinite, allanite) from parental metaluminous felsic magma(s). Furthermore, the systematic decrease in TiO_2 , ΣFeO and $^{\text{IV}}\text{Al}$ of mica with increasing SiO_2 , ΣAl , $^{\text{VI}}\text{Al}$, and F/Cl indicates evolution by fractional crystallization.

Fig. 8.1: Triangular plots of normative Qtz-Ab-Or for NYG complexes (a) and the Kwandonkaya subvolcanic rocks (b). a) Province-wide data on NYG granites, syenites, and rhyolites (MacLeod et al. 1971, Bennett 1981, Bennett et al. 1984, Bowden et al. 1987). The data points for metaluminous to peraluminous granites, rhyolites, and peralkaline granites plot near the granite minimum and are displaced from it along the low-melting-point trough on the liquidus between the pseudoternary minimum and the pseudobinary minimum on the feldspar join. The syenites of the NYG represent liquid compositions that have evolved from the albite-orthoclase join toward the minimum in silica-saturated part of the residua system. The arrow indicates the expected trend of displacement of minimum melt(s) towards the albite apex due to build-up in F content (Manning 1981, Manning & Pichavant 1988). b) Plot of Kwandonkaya data indicating that the subvolcanic rocks span the range from quartz syenite to the granite at the minimum. The cluster of data points is consistent with a fractionation sequence at a confining pressure less than 1 kbar, with (1) being more primitive than (2) in terms of LREE, Ba, Zr, Sr, and K/Rb. Melt compositions responsible for syenitic rocks more primitive than (1) (cf. Fig. 8.1a) may have been present at Kwandonkaya at deeper levels than that currently exposed.

Inset shows a liquidus phase diagram of the granite's residua system (i.e., the silica-saturated, Qtz-Ab-Or segment) modified from Landoll & Foland (1996) and Platt (1996). **m1** represents the minimum on the critically silica-saturated albite-orthoclase join and **m2** is the granite minimum towards which all melts evolve from the silica-saturated field.

8.1a



8.1b

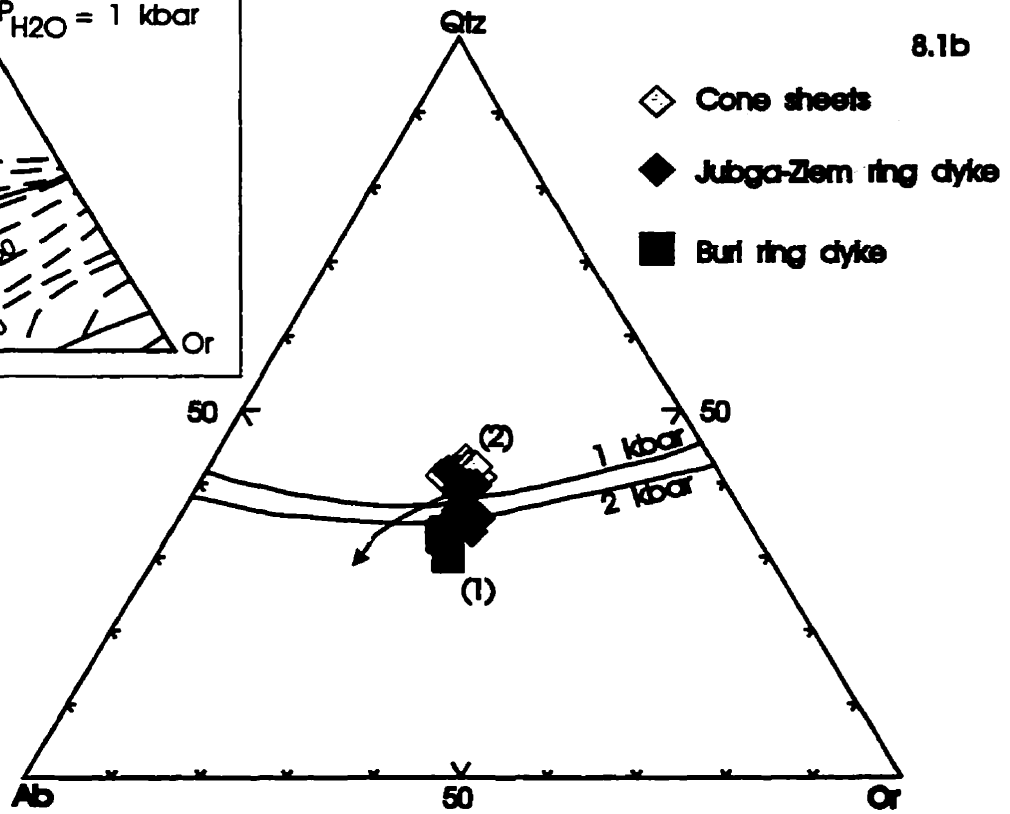
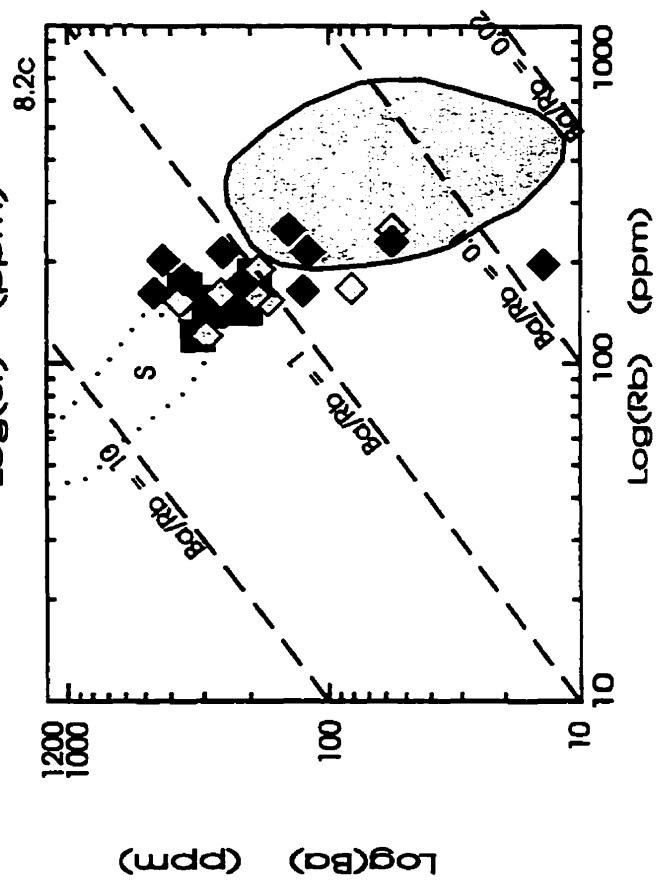
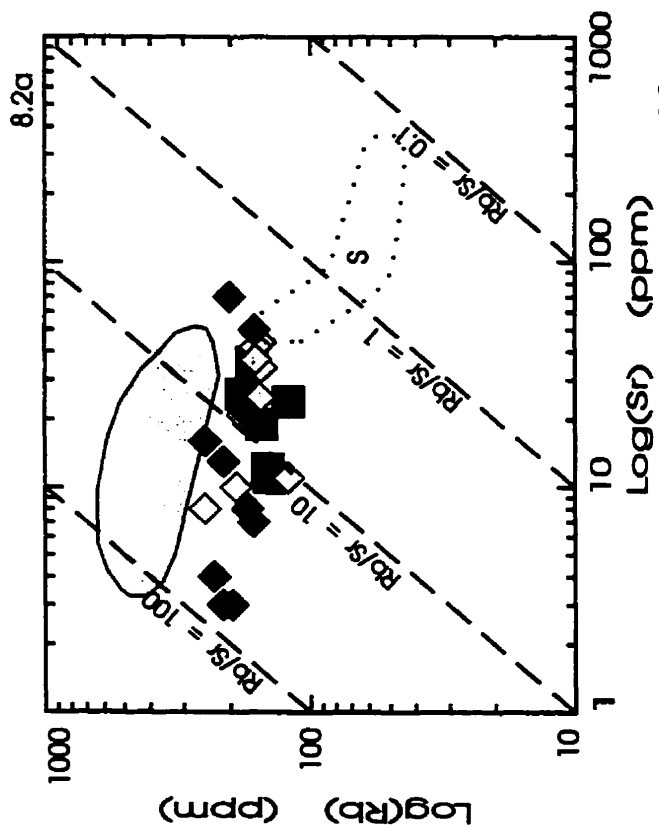
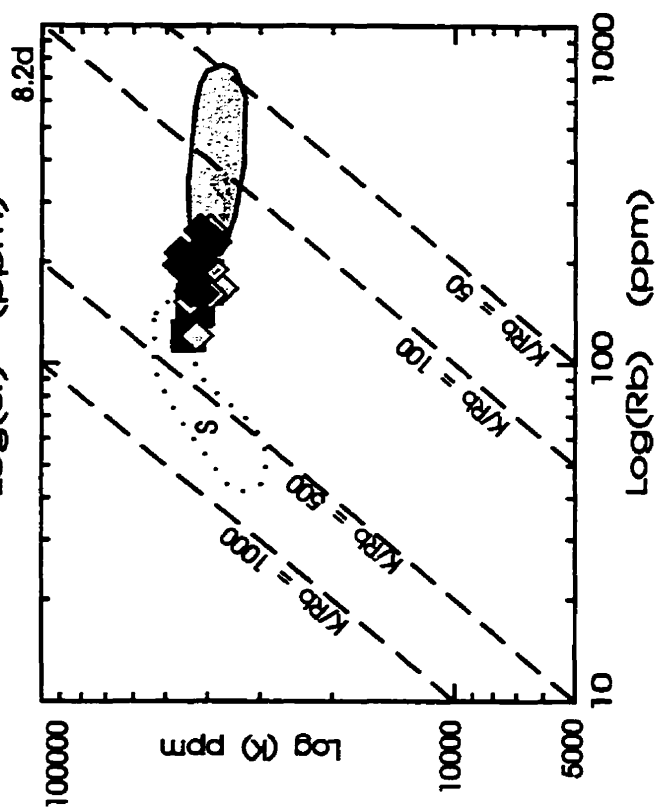
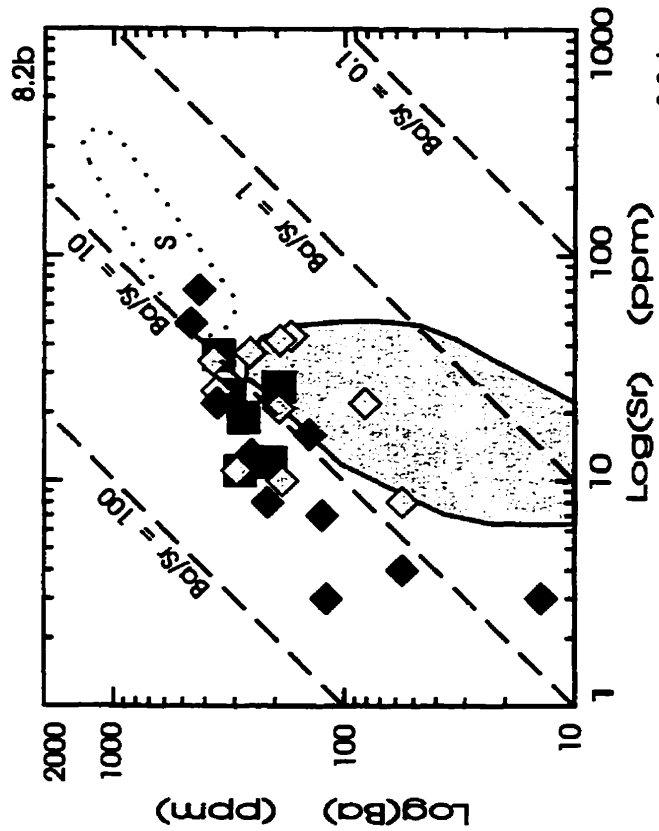


Fig. 8.2a - h. Log-Log trace-element plots, with ratios of concentrations indicated: **a)** Rb vs. Sr, **b)** Ba vs. Sr, **c)** Ba vs. Rb, **d)** K vs. Rb, **e)** K vs. Zr, **f)** Rb vs. Zr, **g)** Y vs. Nb, **h)** Nb vs. Ta . Although biotite granites are not part of the modelling, their range of compositions is plotted for comparative purposes (grey field). Similarly, the range of compositions of syenite from Kila-Warji and Ningi-Burra complexe (S) is plotted to indicate possible relation to subvolcanic rocks. The pristine samples of biotite granite are characterized by higher Rb/Sr, K/Zr, and Rb/Zr, as well as lower Ba/Sr, Ba/Rb, K/Rb and Nb/Ta, with similar Y/Nb compared to subvolcanic rocks (cf. Table 8.2).

Symbols are the same as used in Figure 8.1b.



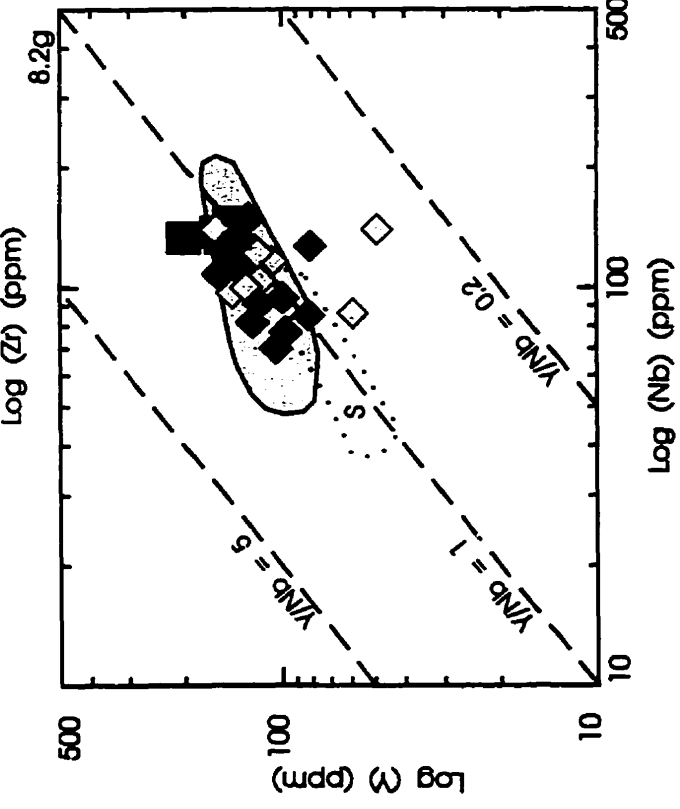
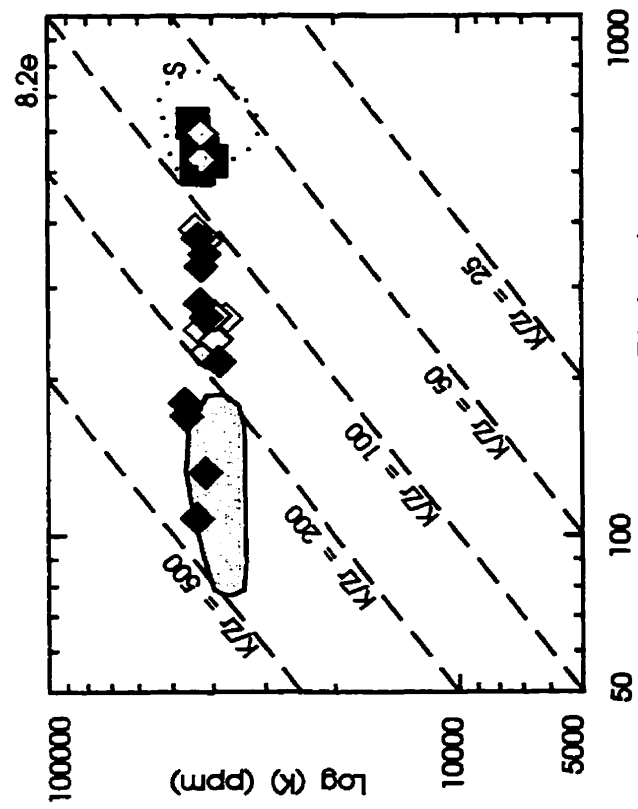
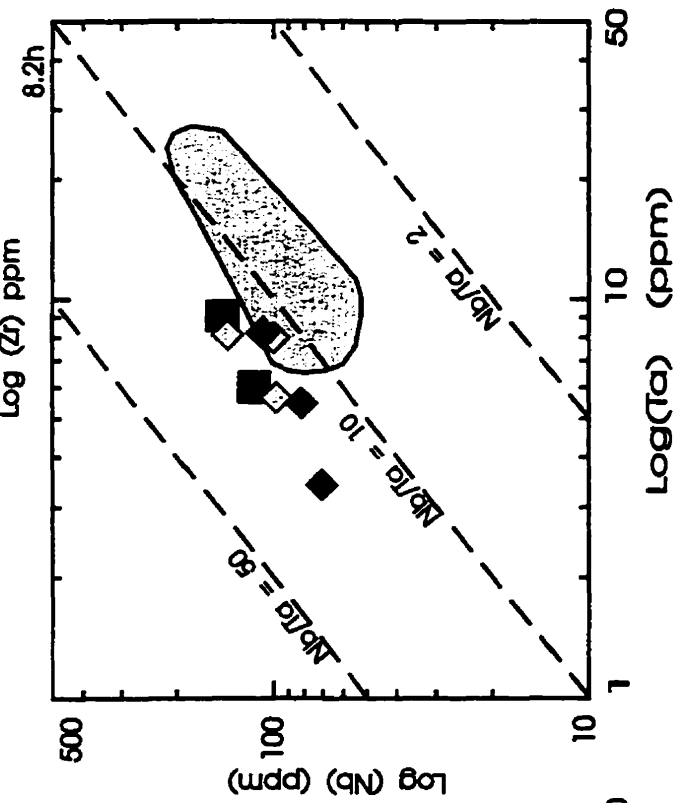
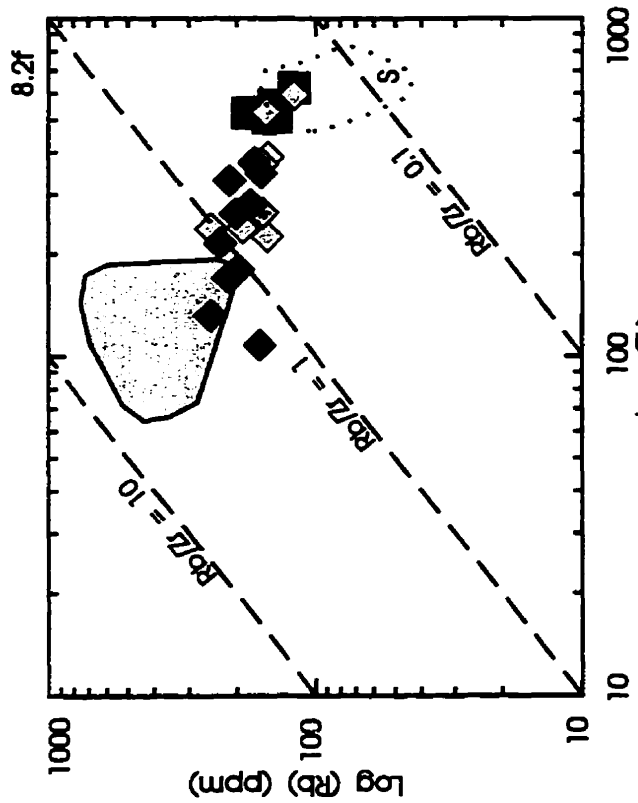


Fig. 8.2i-l: Plot of chondrite-normalized REE, and selected REE ratios that have trends which are consistent with fractional crystallization. i) chondrite-normalized REE plot, j) ΣREE vs. Rb, k) $(\text{La/Lu})_N$ vs. Rb, and l) $(\text{Eu/Eu}^*)_N$ vs. Rb. The ratio $(\text{Eu/Eu}^*)_N$ is the ratio of the Eu-anomaly (Eu) to projected europium content of samples assuming a Eu content collinear with normalized Sm and Gd concentrations (i.e., no anomaly). The variations indicate an increase in the magnitude of Eu anomalies with Rb, which is consistent with fractional crystallization of sodic plagioclase.

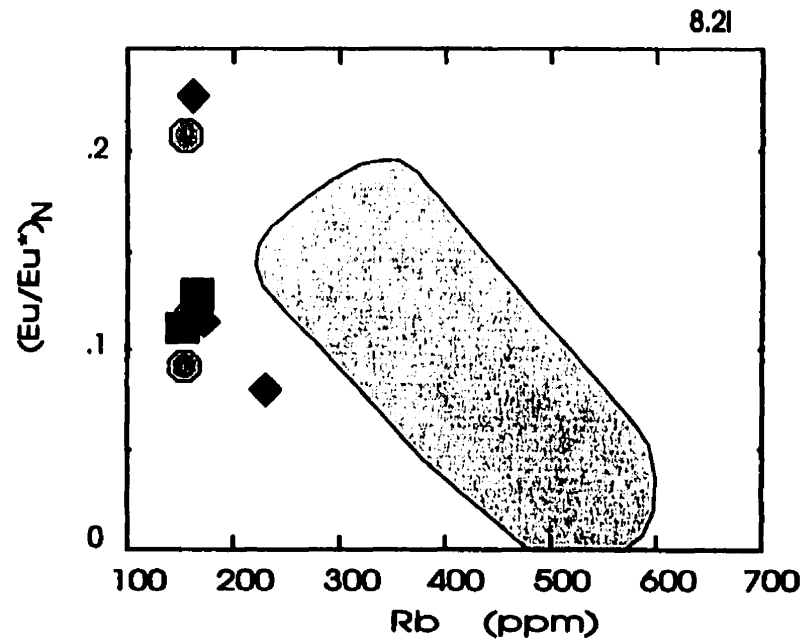
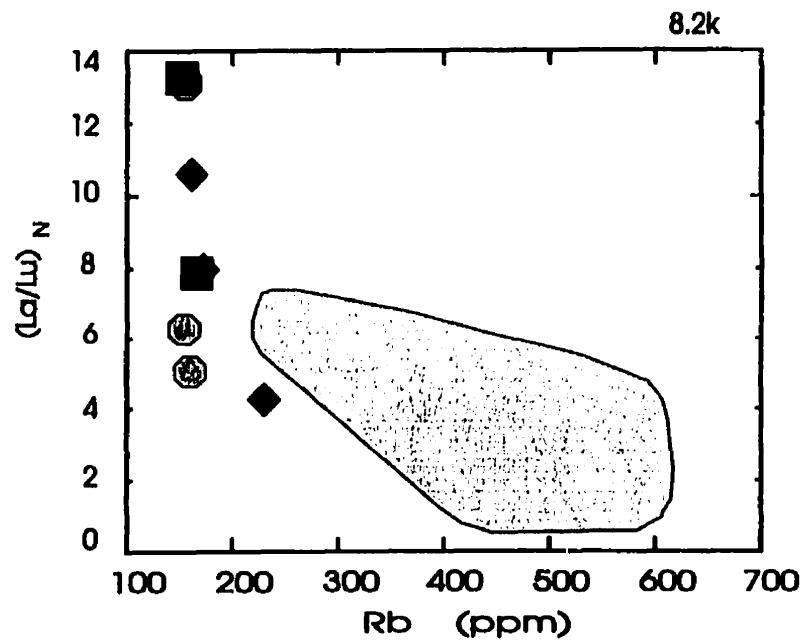
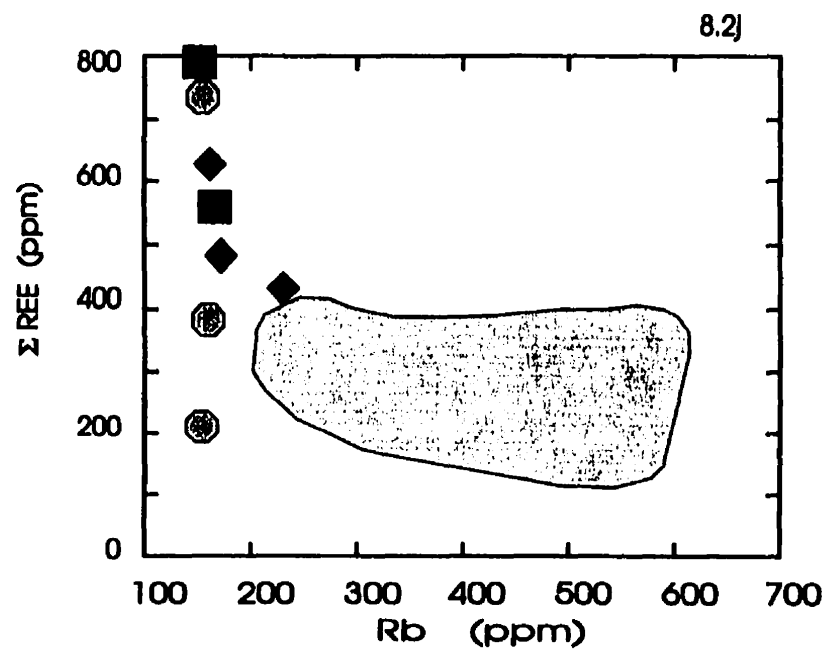
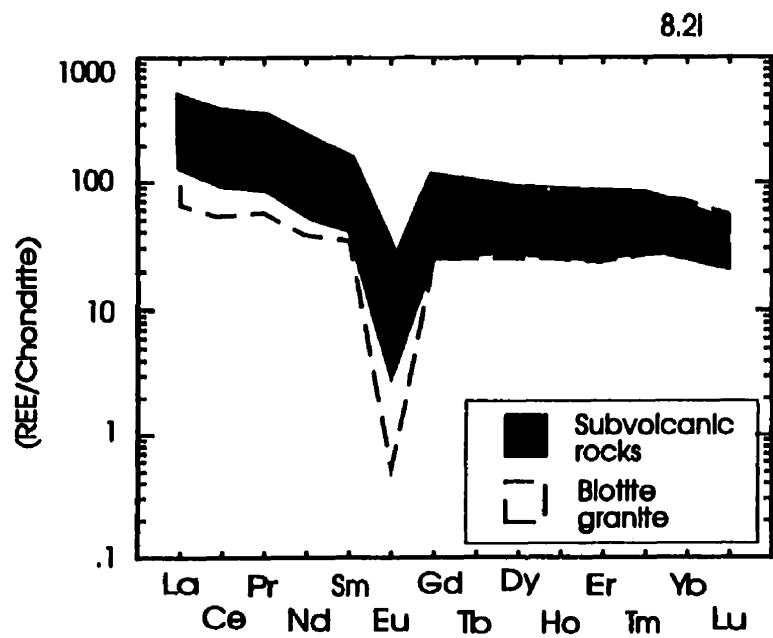


Table 8.1: Trace-element ratios (along with DI, ASI, and AI) for subvolcanic rocks and pristine biotite granites of the Kwandonkaya complex.

Element ratio	Buri ring dyke	Cone Sheets	Jubga-Ziem ring dyke	Pristine biotite granite
Rb/Sr	5 - 13	4 - 31	3 - 71 (1 - 50)	5 - 150
Ba/Sr	7 - 26	4 - 27	5 - 40 (4 - 30)	0.5 - 10
Ba/Rb	1 - 3	0.2 - 3	0.1 - 2	0.03 - 0.8
K/Rb	200 - 300	140 - 300	140 - 250 (100 - 300)	50 - 167
K/Zr	63 - 100	63 - 200	100 - 400	200 - 500
Rb/Zr	0.2 - 0.5	0.2 - 1.0	0.3 - 3	1 - 8
Y/Nb	1 - 2	0.3 - 2	0.5 - 2	0.7 - 3
Nb/Ta	16 - 19	13 - 17	13 - 21	5 - 12
Th/U	0.8 - 7	1 - 3	0.7 - 3.4	0.7 - 3
Zr/Hf	45 - 49	30 - 74	23 - 38	23 - 41
(La/Lu) _N	8 - 13	5 - 13	4 - 10	1 - 6
Eu/Eu*	0.1 - 0.13	0.09 - 0.21	0.08 - 0.23	0.02 - 0.18
ΣREE	558 - 791	211 - 735	432 - 627	169 - 393
DI	93 - 95	92 - 96	93 - 96	93 - 96
ASI	0.98 - 1.07	0.98 - 1.10	0.99 - 1.10	1.04 - 1.10 (up to 1.24)
AI	0.89 - 0.93	0.86 - 0.92	0.84 - 0.93	0.75 - 0.89

Symbols: DI = differentiation index [Σ normative(quartz + orthoclase + albite)], ASI = aluminum saturation index [i.e., molar ratio $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$], and AI = agpaitic index or alkali alumina ratio [i.e., molar ratio $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{Al}_2\text{O}_3$]. The ratios in brackets for Jubga-Ziem ring dyke are from Imeokparia (1984).

8.3 Petrogenetic modelling

In this section, least-squares mixing calculations (LSMC) are used to relate the least and most evolved rock compositions in each subvolcanic intrusive cycle by crystal fractionation. This is followed by a model for the subvolcanic suite as a whole. The basic assumptions made in the fractionation model are that the partition coefficients (K_d) did not change over the compositional spectrum of the subvolcanic suite, and that the fractionating minerals did not change in composition. Although these are clearly over-simplifications, such models can still provide a basis for an explanation of the magmatic evolution of the subvolcanic suite.

8.3.1 Method of modelling

The quantitative modelling of proportions of major elements was undertaken using the LSMC (least-squares mixing calculation) method (cf. Mungall, 1993) to assess the proportions of phases that fractionated. The least and the most evolved rock compositions were determined by visual inspection of variation diagrams and whole-rock chemical data for individual intrusive cycles. The trial parental melt for each intrusive center was chosen to have the least SiO_2 , Rb, and Nb, and the highest CaO, ΣFeO , TiO_2 , Ba, Sr, and Zr contents. For the subvolcanic rock suite as a whole, the average compositions of the least and most evolved compositions were used in the model. The composition of the bulk cumulate in terms of major oxides (C_i^s) was calculated from the compositions and proportions of the fractionating phases from LSMC [$C_i^s = \sum X_i^M Y_i^M$, where X_i^M = proportion of mineral M containing oxide (i) in the solid, and Y_i^M = weight % of oxide (i) in a given mineral]. The compositions of fractionating minerals were determined by microprobe analysis (chapters 3 - 5). The least-squares estimates of fractionating phases were subsequently used to calculate bulk Ds, and the trace-element variations were modelled by both perfect fractional crystallization (after the Rayleigh distillation law) and equilibrium crystallization.

8.3.2 Major-element model (LSMC) for the subvolcanic rock suite

In the case of the Buri ring dyke, the LSMC model (Table 8.2A) indicates that the major-element spectrum requires only 17% crystallization of the trial parental melt (i.e., least evolved composition in the Buri ring dyke). The bulk composition of the cumulates (San: 85%, Hd: 11%, and Ilm: 4%) produces the best fit. The compositional range of the cone sheets in the Jubga-Ziem center requires approximately 32% fractionation (based on major elements). The bulk composition of the cumulates (San: 83%, Amph: 5%, Fa: 5%, Qtz: 6%, and Ilm: 1%) is only slightly more evolved (i.e., more SiO₂-rich) than the cumulates calculated for the Buri ring dyke (Tables 8.2A, B). The spectrum of subvolcanic rocks of the Jubga-Ziem ring dyke requires the least extent of fractionation (12%) (Table 8.2C), with a bulk composition of the cumulate (San: 72%, Amph: 21%, Hd: 6%, and Ilm: 1%) that contains considerable amphibole compared to other calculated cumulates. Overall, the calculated cumulates for the individual intrusive centers have a modal mineralogy that does not resemble that of the microsyenodiorite enclaves. This suggests that the enclaves in the subvolcanic rocks do not correspond to the expected cumulates.

The spectrum of rock compositions of the entire subvolcanic suite corresponds to approximately 29% fractionation of the trial parental melt (Table 8.2D). The bulk cumulate (San: 88%, Fa: 6%, Hd: 5%, and Ilm: 1%) has a bulk composition, in terms of major oxides, similar to that of the calculated cumulates in the models for the Buri ring dyke and the cone sheets, but again different from the composition of the enclaves in subvolcanic rocks.

8.3.3 Trace-element evolution of subvolcanic rock suite

According to the Rayleigh distillation law (McCarthy & Hasty 1976): the trace element concentrations of the residual liquid (C_i^l) are given by $C_i^o F^{(D-1)}$ (where C_i^l = concentration of an element i after a given amount of fractional crystallization, C_i^o = initial concentration of element

Table 8.2. Least-squares major-element calculation for the subvolcanic rocks of Kwandonkaya complex. A) The Buri ring dyke, B) Cone sheets, C) Jubga-Ziem ring dyke, D) Subvolcanic rock cycle, and E) Trachyandesite to rhyolite evolution.

A) BURI RING DYKE LSMC BASED ON MAJOR OXIDES (#65--->#41)

				#41	Ilm	San	Hd	(1-F)
Weight mode				83.00	0.54	14.48	1.98	0.17
Weight mode in solids					3.17	85.18	11.65	
	Bulk							
Oxide	M(obs)	M(calc)	Error					Solid
SiO ₂	71.58	72.25	-0.67	74.21	0.02	66.86	49.30	62.70
TiO ₂	0.24	0.49	-0.25	0.25	50.20	0.00	0.38	1.64
Al ₂ O ₃	13.29	13.30	-0.01	12.83	0.06	18.26	0.50	15.61
FeO	3.58	3.36	0.22	2.92	47.91	0.38	31.25	5.48
MgO	0.07	0.04	0.03	0.00	0.08	0.00	1.96	0.23
CaO	0.49	0.89	-0.40	0.62	0.00	0.09	18.23	2.20
Na ₂ O	3.92	4.08	-0.16	3.86	0.00	5.99	0.41	5.15
K ₂ O	5.34	5.42	-0.08	5.21	0.00	7.53	0.05	6.42
$\sum R^2$ (Sum of squared oxide residuals) = 0.75								

B) CONE SHEETS LSMC BASED ON MAJOR OXIDES (#71--->#63)

				#63	Ilm	San	Fa	Amph	Qtz	(1-F)
Weight mode				67.63	0.37	26.93	1.68	1.56	1.83	0.3237
Weight mode in solids					1.14	83.19	5.19	4.82	5.65	
	Bulk									
Oxide	M(Obs)	M(calc)	Error							Solid
SiO ₂	73.02	73.33	-0.31	77.48	0.02	66.86	30.64	37.37	99.71	64.68
TiO ₂	0.28	0.28	0.00	0.12	50.20	0.00	0.02	0.73	0.09	0.61
Al ₂ O ₃	12.78	12.77	0.01	11.41	0.06	18.26	0.01	8.57	0.02	15.61
FeO	3.13	3.14	-0.01	1.80	47.91	0.38	66.07	34.18	0.14	5.95
MgO	0.00	0.02	-0.02	0.00	0.08	0.00	1.26	0.03	0.00	0.07
CaO	0.49	0.52	-0.03	0.48	0.00	0.09	0.33	10.46	0.01	0.60
Na ₂ O	3.81	3.85	-0.04	3.27	0.00	5.99	0.01	1.72	0.00	5.07
K ₂ O	5.11	5.26	-0.15	4.73	0.00	7.53	0.00	1.84	0.02	6.35
$\sum R^2$ = 0.12										

C) JUBGA-ZIEM LSMC BASED ON MAJOR OXIDES (#56--->#38)

				#38	Ilm	Hd	San	Amph	(1-F)
Weight mode				87.87	0.14	0.68	8.79	2.52	0.1213
Weight mode in solid					1.15	5.61	72.46	20.73	
	Bulk								
Oxide	M(Obs)	M(calc)	Error						Solid
SiO ₂	73.62	74.65	-1.03	76.82	0.02	48.25	66.86	37.37	58.92
TiO ₂	0.19	0.19	0.00	0.11	50.20	0.07	0.00	0.73	0.73
Al ₂ O ₃	13.08	13.23	-0.15	12.98	0.06	0.53	18.26	8.57	15.04
FeO	2.21	2.25	-0.04	1.24	47.91	28.69	0.38	34.18	9.54
MgO	0.00	0.00	0.00	0.00	0.08	0.05	0.00	0.03	0.01
CaO	0.77	0.77	0.00	0.40	0.00	21.33	0.09	10.46	3.44
Na ₂ O	3.72	3.84	-0.12	3.72	0.00	0.41	5.99	1.72	4.72
K ₂ O	5.02	5.10	-0.08	5.00	0.00	0.01	7.53	1.84	5.84
$\sum R^2$ = 1.1									

D) SUBVOLCANIC ROCK SUITE LSMC BASED ON MAJOR OXIDES

				S2D	Ilm	Fa	Hd	San	(1-F)
Weight mode				71.05	0.34	1.70	1.42	25.50	0.2895
Weight mode in solids					1.17	5.87	4.91	88.08	
	Bulk								
Oxide	M(Obs)	M(calc)	Error						Solid
SiO ₂	73.20	73.43	-0.23	77.64	0.02	30.64	49.30	66.86	63.11
TiO ₂	0.26	0.26	0.00	0.12	50.20	0.02	0.38	0.00	0.61
Al ₂ O ₃	12.95	12.90	0.05	11.60	0.06	0.01	0.50	18.26	16.11
FeO	3.01	3.02	-0.01	1.68	47.91	66.07	31.25	0.38	6.31
MgO	0.04	0.05	-0.01	0.00	0.08	1.26	1.96	0.00	0.17
CaO	0.66	0.66	0.00	0.52	0.00	0.33	18.23	0.09	0.99
Na ₂ O	3.73	3.86	-0.13	3.27	0.00	0.01	0.41	5.99	5.30
K ₂ O	5.12	5.23	-0.11	4.66	0.00	0.00	0.05	7.53	6.63
$\Sigma R^2 = 0.08$									

E) TRACHYANDESITE TO RHYOLITE EVOLUTION

				Sub	Ilm	Plag	Fe-Aug	San	Opx	(1-F)
Weight mode				5.60	3.70	17.84	6.47	46.73	19.67	0.944
Mode in solid phases					3.92	18.90	6.85	49.50	20.84	
Bulk										
Oxide	M(Obs)	M(calc)	Error							Solid
SiO2	57.62	58.25	-0.63	73.20	0.00	55.22	50.65	67.39	48.47	57.37
TiO2	1.72	1.94	-0.22	0.26	50.36	0.00	0.45	0.00	0.16	2.04
Al2O3	13.94	14.20	-0.26	12.95	0.03	27.89	1.10	17.91	0.29	14.27
FeO	11.34	11.24	0.10	3.01	46.43	0.43	17.47	0.83	39.45	11.73
MgO	1.73	2.57	-0.84	0.04	0.03	0.00	11.05	0.00	9.43	2.72
CaO	3.65	3.31	0.34	0.66	0.05	10.08	18.14	0.00	1.52	3.47
Na2O	4.53	4.46	0.07	3.73	0.00	5.69	0.27	6.89	0.00	4.50
K2O	3.84	3.45	0.39	5.12	0.00	0.23	0.01	6.67	0.00	3.35
ΣR² = 1.5										

Sample #65 and #41 = hedenbergite fayalite granite porphyry from Buri ring dyke, #71 and #63 = granite porphyries from cone sheets, #56 = hornblende biotite granite porphyry (Jubga-Ziem center), and #38 = biotite granite porphyry (Jubga-Ziem center). In the modelling of subvolcanic rocks, the trial parental melt [M(Obs)] is an averaged composition of relatively less evolved subvolcanic rocks (see text), and the target melt (S2D) is an averaged composition of the most evolved subvolcanic rocks. The trial parental melt in the trachyandesite to rhyolite evolution model is an average of trachyandesite enclaves in the Buri ring dyke, whereas the target melt is the same as the trial parental melt used in the LSMC model for the subvolcanic rock suite.

Mineral Abbreviations: Ilm = ilmenite, San = sanidine, Hd = hedenbergite, Fa = fayalite, Amph = amphibole, Qtz = quartz, Plag = plagioclase, Fe-Aug = ferroaugite, and Opx = orthopyroxene.

i in the parental magma, F = the weight proportion of residual melt relative to the initial amount of melt, and D = bulk distribution coefficient, i.e., $\sum X_i K_D$, which is the summation of distribution coefficients of a given element in proportion to constituent mineral phases). With perfect fractional crystallization, the average concentration of a trace element i in the cumulate (C_i^s) is given by $C_i^s = C_i^0[(1-F^{D-1})/(1-F)]$ (Allègre & Minster 1978). During equilibrium crystallization, the melt's evolution is characterized by the equation $C_i^l = C_i^0(1/[D(1-F)+F])$, whereas the solids' bulk composition (not calculated) is given by $C_i^s = C_i^0[D/(D(1-F)+F)]$.

The calculated trace element evolution paths for the whole spectrum of subvolcanic rocks, by both equilibrium and fractional crystallization, are presented, and the most favourable model is indicated. The partition coefficients (K_D) used in this modelling are given in Table 8.3, and the sources of the data are indicated in the footnotes. The K_D of Mahood & Hildreth (1983) and Rämö (1991) were used because of the similarity between the compositions they studied and the Kwadonkaya subvolcanic rocks.

The paths of trace-element evolution calculated in this study are shown in Figures 8.4a - d, and indicate that the degree of fractionation required to reproduce the spectrum of trace-element concentrations in subvolcanic rocks is much larger than that predicted by LSMC based on major elements. In all cases (Figs. 8.4a - c), whereas perfect fractional crystallization requires 50% fractionation to produce the observed range of Ba, Sr, Zr, and Rb, equilibrium crystallization requires 60%. Thus, there is not much difference between equilibrium crystallization and crystal fractionation in this system. Furthermore, Figure 8.4d is a plot of chondrite-normalized REE concentrations for the inferred compositions based on LSMC estimates and trace-element models; it shows that perfect fractional crystallization produces a slightly better fit in the LREE contents of the observed composition of the residual melt at $F = 0.71$ (i.e., 29% crystallization). However, there is no difference in the HREE pattern of the evolved melt produced either by fractional or

Table 8.3: Table of crystal/liquid partition coefficients used in fractional crystallization model.

	Hd	Fa	San	Ilm	Zrn	Aln
Rb	0.032	0.098	0.64 (MH)	0 (R)	0 (R)	0 (R)
Sr	0.516 (R)	0.014	3.87 (MH)	0 (R)	0 (R)	0 (R)
Ba	0.131 (R)	0.10	6.70 (MH)	0 (R)	0 (R)	0 (R)
Zr	0.50	0.00	0.10 (MH)	1.00	4500.00	2 (R)
Nb	0.40	0 (R)	0.10	0.81	0.00	0
La	3.50	2.03	0.071	1.31	26.60	2362.00
Ce	4.00	1.78	0.041	1.19	23.50	2063.00
Nd	5.90	1.50	0.021	0.96	22.00	1400.00
Sm	6.21	1.06	0.014	0.684	17.70	756.00
Eu	5.60	0.42	3.50	0.40	12.00	122.00
Tb	4.00	0.54	0.035	0.36	37.00	235.00
Dy	1.10	0.015	0.37	95.00	123.00	0.39
Yb	4.50	0.98	0.006	0.55	490.00	24.50
Lu	6.10	1.29	0.05	0.74	635.00	22.00
	MH	MH	PEL	MH	MH	MH

MH = Mahood & Hildreth (1983), R = Rämö (1991, Appendix 3), and PEL = Long (1978).

Fig. 8.3: Schematic log-log plot of concentration of a compatible element (e.g., Ba) vs. that of an incompatible element (e.g., Rb), to illustrate idealized paths of evolution for equilibrium (EC) and perfect fractional crystallization (FC) of magma with initial composition B (cf. McCarthy & Hasty 1976, Tindle & Pearce 1981). AB and BC represent paths of cumulate and melt evolution, respectively, for equilibrium crystallization, whereas AD (cumulate) and BE (melt) are evolutionary trends expected during perfect fractional crystallization. The crystal mushes (i.e., cumulate + intercumulate melt) formed during fractional crystallization are intermediate between AD and BE.

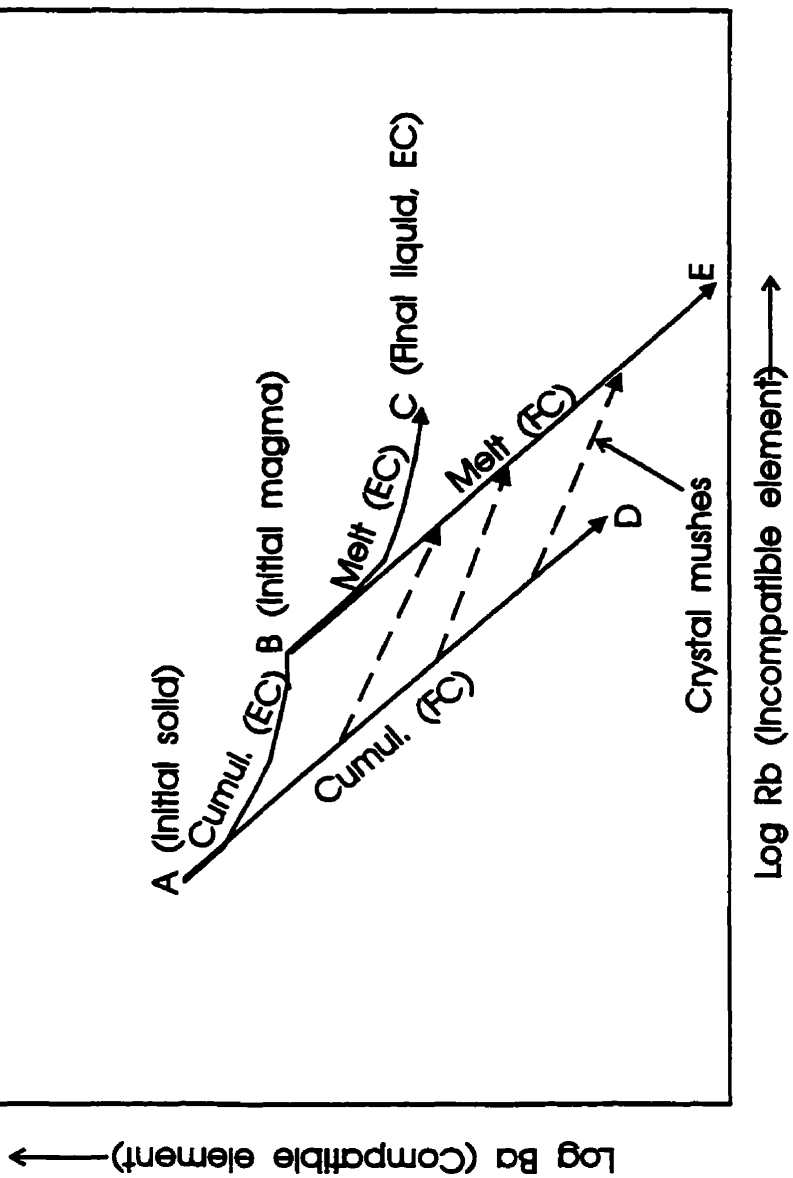
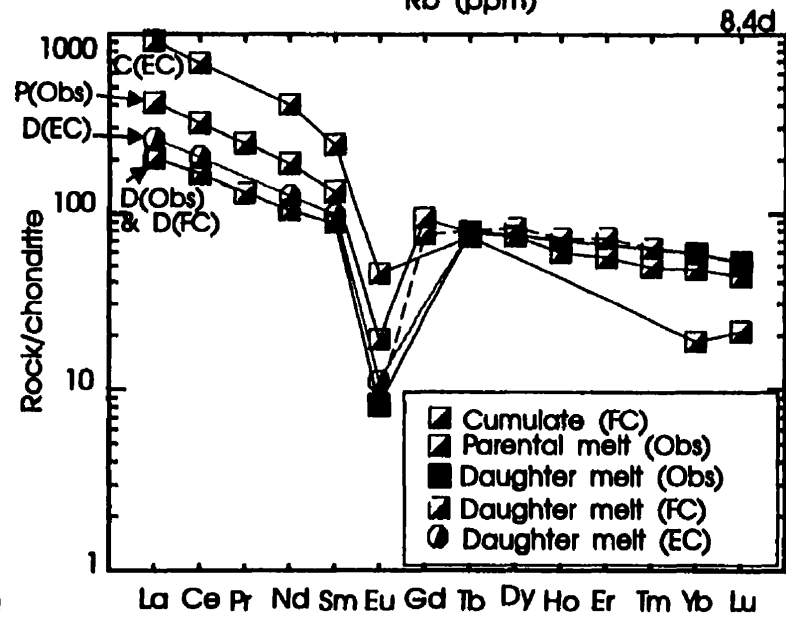
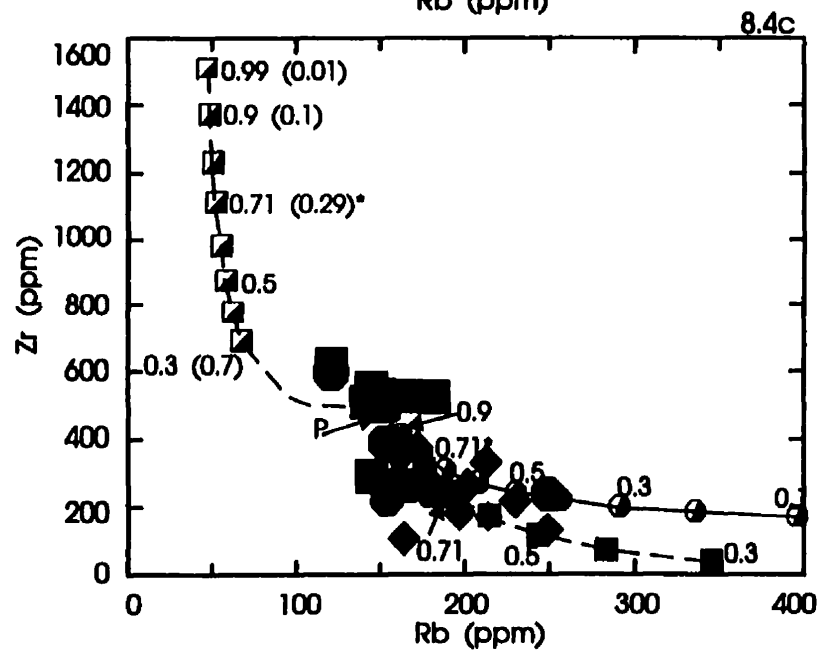
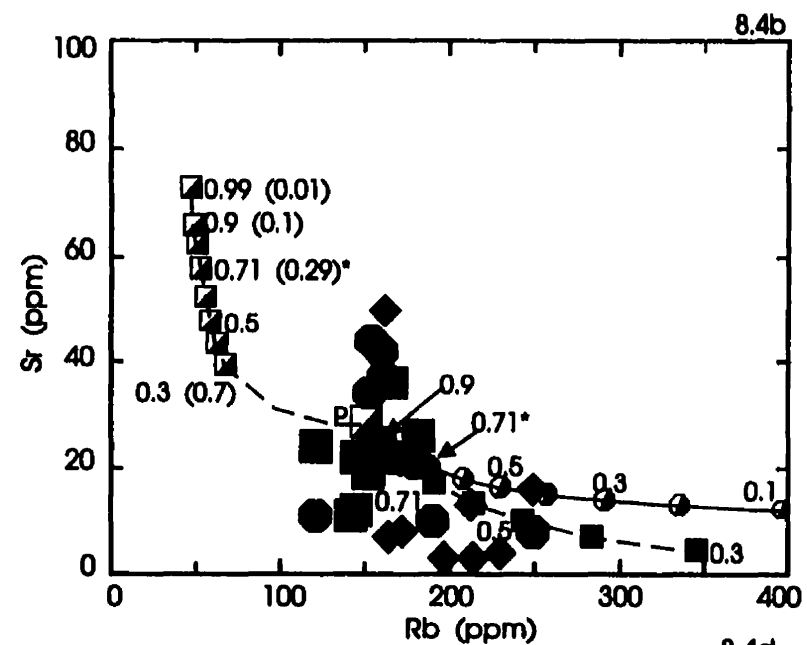
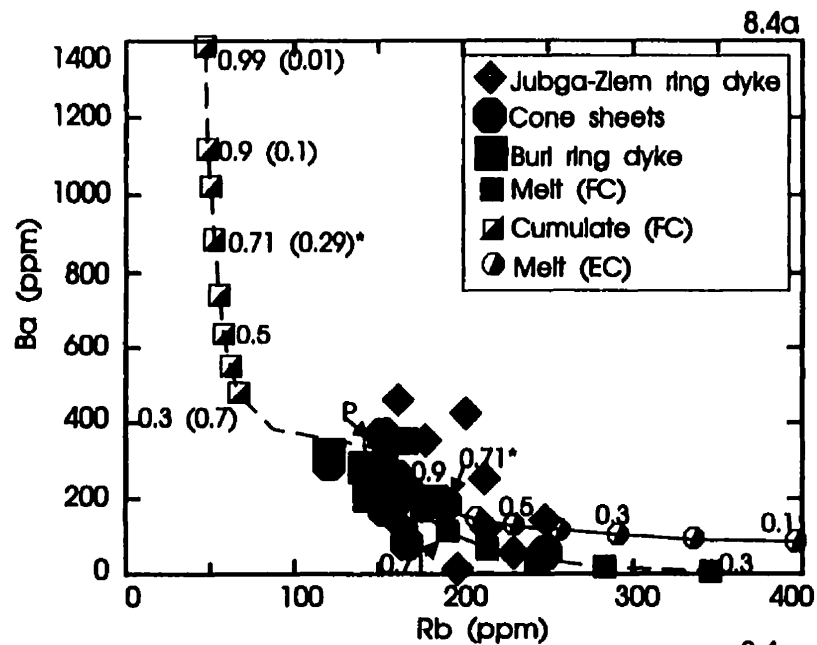


Fig. 8.4: Plots of evolution in trace-element concentrations expected as result of Rayleigh fractional crystallization, and equilibrium crystallization. a) Ba vs. Rb, b) Sr vs. Rb, and c) Zr vs. Rb. The two modes of crystallization do not exhibit significant differences in the melt evolution trends. d) chondrite-normalized REE plots for the various inferred compositions based on LSMC estimates and trace-element models.

The inset in Fig. 8.4a is the legend for Figures 8.4a - c; P in these figures stands for composition of the parental melt. In Fig. 8.4d: Parent (obs) = composition of the parental melt used in LSMC, Daughter (obs) = composition of the daughter melt used in LSMC. Daughter (FC), the composition of daughter melt produced by perfect fractional crystallization, better reproduces the observed composition of the evolved melt than Daughter (EC), the composition of the daughter melt produced by equilibrium crystallization.

The numbers indicate fraction of melt (F) remaining, whereas values in brackets indicate amount of fractional crystallization (1-F). * indicate melt proportion (F) and cumulate proportion (1-F) calculated in LSMC model in Table 8.3D.



equilibrium crystallization. Although both fractional and equilibrium crystallization both appear to be capable of accounting for the evolution of the subvolcanic rocks, the fractional crystallization model is the preferred. Fractional crystallization requires a less extensive interval of crystallization, and the calculated evolutionary trend is closer to the compositions of the subvolcanic rocks. Furthermore, evolutionary trends based on mineral compositions (pyroxene and amphibole in Chapter 3, mica in Chapter 4, and alkali feldspar in Chapter 5) support the fractional crystallization model.

8.4 Review of models previously proposed for the origin of A-type granites

A number of models have been proposed for the origin of the NYG parental magma and A-type magmatic suites in general.

8.4.1 *Mantle source for NYG (A-type) granites*

According to this model, the NYG granites were produced by the fractionation of a mantle-derived mildly alkaline, orthopyroxene-normative olivine basalt to trachyte, followed by formation of rhyolitic compositions by the fractionation of plagioclase, clinopyroxene, and alkali feldspar (Jacobson et al. 1958, 1963, van Breemen et al. 1975, Turner & Bowden 1979, Bowden 1982). The mantle-source model is based on studies of NYG complexes, and is supported by the occurrence of coeval mafic and felsic units (Ike 1979, Bennett et al. 1984, Bowden et al. 1987). This model may adequately explain the evolution of trachytes (syenites) in ring complexes dominantly composed of basalt and trachyte (or gabbro-syenite $\pm$ anorthosite/diorite suite), such as Abontorok, Meugueur-Meugueur, Taguei, Iskou, and Ofoud complexes in Aïr, in the Niger A-type granite province (Bowden et al. 1987, Brown et al. 1989), the sheet-like granite pluton in the Bushveld complex (Kleeman & Twist 1989), or the Mégantic complex in Quebec (Bédard 1989,

Bédard et al. 1987). However, in the Mégantic complex, the biotite granite stock is inferred to have been formed by anatexis of country rocks (Bédard 1989), and the felsic rocks in the anorthosite-bearing complexes of Aïr, in Niger, cannot be adequately accounted for by fractionation from a mafic magma (Bowden et al. 1987).

In complexes where the felsic rock types are predominant, the mantle-source model fails to explain the origin of the associated peralkaline and metaluminous to peraluminous rocks. It is expected that in a transition from mafic to felsic subvolcanic rocks, anorthosite cumulates will occur. However, the mantle-source model fails to explain the observations that only small volumes of mafic lava are documented in the NYG complexes (< 5% of the total exposed area), lavas of intermediate composition are rare (< 1%), and no anorthosite cumulates have been documented so far, even in the deeply eroded NYG complexes in the southern part of the province. Furthermore, gravity surveys across the NYG province indicate negative Bouguer anomalies (Ajakaiye 1976, Ajakaiye et al. 1983) that range from about -30 to -60 milligals (centered on ring complexes), with a regional gravity low of -95 milligals in the centre of the province, in the Jos area. The gravity data are interpreted to indicate the absence of voluminous cumulates (or a mafic base) beneath the predominantly felsic ring complexes of the NYG (Ajakaiye 1976) and the surrounding migmatite-gneiss-granite country rocks. Moreover, the volume of basic magma required to produce the large volumes of granites and rhyolites seems prohibitively large. For example, the Jubga-Ziem center alone, with an area of about 150 km² (length = 13 km, width = 12 km) and a depth estimated to be between 10 and 15 km on the basis of gravity models refined by Ajakaiye (1976), has an estimated volume of 1500 - 2400 km³ of felsic rocks. The volume of basaltic melt required to produce such a volume of felsic rocks, by at least 80% fractionation (cf. Table 8.2E), is 7800 - 11700 km³, and requires a volume of cumulates beneath the complex between 6000 - 9000 km³. These cumulates alone would require a "buried complex" whose size

exceeds the total estimated volume of felsic rocks in the Jubga-Ziem center. If the inferred volume of an eroded felsic (ignimbritic) volcanic cover is included, the required volume of cumulate rocks is nearly doubled. Furthermore, the required volume of cumulates (or mafic base) would yield a higher positive gravity anomaly.

8.4.2 Mixed mantle and crustal sources

Some proponents of the mantle-source model believe that rhyolitic (granitic) magmas are hybrids of mantle-derived magmas and partial melts of the crust (van Breemen et al. 1975, Bowden 1982, Bonin 1990). These hybrid melts are subsequently emplaced at higher levels in the crust, where fractionation and contamination occur (Bowden et al. 1987, Eby et al. 1992). Such a model has been suggested for the NYG complexes because of the wide range of Sr isotopic values ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7048 - 0.7935$) in the felsic rocks of the NYG (Jacobson et al. 1963, van Breemen et al. 1975, Dickin et al. 1987), suggesting variable extents of crustal involvement.

8.4.3 The residual source (or restite) model

This model proposes that A-type felsic magmas are produced by melting of source rocks previously depleted in H_2O by the generation of I-type magmas in the lower crust (Collins et al. 1982, White & Chappell 1983, Chappell et al. 1987, Chappell & White 1991). Creaser et al. (1991) provided a detailed critical review of the restite model, and substantial evidence exists in the NYG and Kwandonkaya to support the objections that they raised. Several characteristics of the Kwandonkaya complex and A-type granites in general do not favour origin from a residual source:

A restite source should be enriched in Ca, Al (in plagioclase), and Mg (in pyroxene and amphibole), and depleted in K, Na, Fe, and Si relative to the original protolith (Creaser et al.

1991). It is therefore unlikely that magmas enriched in K, Na, Si, and Fe, and depleted in Ca, can be generated from such a source. As an example, the range of major-element compositions for the subvolcanic rocks of the Kwandonkaya complex are compared to those of the Pan-African (Older) granite and charnockite from central Nigeria in Table 8.4. Source rocks that have lost a granitic melt fraction similar to the Pan-African Older Granites in composition ($66 \leq \text{SiO}_2 \leq 70\%$, $0.81 \leq \text{F/FM} \leq 0.88$; Kayode 1976) or charnockites ($52 \leq \text{SiO}_2 \leq 71\%$) are unlikely to produce a second generation of melts with compositions of the Kwandonkaya subvolcanic rocks ($71.6 \leq \text{SiO}_2 \leq 76.3\%$, $0.96 \leq \text{F/FM} \leq 1.0$), that are more silica-rich, and depleted in compatible elements (e.g., Cr, Ni) relative to the Pan-African granites (cf. Creaser et al. 1991). Furthermore, high-field-strength elements (e.g., Zr, Nb, Y, and REE) and volatiles are enriched in A-type granites magmas. This seems to make no sense in the context of a residual source model.

8.4.4 *Partial melting model*

This model proposes that the parental felsic magmas in NYG complexes, as well as in A-type granitic complexes in general, were derived by partial melting of nonresidual lower crustal rocks (Bowden et al. 1976). The difference between this and the preceding model lies mainly in the fact that the source rock was not previously melted, and the source areas are invariably enriched in volatiles and some elements by mantle-derived fluids. The melting is assumed to have occurred because of fluxing by mantle-derived volatiles (Bailey 1987). Several source-rock compositions have been previously proposed, and include: charnockite (Brown & Bowden 1973, Beard et al. 1992), granulite (Condie 1991), granodiorite and tonalite (Rämö 1991, Skjerlie & Johnston 1992, 1994).

Table 8.4: Range of concentrations of major and selected trace element for the Pan-African (Older) Granites and subvolcanic rocks of the Kwandonkaya complex.

Oxide wt. %	Pan-African (Older) Granites	Charnockitic suite	Subvolcanic rocks (Kwandonkaya complex)
SiO ₂	66.45 - 69.80	51.97 - 70.59	71.58 - 76.29
TiO ₂	0.50 - 1.11	0.31 - 2.99	0.25 - 0.36
Al ₂ O ₃	14.00 - 15.69	13.90 - 18.59	12.28 - 13.29
Fe ₂ O ₃	0.40 - 1.14	0.77 - 2.59	1.33 - 3.56
FeO	2.67 - 4.15	2.30 - 11.37	0.42 - 1.62
MnO	0.05 - 0.07	0.04 - 0.18	0.04 - 0.07
MgO	0.42 - 1.23	0.30 - 3.16	0.00 - 0.21
CaO	1.68 - 2.81	1.74 - 7.36	0.49 - 0.85
Na ₂ O	2.62 - 3.33	3.19 - 4.14	3.86 - 3.92
K ₂ O	4.68 - 6.33	1.89 - 6.03	4.83 - 5.34
P ₂ O ₅	0.20 - 0.47	0.12 - 0.75	0.02 - 0.16
F/FM	0.81 - 0.88	0.81 - 0.95	0.96 - 1.00
Trace elements (in ppm)			
Rb	181 - 247	31 - 82	121 - 249
Sr	136 - 226	365 - 677	3 - 50
Ba	774 - 1298	n.d.	14 - 460
Zr	316 - 671	n.d.	108 - 622

Data for Pan-African (Older) Granites are taken from Kayode (1976). The charnockites include dioritic, monzonitic, and fayalite granite types (Eborall 1976), with trace-element data for charnockites from Dada et al. (1991). The data for the Kwandonkaya subvolcanic rocks are from this study. n.d. = not determined.

8.5 Partial melting origin for the parental melt of the subvolcanic rocks

This model involves an attempt to produce the parental magma (target melt) by partial melting of differing source compositions corresponding to global averages of the lower crust, granulite, and granodiorite. The source modes and bulk composition (in terms of trace elements) of the trial source-rocks are given in Table 8.5a, and the K_d values (partition coefficients) used are those compiled by Rämö (1991) (Table 8.5b), and were assumed to remain constant during the melting event.

Because the subvolcanic rocks (and associated plutonic rocks) of the Kwandonkaya complex are relatively restite-free, it is estimated that the required extent of melting ranged from 20% to 60% (cf. Christiansen et al. 1986, Rämö 1991). This covers the range of 30 to 50% proposed by Wickham (1987), required to initiate convective motion in zones of melting so as to result in effective segregation of large homogeneous bodies of magma, and the range of 40 to 60%, indicated by Brown & Bowden (1973) in their experimental partial melting of charnockitic fayalite quartz monzonite to produce metaluminous rhyolite in the NYG.

None of the source-rock compositions tested could yield the target melt composition at 20 to 60% melting. The source-rocks are too depleted in all elements except Sr (in the average lower crust), Ba, Sr, Eu, and Lu (in granulite), and Rb, Ba, Sr, and Eu (in granodiorite). Whereas the average lower crust was considered a less likely source-rock owing to the magnitude of enrichment required, granodiorite appears to require significant depletion of an unlikely suite of elements considering the fact that the source of the enriching fluids is the mantle. The granulite is inferred to be the most likely source-rock, as will be shown in following sections.

8.5.1 *Equilibrium modal melting*

In equilibrium modal melting, the proportions of the constituent minerals of a source

Table 8.5a: Source mode and trace-element composition of source rocks (protoliths) used in partial melting model (cf. Table 11 of Rämö 1991).

MINERAL vol. prop.	LOWER CRUST (TM)	GRANULITE (M)	GRANODIORITE (W)
Hornblende	0.05	0	0.045
Biotite	0.10	0.16	0.108
Plagioclase	0.29	0.53	0.402
Alkali feldspar	0.007	0.00	0.1999
Clinopyroxene	0.05	0.00	0.00
Orthopyroxene	0.29	0.058	0.00
Quartz	0.17	0.25	0.246
Oxide	0.04	0.002	0.00
Zircon	0.0001	0.0002	0.0002
Apatite	0.0005	0.0005	0.0005
Allanite	0.0001	0.0002	0.0002
ELEMENT (ppm)			RP
Rb	5.3	58	170
Sr	230	285	372
Ba	150	483	938
Zr	70	241	191
La	11	31	36
Ce	23	72	74
Nd	12.7	36	37
Sm	3.17	9.75	6.41
Eu	1.17	4.13	1.75
Tb	0.59	1.18	0.97
Yb	2.20	3.73	3.22
Lu	3.37	0.9	0.65

Symbols: TM = Taylor & McLennan (1985), M = Merilainen (1976), W = Wyllie (1977). The trace-element data for the average granulite are from RP = Rudnic & Presper (1990).

Table 8.5b: Partition coefficients used in evaluation of partial melting of trial source-rocks.

Element	Hbl	Bt	Plag	San	Opx	Cpx	Fa	Qtz	Mag	Ilm	Zrn	Apat	Aln	Grn
Rb	0.014	5.50	0.041	0.34	0.003	0.032	0.00	0.014	0.50	0.00	0.00	0.00	0.00	0.01
Sr	0.022	0.12	9.00	3.87	0.009	0.516	0.00	0.00	0.70	0.00	0.00	5.00	0.00	0.15
Ba	0.044	6.40	0.308	6.12	0.003	0.131	0.00	0.023	0.40	0.00	0.00	0.00	0.00	0.015
Zr	1.40	2.00	0.10	0.10	0.20	0.50	0.00	0.00	0.80	1.00	4500.00	0.10	2.00	0.50
La	0.33	0.11	0.30	0.071	0.10	0.30	1.50	0.015	0.89	7.10	2.00	20.00	2600.00	0.35
Ce	1.52	0.32	0.27	0.044	0.015	0.50	1.40	0.014	1.10	7.80	2.64	34.70	2250.00	0.35
Nd	4.26	0.29	0.21	0.025	0.22	1.11	1.10	0.016	1.65	7.60	2.2	57.10	1620.00	0.50
Sm	7.77	0.26	0.13	0.018	0.27	1.67	0.75	0.014	2.00	6.90	3.14	62.80	850.00	2.60
Eu	5.14	0.24	2.15	1.13	0.17	1.56	0.25	0.056	0.47	2.50	3.14	30.40	110.00	1.00
Tb	12.00	0.35	0.06	0.006	0.65	1.80	0.43	0.017	2.45	6.50	100.00	20.00	270.00	35.00
Yb	8.38	0.44	0.049	0.012	0.86	1.58	0.95	0.017	1.10	4.10	270.00	23.10	30.00	40.00
Lu	5.50	0.33	0.046	0.006	0.9	1.54	1.27	0.014	1.05	3.60	323.00	20.20	30.00	30.00

*Source: Kd file used by Rämö (1991) in modelling studies of Finnish A-type Rapakivi granites; Rämö compiled Kd values from various sources, including: Schnetzler & Philpotts (1970), Nagasawa & Schnetzler (1971), Mahood & Hildreth (1983), and Nash & Crecraft (1985).

entering the melt is the same as their proportion in the source. The trace-element evolution during this type of melting can be modelled by the equation: $C_L/C_o = 1/[D_o(1-F) + F]$ (Shaw 1970), where C_o = initial concentration of trace element i in the source rock (protolith), C_L = weight concentration of the trace element in the derived melt, $D_o = \sum X K_D$, i.e., the summation of modal melt proportion multiplied by the partition coefficients of the respective phases, and F = weight fraction of minerals going into the melt relative to the source rock.

Figures 8.5a - c present the predicted composition of melts generated by modal melting of non-enriched source-rocks, normalized to chondritic values, and the derived melts are compared on an equivalent plot in Figure 8.5d. It is clear that none of the proposed lower crustal rocks can yield target-melt compositions without selective enrichment in trace elements. There are only slight differences in pattern between 20 and 60% modal melting. The granodiorite source yields melts more enriched in Rb, Ba, Sr, and Eu than the target melt between 20 and 60% modal melting (Figs. 8.5c, d).

8.5.2 *Equilibrium nonmodal melting*

Unlike modal melting, the proportion of minerals entering the melt in nonmodal melting differs from the initial mode of the source. This type of melting can be described by the equation: $C_L/C_o = 1/[D_o + F(1-P)]$ (Shaw 1970), in which the terms C_L , C_o , D_o , and F are the same as described in the case of modal melting, and P is equal to $\sum p^i K_d^i$ (i.e., bulk distribution coefficient for the minerals that are entering the melt, in which p^i is the weight fraction of mineral i entering the melt). The initial values of p^i used (i.e., the weight fraction of minerals entering the melt) in this modelling were taken from previous studies on the melting of lower crustal compositions (Table 8.5a, Appendix VIII.2a, cf. Rämö 1991).

The trace element pattern of the derived melts (Figs. 8.6a - d) are not significantly **Fig.**

8.5: Chondrite-normalized plots of trace-element concentrations in modal melts (generated by 20 to 60% melting of selected non-enriched source-rocks). T-melt in all Figures refers to the target melt. **a)** Average lower crust (of Taylor & McLennan 1985); ALC20MM and ALC60MM are composition of melts produced by 20 and 60% modal melting, respectively, of the average lower crust. **b)** Average granulite (of Rudnic & Presper 1990); AGr120MM and AGr160MM are composition of melts produced by 20 and 60% modal melting, respectively, of granulite. **c)** Average granodiorite (of Wyllie 1977); AGrd20MM and AGrd60MM are compositions of melts produced by 20 and 60% modal melting, respectively, of granodiorite. **d)** A comparative plot of composition of melts produced by modal melting (cf. Figs. 8.5a - c).

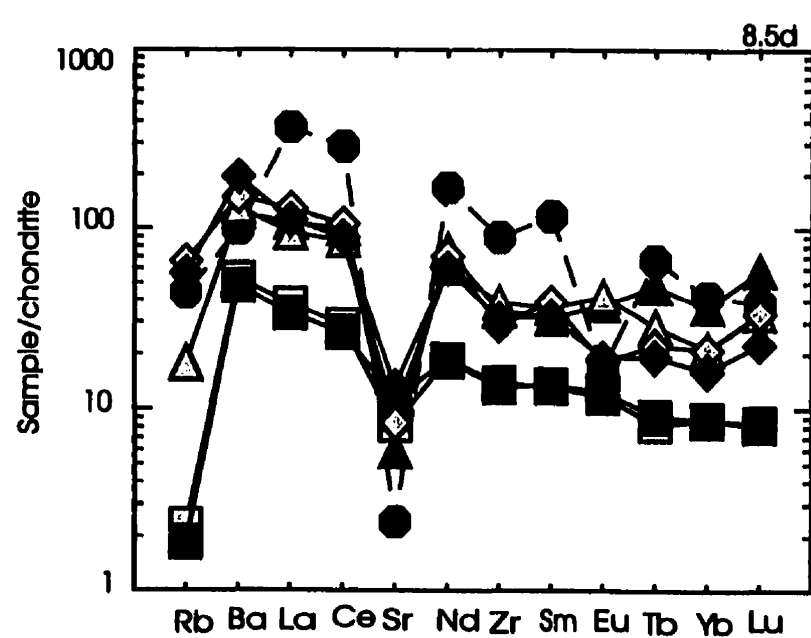
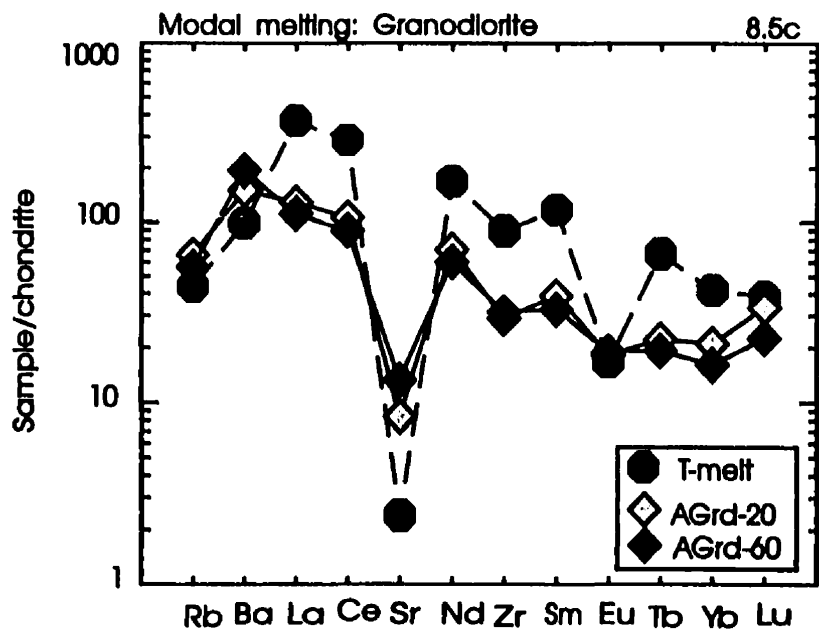
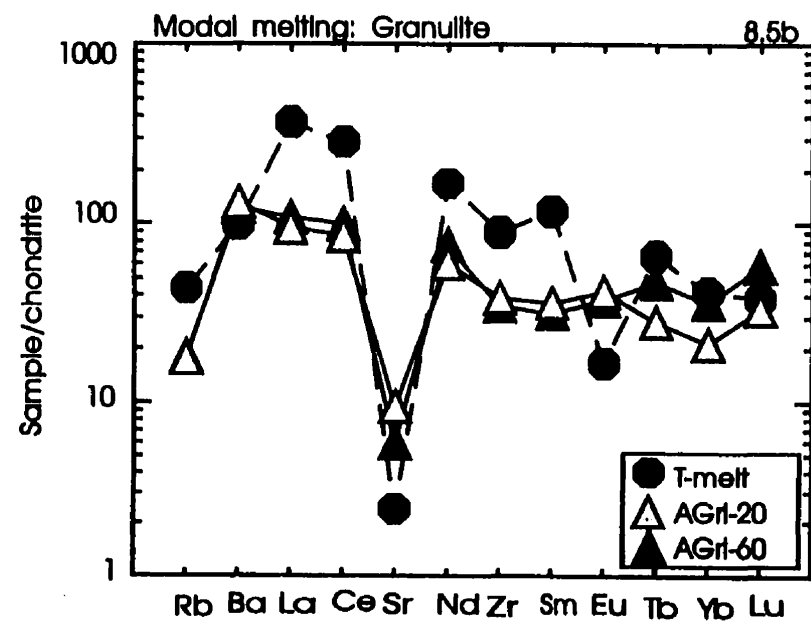
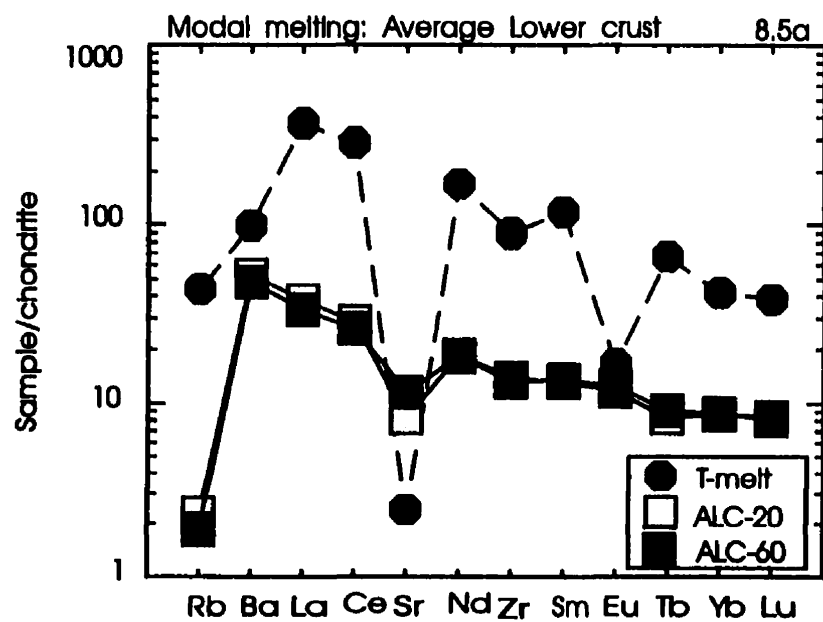
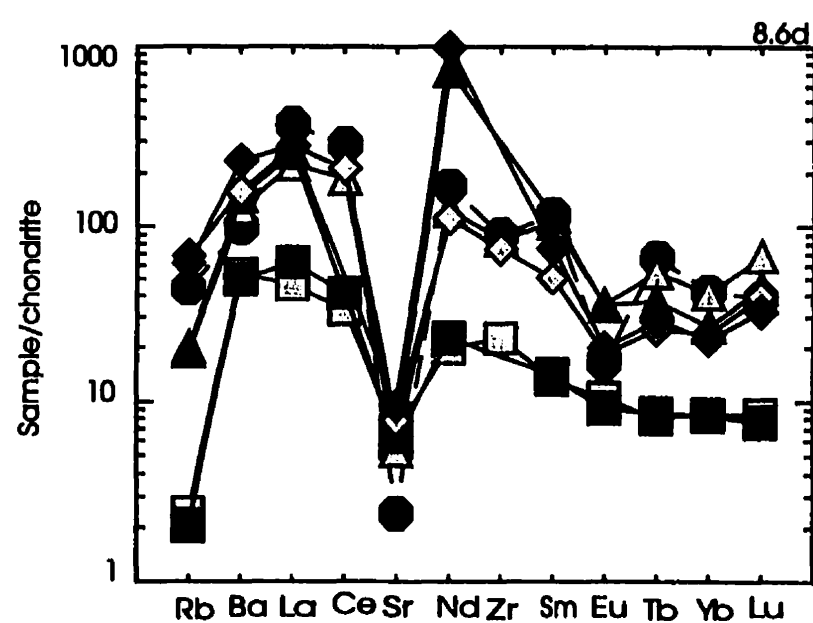
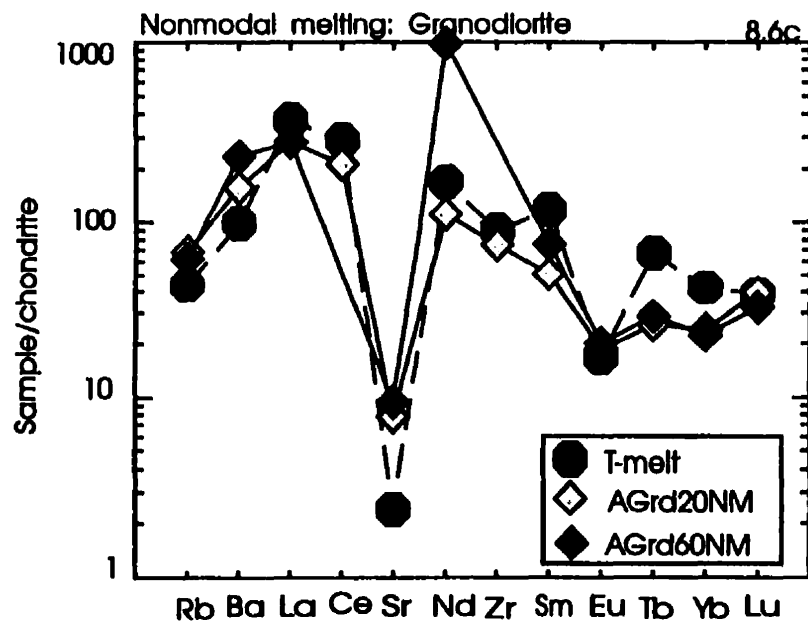
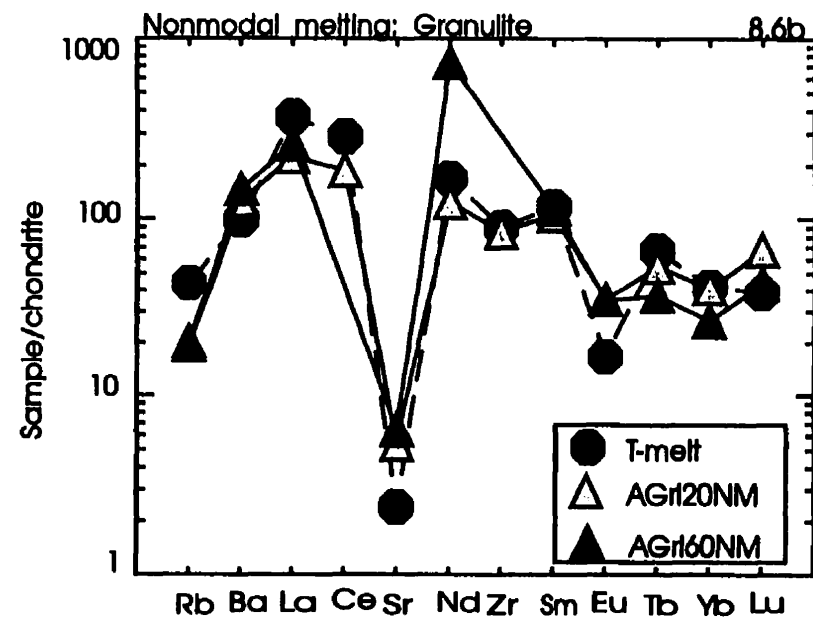
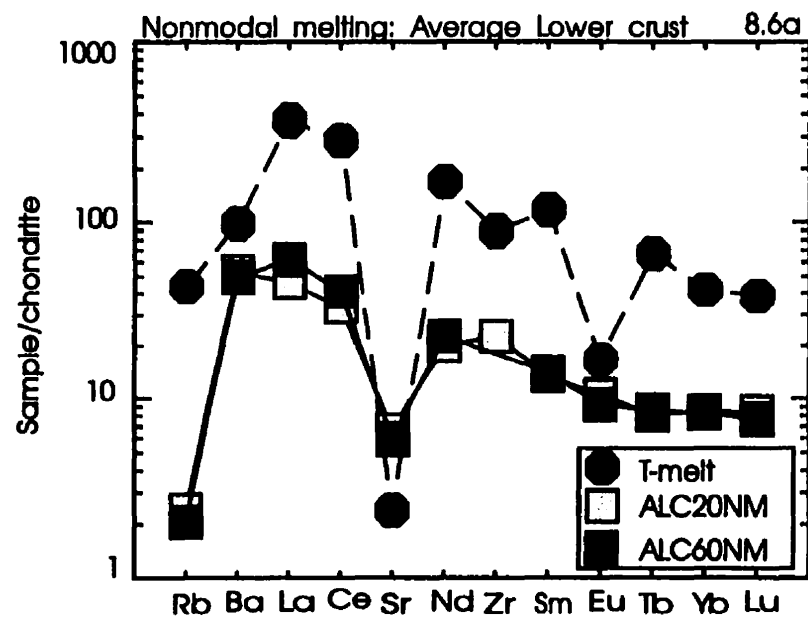


Fig. 8.6: Chondrite-normalized plots of nonmodal melts generated by 20 to 60% melting of non-enriched trial source-rocks: **a)** Average lower crust (ALC20NM and ALC60NM are melts produced by 20 and 60% nonmodal melting, respectively, of the average lower crust). **b)** Average granulite (AGrl20NM and AGrl60NM are melts generated by 20 and 60% nonmodal melting, respectively, of granulite). **c)** Average granodiorite (AGrd20NM and AGrd60NM are melts derived by 20 and 60% nonmodal melting, respectively, of granodiorite). **d)** Melts generated from the three protoliths compared (symbols same as in Figs. 8.7a - c). Source modes (Table 8.5a) and p^i (Table 8.5b) from various sources, as indicated in footnotes. In all plots, T-melt = target melt.



different from those derived by modal melting (Figs. 8.5a - d). Furthermore, nonmodal melting of a granulitic protolith yields the closest match to the target melt (Fig. 8.6b), with the HREE pattern being best matched at 60% melting. The melts generated by nonmodal melting of a granodiorite source have higher levels of Rb, Ba, Sr, and Eu than the target-melt (Fig. 8.7c). The only options through which granodiorite can yield the target-melt either involve selective metasomatic enrichment and depletion, or adjustment of p^i (i.e., the weight fraction of mineral i entering the melt). Attempts to adjust the p^i in the case of granodiorite were futile with respect to Sr and Ba, even at $P_{Sr}, P_{Ba} = 0$ (i.e., even if, for example, Sr and Ba are not entering the target-melt). The content of these elements (at both 20 and 60% nonmodal melting) in the derived melts was found still to be in excess of the target melt composition. The other alternative would be to deplete Rb, Ba, Sr, and Eu from the granodiorite during alteration and to enrich the rest of the deficient elements in Figure 8.7c. Such enrichment-depletion processes are difficult to envisage, particularly in comparison to documented trends of trace-element enrichment during metasomatism by mantle-derived melt or fluids (cf. Harte 1987, Morogan & Martin 1985, Morogan 1988). For the foregoing reasons, granodiorite is not considered further as a possible source-rock.

To derive the target-melt from either the average lower crust or granulite, either the p^i values have to be modified, or the source-rocks have to be selectively enriched so as to yield the parental melt. These possibilities are examined below.

8.5.3 Parental melt generation by modifying p^i values

One approach that can be used to generate suitable parental melt compositions is to calculate the p^i values that will yield the target-melt composition by 20 to 60% nonmodal melting. The adjusted p^i values were calculated (using a spreadsheet) by placing the K_d values listed in Table 8.5b into a twelve by twelve (square) matrix. Then the new p^i values are calculated by

matrix algebra from the equation: $Ax = P$ (where A represents the twelve by twelve matrix or K_d values, $x = p^i$ values, and P is the bulk melt partition coefficient for the minerals that are entering the melt). The expression for the modified p^i values was derived by multiplying both sides by an inverse of A (i.e., $A^{-1}Ax = A^{-1}P$) to give an expression for the term x' (adjusted p^i values): $x' = A^{-1}P$. The required modifications to the p^i values are unrealistic because some of the adjusted values need to be greater than 100% or 1 for some individual minerals in order to generate the parental melt and correct source-rock compositions. This suggests that the parental melt of Kwadonkaya complex cannot be generated by simply modifying the p^i values (i.e., the weight fraction of mineral i entering the melt) of either the average lower crust or granulite.

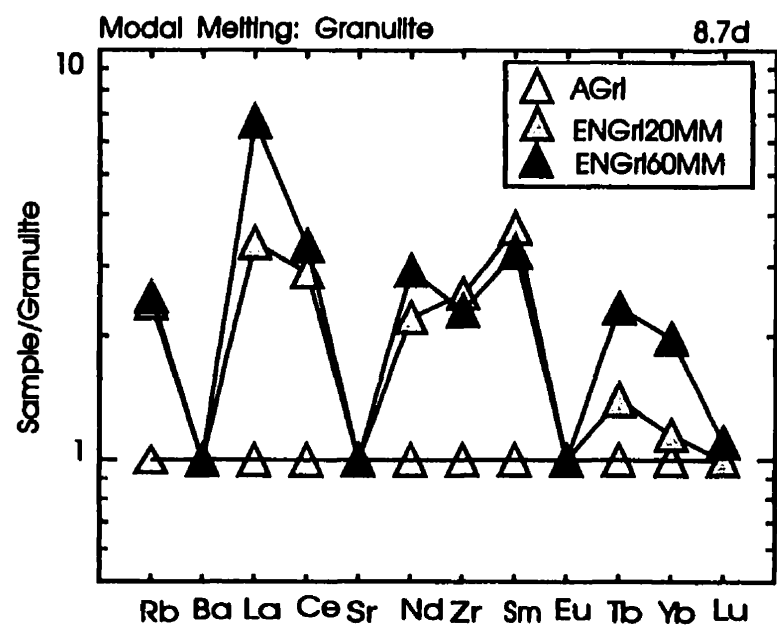
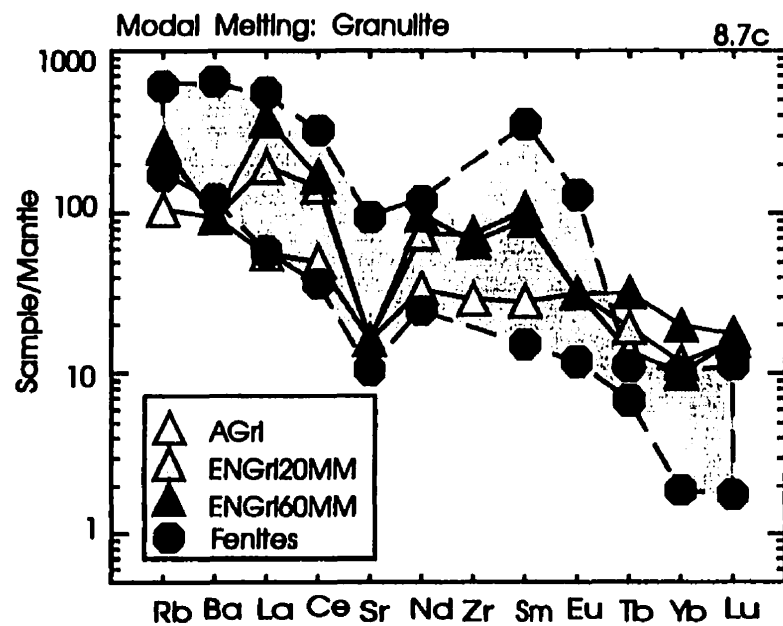
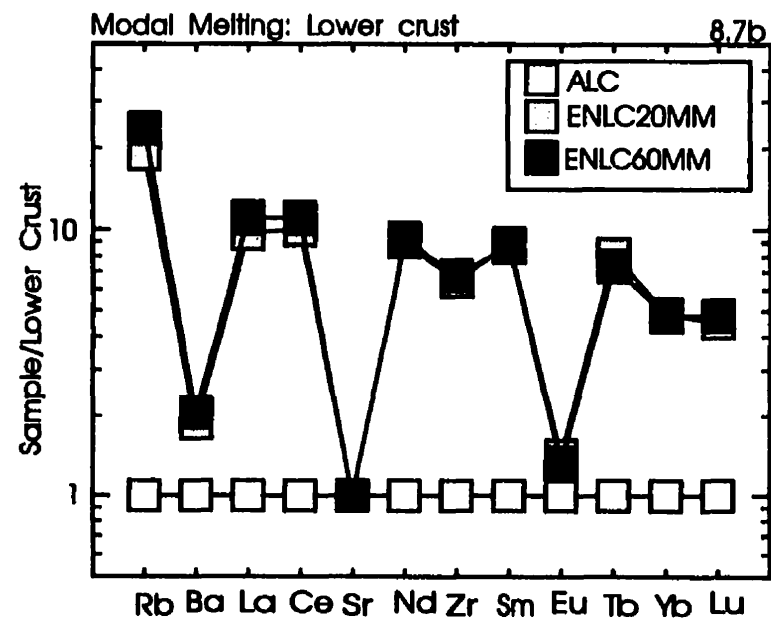
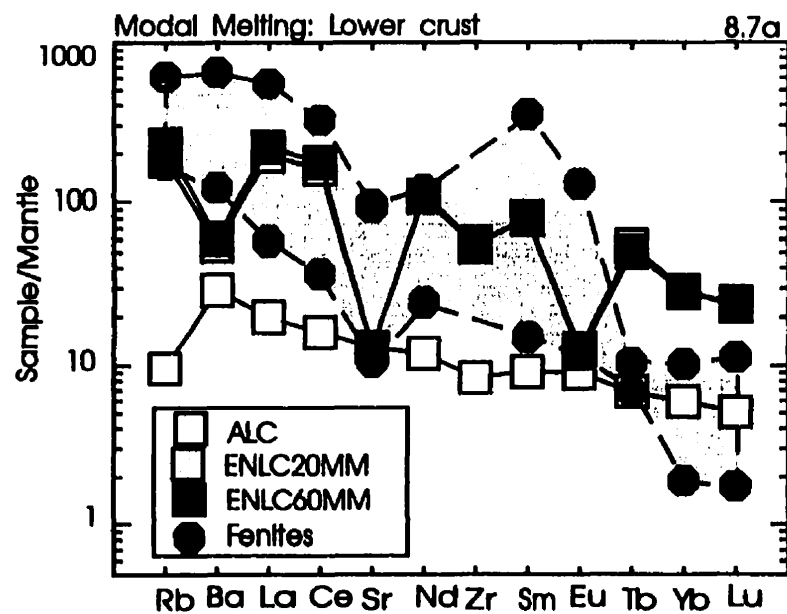
8.5.4 Generation of parental melt by metasomatic enrichment followed by partial melting

Since the normal compositions of the trial source-rocks could not yield the appropriate parental melt composition, the source rocks therefore require metasomatic enrichment. This was done by calculating the appropriate trace-element composition of source rocks that will be capable of producing the parental melt for the subvolcanic rocks at 20% and 60% modal and nonmodal melting. In the case of modal melting, the appropriate source-rock composition was calculated using the equation $C_{\alpha(M)}^i = C_{PL}^i[D_o^i(1-F) + F]$ (where $C_{\alpha(M)}^i$ = concentration of trace element i in the source required to yield parental melt, C_{PL}^i = concentration of trace element i in the parental melt, D_o^i = bulk distribution coefficient, and F = weight fraction of minerals going into the melt relative to the source rock. For nonmodal melting the equation: $C_{\alpha(M)}^i = C_{PL}^i[D_o^i + F(1-P)]$ was used (where P = bulk distribution coefficient of the minerals that are entering the melt, and other terms are the same as in the equation for modal melting). The F values ranged from 0.2 to 0.6 (i.e., 20 to 60% partial melting).

The enrichments required to modify the average lower crust and granulite so as to yield

Fig. 8.7: Composition of source rocks in terms of selected trace elements and required enrichments (for generation of target melt by 20% and 60% modal melting). a) Average lower crust and enriched lower crust normalized to the mantle. b) Enriched lower crust normalized to non-enriched equivalent. In Figures 8.7a and b: ALC represents the average lower crust (of Taylor & McLennan 1985), ENLC20MM and ENLC60MM are enriched lower crust (protoliths) that yield target-melt at 20 and 60% modal melting, respectively. c) Average granulite and enriched granulite normalized to the mantle. d) Enriched granulite normalized to non-enriched equivalent. In Figures 8.7c and d: AGr1 represents average granulite (of Rudnic & Presper 1990), ENGr120MM and ENGr160MM are enriched granulite (protoliths) that yield target-melt composition at 20 and 60% modal melting, respectively.

The fenites are from the Berang and Stornaset areas of the Fen complex, and were formed by the metasomatism of a migmatite protolith (Morogan 1988).



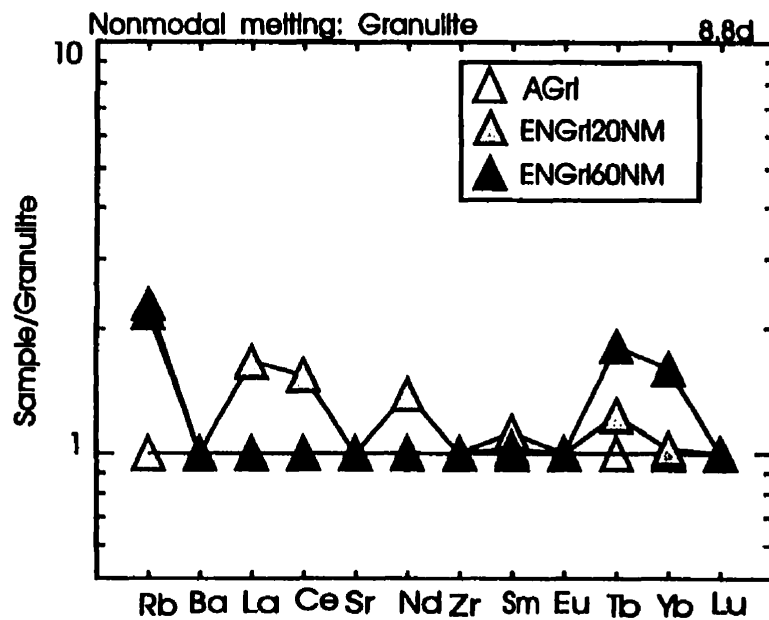
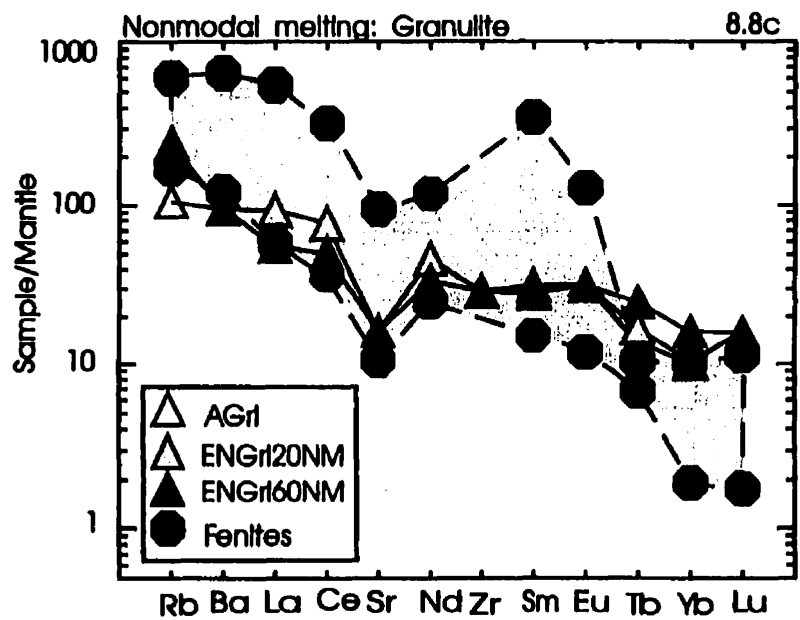
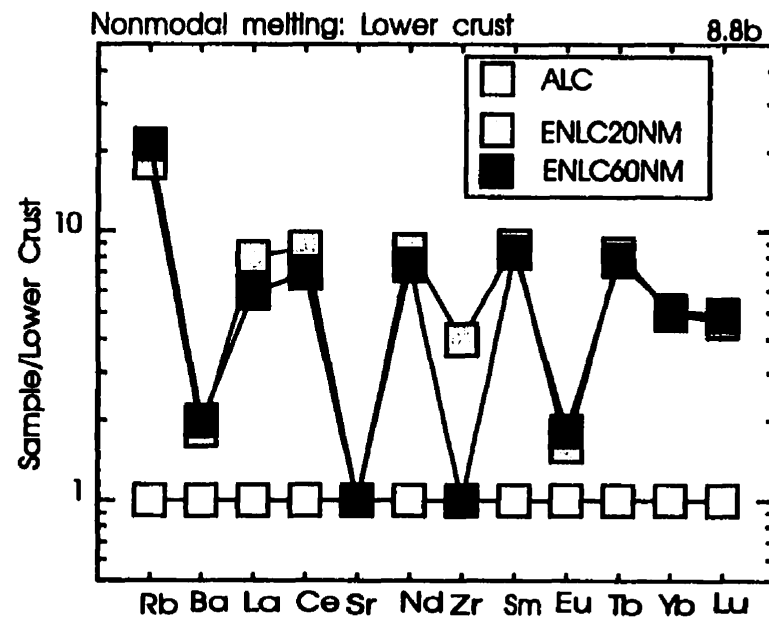
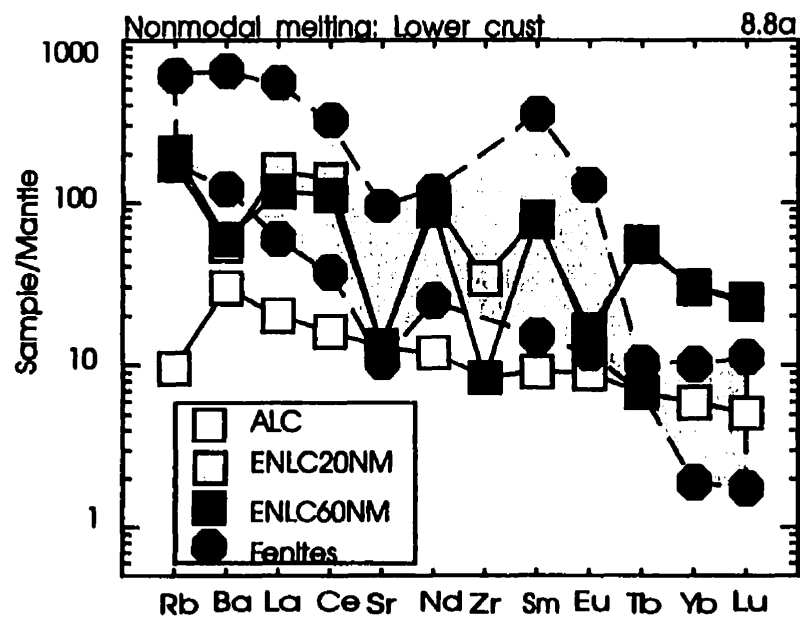
an appropriate parental melt composition by modal melting are shown in Figures 8.7a - d (data Appendix VIII.1b-i). The enriched source-rock compositions are compared to fenitized migmatite from the Berang and Stornaset areas of the Fen complex (Figs. 8.7a, c), taken as an example of metasomatic modifications associated with carbonatites and alkaline rocks. The observation that the necessary enrichments fall within the range of documented metasomatic modifications of crustal rocks (i.e., fenitization) suggests that this mechanism, in theory, is a possible mode of generating the parental melt composition. However, as noted in fenitized zones, it is expected that the mode and mineral assemblages of the source-rock will be modified. Therefore, the requirement in modal melting that the proportions of the constituent minerals of a source-rock participating in a melting reaction be the same as their proportion in the source makes it doubtful that modal melting was the mode of generation of the parental melt.

Figures 8.8a - d display the enrichments required to modify the average lower crust and granulite in order to generate the parental melt by nonmodal melting. As in the case of modal melting (cf. Figs. 8.7a and c), the required enrichment for both average lower crust and granulite are similar to those documented in zones of fenitization, except that the required HREE enrichment for both the average lower crust and granulite are higher than that in fenites (Figs. 8.8a, c). This might be explained by higher contents of F in metasomatizing fluids that altered the source rocks for the NYG (Kwandonkaya) parental magma. According to Keppler (1993), in a F-bearing fluid, the solubility of REEs increase from LREE to HREE. Therefore, the higher HREE enrichment required for the source rocks necessitates a higher F content in the metasomatizing fluids that affected the lower crust.

All the source-rocks considered yield melts with higher Sr and Eu than the target melt. This might reflect a higher content of plagioclase than in the residue and low $f(\text{O}_2)$, favouring Eu^{2+} over Eu^{3+} , and consequently greater retention of Eu in residual plagioclase. Furthermore, the

Fig. 8.8: Plots of enrichments required to modify trial source-rock (for the generation of target-melt compositions by 20 to 60% nonmodal melting). a) Average lower crust and enriched lower crust normalized against mantle compositions and compared to fenites from the Fen complex. b) Average lower crust normalized against nonenriched equivalent. c) Average granulite and enriched granulite normalized against mantle compositions and compared to fenites from the Fen complex. d) Enriched granulite normalized to nonenriched equivalent.

In Figures 8.8a and b: ALC = average lower crust (of Taylor & McLennan 1985), ENLC20NM and ENLC60NM = enriched lower crust that yield target melts at 20 and 60% nonmodal melting, respectively; in Figures 8.8c and d: AGrl = average granulite (of Rudnic & Presper 1990), ENGrl20NM and ENGrl60NM = enriched granulite that yield target melt at 20 and 60% nonmodal melting, respectively. The fenite data are taken from Morogan (1988).



normalized plot of trace elements for the enriched lower-crust protoliths has a less steep slope than that of the fenites (Figs. 8.8a, c). The observed difference in slope suggests that the ratio of incompatible to compatible elements in metasomatizing fluids that affected the protoliths for the Kwandonkaya complex was lower than that in fenitizing fluids. This could also be an artefact of degassing and loss of alkalis, F, and incompatible elements in a low-pressure subvolcanic environment. Overall, an enriched granulite is the most favoured as the source-rock because the unenriched granulite produces the closest matching target melts (Figs. 8.6b, d), and thus requires the least magnitude of enrichments to yield the appropriate parental melt (Figs. 8.8b, d). Nonmodal melting following metasomatic enrichments is the favoured mode of generation of the parental melt because of the anticipated changes in either the mineral compositions or modal mineralogy of the source-rock.

8.6 Discussion

8.6.1 *Evolution of the subvolcanic suite*

Figures 8.4 a - d shows that both perfect fractional crystallization and equilibrium crystallization yields similar trends, and could possibly account for the evolution of the subvolcanic rocks in the Kwandonkaya complex. However, Rayleigh fractionation is more effective in enriching or depleting trace elements in the suite of subvolcanic rocks and is therefore favoured.

The observation that the extent of fractionation estimated by LSMC is relatively less than the extent of fractionation required to produce the spectrum of observed compositions in terms of trace elements is a commonly observed petrological phenomenon. This discrepancy can arise from the fact that some prominent magmatic accessory minerals (zircon, chevkinite, and allanite)

that could significantly affect trace-element enrichments or depletions are not required or accommodated in the major-element model.

8.6.2 Genesis of parental melt for the subvolcanic suite

Any model involving a crustal source-rock for the NYG complexes must take into account the role of the ensialic Pan-African orogenic event as a ground-preparation stage. Features produced during the Pan-African orogeny that are critical to the development of suitable source-rock(s) for NYG magmas include the presence of shear zones and transcurrent faults (Bowden 1985, Black et al. 1985, Bowden et al. 1987, Bonin 1990, Ferré et al. 1995). Recent interpretations of radar imagery of basement rocks around the Kwandonkaya complex by Ferré et al. (1995) have shown clearly that the complex was emplaced along a shear zone subparallel to major fold axes in the Pan-African basement (Fig. 2.2b). According to Bowden et al. (1987), the reactivation of lithospheric shear-zones and within-plate incipient rifting was accompanied by pressure release, channelling of volatiles, partial melting and the generation of NYG magma in the Mesozoic. Such transcurrent faults provided access for mantle-derived F, LILE, HFSE ($\pm\text{CO}_2$, $\pm\text{H}_2\text{O}$) into the overlying lower crust. The second feature involves the recrystallization of the lower crust to yield the Pan-African granulites (of dioritic to granitic composition: Ajibade et al. 1987).

In this study, models testing the partial melting of a variety of source-rock compositions, including global averages of lower crust, granodiorite, and granulite, could not reproduce the composition of the target parental magma for the subvolcanic rock suite. These source rocks had to be enriched "from below", i.e., by mantle-derived metasomatic fluids, the magnitudes of the enrichments being displayed in Figures 8.7a - d and 8.8a - d.

The favoured model requires metasomatic enrichment of source-rocks in the lower crust

followed by nonmodal melting. Of the source-rock compositions tested, the granulitic composition is the most strongly favoured, followed by an inferred average lower crust. In extensional (and epeirogenic) regimes, such as that displayed by the Niger-Nigeria A-type granite province, metasomatic enrichment is envisaged to have accompanied or followed the emplacement of mantle-derived melts into the lower crust (some of which may have been extruded at some ring centers). Such infiltration of metasomatic fluids could have caused alterations accompanied by changes in the modal mineralogy (also known as modal metasomatism: Harte 1987) in the lower crust of north-central Nigeria. Modal metasomatism usually leads to the addition of amphibole, biotite, $\pm$ clinopyroxene, apatite, and Fe-Ti oxides, as indicated by Harte (1987). Such modal metasomatic enrichments reflect the addition of alkalis (K, Na), Rb, Ba, Sr, Zr, REEs, volatiles (F, Cl, $\pm$ CO₂), ($\pm$ Ti, Ni, Cr, V, Nb, Th, U) (Le Bas 1977, Bowden 1985, Morogan & Martin 1985, Harte 1987, Morogan 1988, Martin & Morogan 1988). The relative enrichment in trace elements estimated for the crustal source of alkaline, silica-saturated magmas falls within the range observed in the "type locality" of fenitization (Figs. 8.7a - d, and 8.8a - d). The observation that LREE are inferred to be enriched to a greater extent than the HREE at Kwandonkaya indicates relatively low F contents; on the other hand, the fact that inferred levels of HREE for the enriched source-rock are higher than those in the fenites may reflect higher fugacities of F during the required enrichment in the source-rock of the Kwandonkaya parental magma than in the fenitizing fluids at Fen.

It is likely that the source for the Kwandonkaya magmas was Fe-rich syenitic fenite, produced by fenitization-type metasomatic alteration of a granulitic lower crust (cf. Morogan & Martin, 1985). The global occurrence of A-type granites suggests that there is no unique source-rock, and that the generation of A-type felsic magmas involves a relatively common process in the lower crust. Although this study points to granulite as the favoured source-rock composition,

the protolith for A-type granites in the Niger-Nigeria Province is most likely not a single or unique lithology, but rather may involve a heterogeneous source, affected by such selective metasomatic enrichment of major and trace elements, which could explain the large range of isotopic ratios documented in literature. In other words, some "ground-preparation" (cf. Abdel-Rahman & Martin 1990) must have occurred prior to the generation of A-type felsic magmas. The study of modern-day analogues of A-type magmatism in an extensional (epeirogenic) regime and of metasomatism accompanying emplacement of mantle-derived rocks in such settings may provide insight into the actual nature of such "ground-preparation".

8.7 Summary and conclusions

1) The petrogenetic modelling of the subvolcanic rock suite indicates that both closed-system fractional crystallization and equilibrium crystallization can adequately account for the observed spectrum of compositions in terms of both major and trace elements. However, the fractional crystallization model is preferred because it requires a lesser extent of crystallization, and it is more effective in enriching incompatible trace elements (e.g., Rb, Li, $\pm$ Sn-W-Nb) in residual melts. Furthermore, the calculated fractional crystallization trend is close to the composition of subvolcanic rocks, and is consistent with petrographic, mineral chemical, and petrochemical evidence described in this thesis.

2) The disparity between the major and trace element models is due to magmatic accessory minerals such as zircon, allanite, and chevkinite, not required in the major element model.

3) Attempts to produce the parental magma of subvolcanic rocks by partial melting models for a variety of lower crustal protoliths were unsuccessful. This study has shown, however, that any reasonable source-rock requires enrichment before partial melting. In the case

of the Kwandonkaya subvolcanic rocks, the source-rock seems to have been a suite of lower crust granulites metasomatically enriched in alkalis, Rb, Ba, Zr, REEs $\pm$ (Fe, Nb, Th, U, Sr, Ni, Cr, V, F, Cl, CO₂), similar to patterns of enrichment observed in classic examples of fenitization. Only the HREE of the source rock are required to be enriched above what is found in fenites. This disparity in HREE pattern indicates that the metasomatizing fluids that altered the source rocks in the Kwandonkaya case likely had higher contents of F. Nonmodal melting following such a metasomatic enrichment is the favoured mode of generation of a suitable parental melt at Kwandonkaya and, by implication, in other plutons of the Nigerian Younger Granite Province.

Chapter 9: Summary, contributions of this study, and recommendations for further studies

9.1 Summary and conclusions

In this study, I have shown that the documented petrochemical and mineral chemical variations in the Kwandonkaya complex are due to an interplay of related factors that include fractional crystallization, differences in the degree of volatile build-up and loss, as well as degrees of subsolidus modifications, and variations in $f(\text{O}_2)$ and temperature at magmatic and postmagmatic stages. The main conclusions, provided at the end of each chapter, are summarized here.

9.1.1 Generation of the parental melt

Attempts to produce the parental magma of subvolcanic rocks by partial melting models for a variety of lower crust protoliths were unsuccessful. In this study, I have shown, however, that any reasonable source-rock requires enrichment before partial melting. In the case of the Kwandonkaya subvolcanic rocks, the source-rock seems to have been a suite of lower-crust granulites metasomatically enriched in alkalis, Rb, Ba, Zr, REEs $\pm$ (Fe, Nb, Th, U, Sr, Ni, Cr, V, F, Cl, CO_2); the phenomenon of enrichment is similar to that observed in classic examples of fenitization. Only the HREE of the source rock are required to be enriched above what is typically found in fenites. This disparity in degree of HREE enrichment indicates that the metasomatizing fluids that altered the source rocks likely had high contents of F. Nonmodal melting (20 - 60%) following such a metasomatic enrichment is the favoured mode of generation of a suitable parental melt at Kwandonkaya and, by implication, in other plutons of the Nigerian Younger Granite Province.

It is envisaged that shear zones, formed during the Pan-African orogeny, were reactivated

during Triassic-Jurassic times in northern Nigeria at the onset of fracturing of the Gondwana plate so as to establish the present-day global plate-tectonic framework. These faults were avenues for decompression and metasomatic modification of the granulitic source-rock. Accompanying underplating by mafic magma and selective trace-element enrichment enabled the generation of a suitable source-rock composition. The derived melt was emplaced into a shallow magma-chamber, where it underwent low-pressure fractional crystallization.

9.1.2 Implications of melting model to NYG and A-type granite magmatism

The above model of genesis proposed for the parental melt of the Kwandonkaya complex has an important bearing on the origin of the parental magma of the NYG complexes, and silica-saturated A-type granites in general. The envisaged model of metasomatism followed by melting is presented in Figures 9.1 and 9.2. Figure 9.1 is a schematic east-to-west section through the lithosphere of the Jos Plateau region indicating a pre-rift stage characterized by: extension, updoming, reactivation of transcurrent faults and shear zones, underplating of mafic magma beneath the lower crust, and metasomatic modification of the granulitic lower crust. Such metasomatic modification, envisaged to have been optimum along the transcurrent faults and shear zones, involved the addition of alkalis (K, Na), Rb, Ba, Sr, Zr, REEs, volatiles (F, Cl, $\pm\text{CO}_2$), and ($\pm\text{Ti}$, Ni, Cr, V, Nb, Th, U).

The occurrence of extension and updoming, as well as the underplating of mafic magma, usually result in the perturbation of lower-crust geotherms and, consequently, in partial melting (Figs. 9.1, 2). The geotherms for the continental crust, as computed by Chapman & Furlong (1992) assuming steady-state, static conditions and conductive transfer of heat, are shown in Figure 9.1. It is envisaged that the underplating of mantle-derived mafic magma in the lower crust, extension, and updoming will shift the geotherms from the steady-state P-T field of the

Fig. 9.1: A schematic lithosphere section (NNE-SSW) through the crust of the Jos Plateau region in northern Nigeria. This section illustrates the pre-rift stage associated with extension, reactivation of transcurrent faults and shear zones, updoming (epeirogenic uplift), underplating of mafic magma beneath the lower crust, and metasomatic modification of the granulitic lower crust. The area of metasomatism (above the underplated mafic magma) most likely coincided with the zone of melting, which is envisaged to have had a maximum extent along the transcurrent faults and shear zones owing to associated higher intensity of alteration and decompression. **Moho** represents the Mohorovičić discontinuity, the crust-mantle boundary.

This diagram was modified from the mantle-metasomatism model of Wooley (1991) and Platt (1996) for the origin of carbonatite-phonolite-syenite complexes associated with rift zones. The model proposed for Kwandonkaya complex (NYG complexes and A-type granites), inferred to have been derived from the lower crust, differs from the rift-related carbonatite-bearing complexes, proposed to have been derived from the upper mantle.

← DOME →
(Jos Plateau, ~150 km wide)

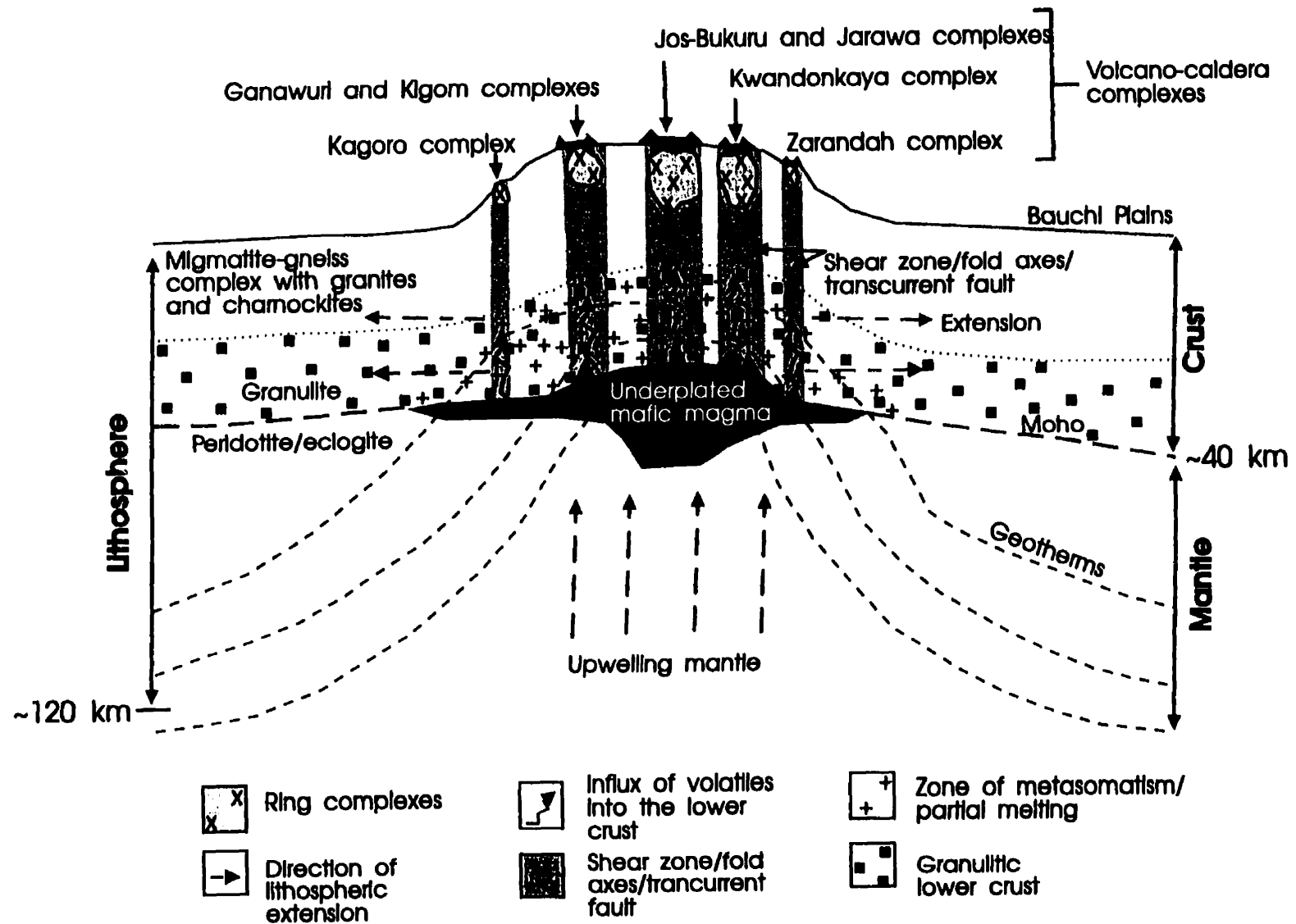
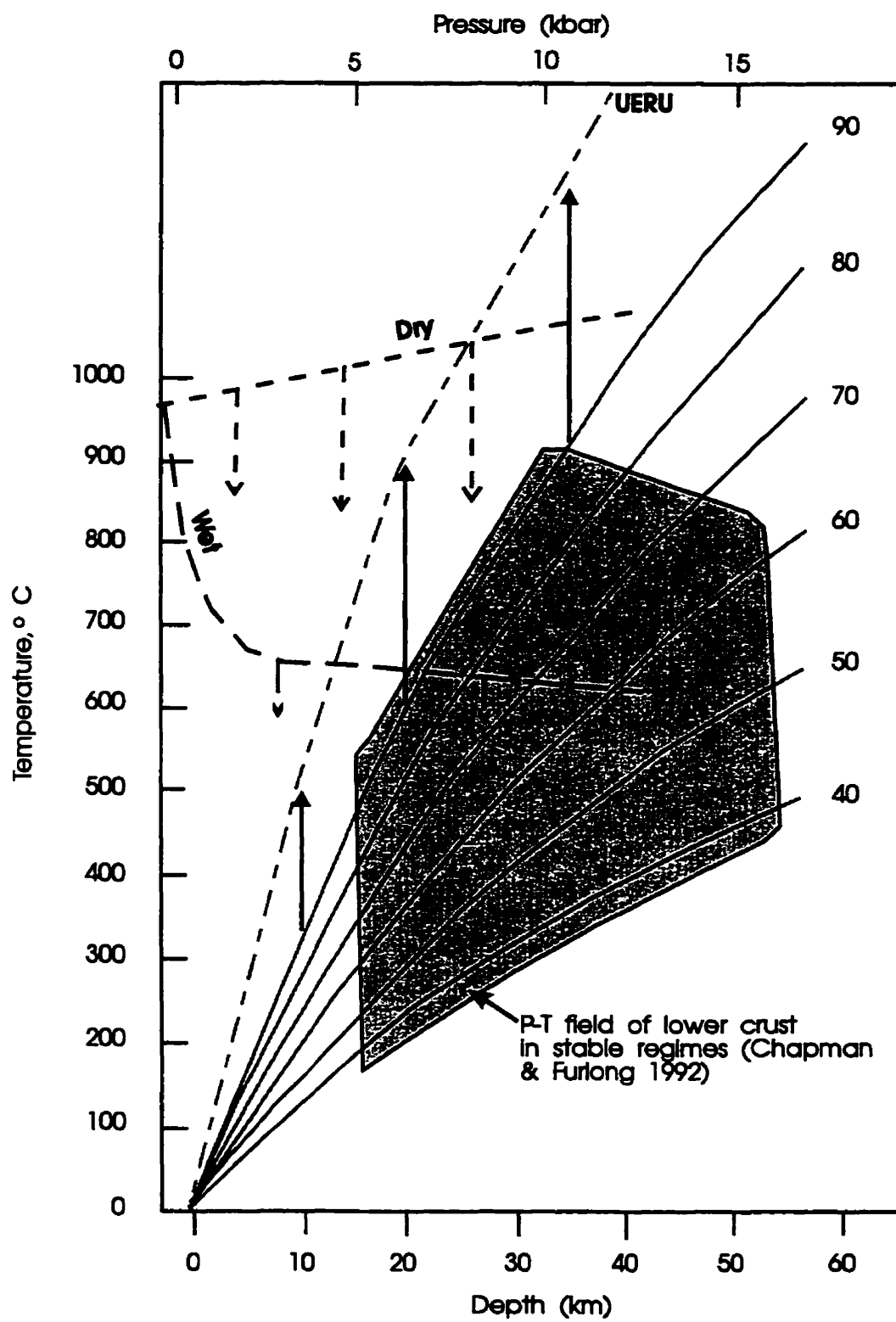


Fig. 9.2: Framework geotherms for the continental crust (modified from Chapman & Furlong 1992). These geotherms were computed by Chapman & Furlong (1992), assuming steady-state, static conditions and conductive transfer of heat. The geotherms, labelled on the right of the curves, represent characteristic heat-flow (q_0) in mW m^{-2} . The shaded area, according to Chapman & Furlong (1992), encloses the steady-state P-T field of the lower crust in stable regimes. It is envisaged that lithospheric extension and updoming, and the underplating by mantle-derived mafic magma, will shift the geotherms to higher temperatures (solid arrow and dashed curve labelled UERU indicate, qualitatively, such effects). The magnitude of such a shift in the geotherms is a function of the ambient geothermal regime, volume and composition of underplated magma, and the extent and rate of lithospheric extension and uplift.

The curves labelled dry and wet show granite solidi plotted from the data of Ebadi & Johannes (1991). The wet solidus depicts water-saturated minimum (granite) melting curve [i.e., $a(\text{H}_2\text{O}) = 1$]. The dry solidus represents the maximum (granite) melting curve [i.e., $a(\text{H}_2\text{O}) = 0$]. Metasomatic alteration will shift the dry melting curve to lower temperatures (dashed arrows indicate, qualitatively, such effects). Such alterations, usually accompanied by increase in F and other volatiles, will also shift the wet solidus to lower temperatures (short arrows). The combined effect of perturbation of lower crustal geotherms and mantle metasomatism is to induce partial melting.



lower crust in stable regions to higher temperatures (Fountain 1989, Wooley 1991, Chapman & Furlong 1992, Platt 1996). The magnitude of such a shift in the geotherms is expected to be a function of the ambient geothermal regime, the extent and rate of lithospheric extension, and the volume and composition of the underplated mantle-derived mafic magma. The combined effect of metasomatism and modification of the lower crustal geotherms as shown in Figure 9.2 is to induce in partial melting that will yield silica-oversaturated (predominantly granitic) melt. In view of the number of variables implied in the model of formation of the parental melt (Fig. 9.1), it is likely that peralkaline centers can be explained by the same overall models of metasomatism followed by melting.

9.1.3 Structural and Magmatic Evolution of the Kwandonkaya complex

The Kwandonkaya complex evolved in a series of migrating magmatic centers at the volcano-caldera and epizonal levels. The subvolcanic rocks of the Kwandonkaya complex are interpreted as relics of intra-caldera volcanism. Transitional rock-types, between truly volcanic and truly plutonic in character, are preserved as rhyolite porphyry along with granite porphyries (degassed subvolcanic intrusions). The first (Buri) and second (Jubga-Ziem) centers were controlled by ring faulting and possibly preceded by cauldron subsidence and intra-caldera volcanism, whereas the third center (Panshanu) has no preserved evidence for caldera formation (owing to the absence of ring faulting and ring-dyke emplacement) (Fig. 2.2b). The stage of intra-caldera volcanism was followed by the emplacement of a variety of porphyries as crystal-rich batches of melt in a series of cone sheets, ring dykes, and composite (in cases zoned) plutons. Currently, the complex is exposed as three overlapping subvolcanic-plutonic igneous centers exposed in their intermediate and root zones (relative to ring dykes) in contact with apical parts of hypabyssal biotite granite intrusions.

From petrographic and petrochemical studies, the enclaves in the Kwandonkaya subvolcanic and plutonic rocks do not correspond to restite produced by melting that yielded the felsic rock-units. The petrographic evidence rather suggests liquid-liquid contacts; mineral chemical evidence indicates various degrees of mixing. The hybridization in most cases occurred initially in mesozonal magma chambers, to produce mixed enclaves, before mingling with their rhyolitic host-rocks. The petrochemical data for the enclaves cannot be modelled as resulting from simple two end-member mixing, because a variety of processes followed the initial mixing and the subsequent stages of mingling: fractionation, phenocryst recycling (i.e., the multistage incorporation of phenocrysts from the felsic host-rocks by enclaves, and *vice versa*), and diffusive exchange between the enclaves and their host rocks. These processes contributed to different extents to a shift in the enclave compositions from trends expected by mixing or fractionation. Overall, the presence of dark enclaves indicates that mafic melt was available throughout the evolution of the Kwandonkaya complex.

The presence of Fe²⁺-rich ferromagnesian phases in the felsic subvolcanic rocks at the initial and final stages of crystallization indicates relatively reduced melts throughout. The crystallization of fayalite and hedenbergite was initiated above or close to the alkali feldspar solvus, since orthoclase microperthite (originally sanidine₂) terminated the growth of both fayalite and hedenbergite. Amphibole crystallization invariably spanned the late-magmatic to subsolidus range (i.e., postdates alkali feldspar crystallization). The limited incorporation of Na in hedenbergite and calcic amphibole indicates lack of substantial alkali enrichment in melts at both early and final stages of crystallization of the subvolcanic rocks; Kwandonkaya thus does not show the late-magmatic peralkaline trend documented in some metaluminous-peraluminous complexes in Nigeria.

The dark mica in subvolcanic and plutonic rocks mostly crystallized mainly from an

interstitial residual melt enriched in volatile components. The variations in the dark mica chemistry include progressive increases in Si, $^{\text{VI}}\text{Al}$ (and ΣAl), Li and F/Cl with decreasing Fe^{2+} and $^{\text{IV}}\text{Al}$. Such variations in mica compositions (from annitic biotite to ferroprotolithionite - siderophyllite - zinnwaldite) suggest an evolution from subvolcanic to plutonic rocks by efficient fractional crystallization. The range of F/FM (0.80 to 1) reflects the highly evolved composition of the suite of rocks, whereas the irregular variations in the F/FM of mica, even in single intrusive units, are due to an interplay of factors that include the degree of oxidation of the residual melt in the subvolcanic rocks and the more pristine plutonic rocks, and the appearance of ilmenite and magnetite before, with, or after the formation of the mica.

The crystallization of early magmatic accessory minerals like chevkinite, zircon, and ilmenite was controlled by saturation in the melt. The late-magmatic inclusions in amphibole and biotite were produced by build-up of ESCs (essential structural constituents) during crystallization or owing to decrease in solubility of ESCs in a differentiated boundary-layer along the crystal-melt interface. In most instances, the residual F-rich melt, from which ferromagnesian minerals (amphibole and biotite) crystallized, coprecipitated the accessory phases in some subvolcanic rocks and biotite granites.

Increases in DI, SiO_2 , and ASI (hence metaluminous to peraluminous transition) of the felsic rock suite resulted from an interplay of factors that included fractional crystallization of sanidine, at water-undersaturated conditions in subvolcanic rocks [$0.3 \leq a(\text{H}_2\text{O}) \leq 0.5$, at 0.5 to 2.5 kbar]; biotite granites crystallized from near-minimum melts that were volatile-enriched to the point of achieving second boiling and developed a peraluminous character by alkali loss at a late magmatic stage. Magmatic evolution by fractional crystallization, from the least evolved to the most evolved felsic rock, is indicated by decreasing ΣFeO , TiO_2 , and Zr with increasing SiO_2 and Rb. The overall similarity in the REE patterns suggests a comagmatic suite, and the progressive

slight depletions in LREE indicates fractionation of LREE-rich phases such as chevkinite and allanite at a magmatic stage. The progressive increases in magnitude of the negative Eu anomaly in pristine subvolcanic rocks and biotite granite indicate the fractional crystallization of a ternary feldspar.

The petrogenetic modelling of the subvolcanic rock suite indicates that both closed-system fractional crystallization and equilibrium crystallization can adequately account for the observed spectrum of compositions in terms of both major and trace elements. The fractional crystallization model is preferred because it requires a lesser extent of crystallization, and it is more effective in enriching incompatible trace elements (e.g., Rb, Li, $\pm$ Sn-W-Nb) in residual melts. Furthermore, the calculated fractional crystallization trend is close to the composition of subvolcanic rocks, and is consistent with petrographic, mineral chemical, and petrochemical evidence described in this thesis. The felsic subvolcanic rock suite possibly could have evolved by 29% Rayleigh fractionation of liquidus phases from the least evolved to the most evolved composition. The cumulate has a composition: sanidine_{ss}: 88%, fayalite: 6%, hedenbergite: 5%, ilmenite: 1%. Therefore, fractional crystallization of alkali feldspar predominantly accounts for the major-element evolution of the subvolcanic rocks. Trace-element evolution requires 50% fractionation of the same phases. The disparity between these two models (i.e., the trace and major elements) is due mainly to the role of zircon, chevkinite, and allanite, not called upon in the major-element model.

9.1.3 Postmagmatic evolution

Subsolidus assemblages in place of the magmatic ferromagnesian phases consistently indicate a shift of $f(\text{O}_2)$ conditions to above QFM, and an increase in $a(\text{H}_2\text{O})$, and F/Cl with decreasing temperature. Variations of Fe/Mg of fayalite, hedenbergite, and amphibole are not

sequential with respect to the relative age of emplacement and degree of evolution of their host rocks. Factors other than fractionation, namely, different extents of volatile build-up and loss, and different degrees of subsolidus equilibration, thus also influenced the composition of the ferromagnesian minerals in the subvolcanic rocks of the complex.

Despite the fact that most subvolcanic rocks contain alkali feldspar apparently devoid of exsolution lamellae, the data show that all rocks in the Kwandonkaya complex, without exception, have been equilibrated to various extents in the subsolidus region. Small-volume intrusions, such as the Buri ring dyke and cone sheets, which underwent high undercoolings and were relatively volatile-poor, are characterized by orthoclase without microcline. The rock types in which orthoclase occurs as the only K-feldspar indicates the absence of extensive reordering due to the water-undersaturated character of subvolcanic rocks, which resulted in limited rock-fluid interactions in the subsolidus, and the presence of Ca in the primary feldspar. On the other hand, subvolcanic rocks in the Jubga-Ziem ring dyke contain orthoclase + microcline in their phenocryst-matrix pairs, possibly because of the larger size of the intrusive complex, and thus slower cooling, lower An contents, and higher volatile contents. Subsolidus re-equilibration was likely enhanced by interaction with fluids exsolved from the (younger) biotite granite. These rock types preserve intermediate stages of the monoclinic-to-triclinic inversion accompanying the Al-Si order during the transition: sanidine_{ss} ($\rightarrow$ orthoclase) + albite $\rightarrow$ microcline + albite. The larger intrusive masses of biotite granite contain mainly microcline, which reflects the fact that these units are relatively Ca-poor, represent larger intrusive bodies that cooled down at a relatively slower rate, and were characterized by relatively more significant build-up of volatiles, which catalyzed the ordering reaction. Overall, the textural variants of biotite granite display consistent similarity in their t_{10} , N_{Or} , and Δ , both at the margins and central part of the pluton. This consistently suggests that the composite biotite granite experienced a rapid (uniform) rate of

cooling and extent of postmagmatic equilibration, except in localized apical areas that experienced volatile loss to yield drusy biotite granite. It is interesting to note that end-member microcline does not occur here; in spite of evidence that exsolution of albite has invariably occurred, the Al-Si ordering has not reached completion in the time available. Furthermore, there is no evidence that the earlier centers (Buri and Jubga-Ziem) were heated up by the youngest one, Panshanu, to result in the resetting of the Al-Si order of earlier rock units. Thus, all three intrusive centers were penecontemporaneous. The K-feldspar in the subvolcanic and plutonic rocks of the Kwandonkaya complex were mostly equilibrated at compositions defined by the strain-free solvus: the inferred temperatures of closure of rock types bearing orthoclase (from $0.9 \leq N_{Or} \leq 1$) are between 300 and 450°C, whereas those bearing only microcline range from 200 to 450°C.

The inferred halogen fugacity ratios [$\text{Log}(f_{\text{HF}}/f_{\text{HCl}})$] increase in the sequence: subvolcanic rocks through pristine plutonic rocks to altered biotite granites. The variation in the halogen fugacities in the various rock-units reflects a dropping temperature of equilibration with fluids evolved to various extents, the inherent variation in mica chemistry (particularly the X_{Fe} or X_{Mg}) that affects the degree of accommodation of the halogens, and various extents of volatile loss. Increases in inferred concentration of Li, and ratio of halogen fugacities ($\text{Log } f_{\text{HF}}/f_{\text{HCl}}$) of fluids, from subvolcanic to plutonic rocks, defines an important link with the observed alterations and mineralization of biotite granites. Such alterations and mineralization are absent in the low-Li, low- f_{HF} subvolcanic rocks. The composition of the dark mica in drusy as well as albitized biotite granite indicates late re-equilibration with F-rich fluids. Except in the drusy biotite granite, the pale-coloured trilithionite in biotite granites, albitized biotite granite and greisens is entirely secondary. In both cases, the pale mica is predicted to be a mixed structure (i.e., an intermediate composition between dioctahedral lithian muscovite and trioctahedral lepidolite), generally enriched in Li, Al, and Si, and inferred to have equilibrated with fluids highly enriched in F

relative to Cl.

This study has also shown that the accessory minerals like chevkinite, zircon, and ilmenite crystallized early, and were therefore controlled by saturation in the melt. The late magmatic inclusions in amphibole and biotite (e.g., allanite, ilmenite, and zircon) were apparently produced by build-up of ESCs during crystallization or due to decrease in solubility of ESCs in a differentiated boundary-layer along crystal-melt interface. In some samples of porphyry and biotite granite, the accessory minerals were produced either by subsolidus replacement or by hydrothermal vein-filling. In the altered variants of biotite granite, the accessory and ore mineral suite was produced by either replacement or by crystallization from hydrothermal fluids. The hydrothermally affected units are not unusually enriched in most of the trace elements and REE. Therefore, the accessory and ore-mineral suite in these rock types likely resulted from local dissolution and redeposition.

The petrographic, mineral chemical and petrochemical features illustrate well the fact that the characteristics of the biotite granites have been modified from what can be considered as a magmatic mineral assemblage, even in the most pristine samples. There is also a strong indication that the magma that crystallized to form biotite granite was metaluminous. Although there are similarities in patterns of trace element enrichment of topaz-bearing granites in the Kwandonkaya complex and the low-phosphorus type topaz granites of Taylor (1992), in detail, the topaz-bearing granites in Kwandonkaya have slightly lower Rb, Pb, Ba, and HREE compared to his low-P topaz granites. Such differences are possibly due to a higher degree of volatile loss and rock-fluid interactions in the biotite granite of the Kwandonkaya complex. Postmagmatic alteration is indicated by the apparent mobility of Ca, Sr, Ba, Nb, Y, Th and U in the whole suite, and alkali loss is suggested by the slight decrease in $\Sigma(\text{Na}_2\text{O} + \text{K}_2\text{O})$ with increasing bulk SiO_2 . The granitic melt that led to the biotite granite was volatile-enriched to the point of achieving second boiling

and developed a peraluminous character by alkali loss, first at a late magmatic stage; then, the aluminosity (ASI) was enhanced locally by hydrothermal leaching of alkalis. In the greisens, increased concentrations of Ti, ΣFe , Mn, and Pb point to the mobilization of these elements in the later-stage acidic fluids. Later enrichments in HREE are related to build-up in F and other volatile constituents in altered variants of biotite granite, whereas a more pronounced Eu anomaly in such altered biotite granite indicates selective mobility of Eu associated with loss of H_2 and accompanying oxidation during degassing.

9.2 Contributions of this study

- 1) In this study, modifications are proposed to the geological map of Kwandonkaya complex. This is with respect to the Buri ring dyke, hitherto not included in other studies. It is hereby proposed that textural and mineralogical criteria be used in naming rock types, rather than the "type locality" names. This will enable comparison with rock types in better known complexes such as Jos-Bukuru, Tibchi, and Ririwai.
- 2) This study of the ferromagnesian minerals in the Kwandonkaya complex represents, to my knowledge, the most extensive study of an individual ring complex in the NYG province.
- 3) This study presents the first attempt to infer halogen fugacity ratios in the fluid from mica chemistry. These results provide a framework for future studies of fluid inclusions and mass balance calculations.
- 4) The structural state of alkali feldspar and the inferred variations in postmagmatic re-equilibration have been studied more thoroughly in the Kwandonkaya complex than in any other of the NYG complexes, or any other subvolcanic-plutonic A-type granitic complex.
- 5) The occurrence of chevkinite, allanite, and a variety of REE-bearing minerals in a NYG complex is documented for the first time by this study. This results set the stage for future studies

on specific minerals of petrogenetic or economic importance.

- 6) This study reports, for the first time, the occurrence of enclaves and magma mixing in the Kwandonkaya complex. This thesis contains the most detailed study of the field, petrographic, and mineral chemical characteristics as well as petrochemical studies of enclaves in NYG complexes.
- 7) For the first time in the NYG complexes, attempts are made to quantify the extent of fractionation required to produce the spectrum of compositions documented.
- 8) This study also presents the first attempt to produce a quantitative model on the nature of the source-rock and the mode of generation of the parental melt. A model of metasomatic enrichment of granulitic lower crust rocks followed by 20 to 60% nonmodal (partial) melting is proposed.

9.3. Recommendations for further studies.

The nature and effect of metasomatism at the source (cf. Bowden et al. 1987, Bonin 1990) will require further study so as to shed further light on mantle-lower crust interaction in regions containing A-type granite provinces. The studies of the Cenozoic alkaline volcanic suites in the area containing A-type granite complexes in Nigeria and adjacent Benue rift (although unrelated to NYG magmatism) may provide direct information on the nature of lower crust and mantle beneath the NYG province. So far, these rocks have received little attention, although it is known that the volcanic suite carries xenoliths of lherzolite, pyroxenite, and diorite, as well as xenocrysts of garnet and zircon (Wright 1975). The detailed studies of these xenoliths and inclusions may provide precious information on the possible source-rock compositions for the NYG complexes, and on the importance of metasomatic "fertilization" in the lower crust.

The nature of the lower crust, metasomatized or not, needs to be better documented isotopically. Only once this documentation is available will it be possible to properly interpret the isotope geochemistry of the Kwandonkaya suite and of other suites in the belt, in particular to

determine the respective roles of crust and mantle.

Studies involving Sn and halogen contents of the suite of rocks, along with fluid inclusion investigations, are required to better understand alteration events and mineralization. The documentation of topaz and beryl occurrences and characterization of their fluid inclusions are being currently pursued.

References

- ABDEL-RAHMAN, A-F.M. & MARTIN, R.F. 1990: The Mount Gharib A-type granite, Nubian Shield: petrogenesis and role of metasomatism at the source. *Contrib. Mineral. Petrol.* **104**, 173-183.
- AINES, R.D. & ROSSMAN, G.R. 1986: Relationships between radiation damage and trace water in zircon, quartz, and topaz. *Am. Mineral.* **71**, 1186-1193.
- AJAKAIYE, D.E. 1976: A gravity survey over the Nigerian Younger Granite Province. In: *Geology of Nigeria* (C.A. Kogbe, ed.). Elizabethan Publishing Company, Lagos, Nigeria (207-224).
- _____, HALL, D.H. & MILLAR, T. 1983: Aeromagnetic anomalies across the Nigerian Younger Granite province. *Abstr. 1409. IAGA Bull, 28th Gen. Assemb., Hamburg*, **48**, p14
- AJIBADE, A.C., WOAKES, M. & RAHAMAN, M.A. 1987: Proterozoic crustal development in the Pan-African regime of Nigeria. In: *Proterozoic lithospheric evolution* (A. Kröner, ed.). (259-271).
- ALLÈGRE, C.J. & MINSTER, J.F. 1978: Quantitative models of trace element behaviour in magmatic processes. *Earth Planet. Sci. Lett.* **38**, 1-25.
- _____, PROVOST, A. & JAUPERT, C. 1981: Oscillatory zoning: a pathological case of crystal growth. *Nature* **294**, 223-228.

- ALLEN, C.M. 1991: Local equilibrium of mafic enclaves and granitoids of the Turtle pluton, southeast California: mineral, chemical, and isotopic evidence. *Am. Mineral.* **76**, 574-588.
- ANDERSON, A.T. 1976: Magma mixing: petrological process and volcanological tool. *J. Volcan. Geotherm. Res.* **1**, 3-33.
- ANDERSON, J.L., CULLERS, R.L. & VAN SCHMUS, W. R. 1980: Anorogenic metaluminous and peraluminous granite plutonism in the mid-Proterozoic of Wisconsin, U.S.A. *Contrib. Mineral. Petrol.* **74**, 311-328.
- ANFIGOLOV, V.W., BUSHLYAKOV, V.A., VILISOV, V.A., & BRAGINA, G.I. 1977: Distribution of fluorine between coexisting biotite and amphibole and granitic melt at 780°C and 1000 atm. pressure. *Geochem.* **14**(2), 95-98.
- BACON, C.R. 1989: Crystallization of accessory phases in magmas by local saturation adjacent to phenocrysts. *Geochim. Cosmochim. Acta* **53**, 1055-1066.
- BADEJOKO, T.A. 1977: *The composition and structural state of K-feldspar and their bearing on the evolution of Younger Granites of central Nigeria*. Unpubl. Ph.D. Thesis, Univ. of Ibadan, Nigeria
- _____, 1984: Correlations between microstructures, K-feldspar triclinicity and trace element geochemistry in stanniferous and barren granites, northern Nigeria. *Lithos* **17**, 259-271
- _____, 1986: The petrogenesis of the Younger Granites of Nigeria. *J. African Earth Sci.* **5**, 233-242.
- BAILEY, D.K. 1987: Mantle metasomatism - perspective and prospects. In: **Alkaline igneous rocks** (J.G. Fitton & B.J.G. Upton, eds.). *Geol Soc. Am., Spec. Publ.* **30**, 1-13.
- BAILEY, S.W. 1984a: Classification and structures of the micas. In: **Micas** (S.W. Bailey, ed.). *Rev. Mineral.* **13**, 1-12.
- _____, 1984b: Crystal chemistry of the true micas. In: **Micas** (S.W. Bailey, ed.). *Rev. Mineral.*

- BAIN, A.D.N. & RUSS, W. 1923: Kwandonkaya complex. In: **The northern tinfields of Bauchi province** (J.D. Falconer & C. Raeburn, eds.). *Geol. Surv. Nig., Bull.* **4**.
- BAKER, D.R. 1990: Chemical interdiffusion of dacite and rhyolite: anhydrous measurements at 1 atm and 10 kbar, application of transition state theory, and diffusion of zoned magma chambers. *Contrib. Mineral. Petrol.* **104**, 407-423.
- _____ 1991: Interdiffusion of hydrous dacitic and rhyolitic melts and the efficacy of rhyolite contamination of dacite enclaves. *Contrib. Mineral. Petrol.* **106**, 462-473.
- _____ & BOSSANYI, H. 1994: The combined effect of F and H₂O on interdiffusion between peralkaline dacitic and rhyolitic melts. *Contrib. Mineral. Petrol.* **117**, 203-214.
- BARBARIN, B. & DIDIER, J. 1991: Review of the main hypotheses proposed for the genesis and evolution of the mafic microgranular enclaves. In: **Enclaves and granite petrology**. (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (367-373).
- _____, _____ & GOURGAUD, A. 1984: Magma mixing: origin of intermediate rocks and "enclaves" from volcanism to plutonism. *Phys. Earth Planet. Interiors* **35**, 63-76.
- BARKER, F., WONES, D.R., SHARPE, W.N. & DESBOROUGH, G.A. 1975: The Pikes Peak batholith, Colorado Front Range, and a model for the origin of the gabbro-anorthosite-potassic granite suite. *Precamb. Res.* **2**, 97-160.
- BEARD, J.S., LOFGREN, G.E., SINHA, A.K. & TOLLO, R.P. 1992: Experimental constraints on the origin of Late Proterozoic granitoids in the Central Appalachians. *Eos, Trans. Am. Geophys. Union* **73**(4), 347 (Abstr.).
- BÉDARD, J.H. 1989: The Mégantic complex: the northernmost intrusion of the White Mountain Magma series. **GAC/MAC Fieldtrip B3: Alkaline and calc-alkaline complexes of southern Québec**. *GAC/MAC Field trip guidebook*, 75-106.

- BÉDARD, J.H., LUDDEN, J.N., & FRANCIS, D.H. 1987: The Mégantic intrusive complex, Québec: a study of derivation of silica-oversaturated anorogenic magmas of alkaline affinity. *J. Petrol.* **28**, 355-388.
- BENNETT, J.N. 1981: *The petrology and mineral chemistry of the Shira ring complex, northern Nigeria*. Unpubl. Ph.D. thesis, University of St Andrews, St. Andrews, U. K.
- _____, TURNER, D.C., IKE, E.C. & BOWDEN, P. 1984: The geology of some northern Nigerian anorogenic ring complexes. *Overseas Geol. & Mineral. Res.* **61**, 65p.
- BERGEN, M.J. 1984: Perrierite in siliceous lavas from Mt. Amiata, central Italy. *Mineral. Mag.* **48**, 553-556.
- BLACK, R. 1965: Sur la signification pétrogénétique de la découverte d'anorthosites associés aux complexes annulaires subvolcaniques du Niger. *C. R. Acad. Sci. Fr.* **260**, 5829-5832.
- _____, LAMEYRE, J. & BONIN, B. 1985: The structural setting of alkaline ring complexes. *J. Afr. Earth Sci.* **3**, 5-16.
- BLAKE, S. & CAMPBELL, I.H. 1986: The dynamics of magma mixing during flow in volcanic conduits. *Contrib. Mineral. Petrol.* **94**, 72-81.
- BONIN, B. 1986: **Ring complex granites and anorogenic magmatism**. Elsevier, New York. 188p.
- _____. 1990: From orogenic to anorogenic settings: evolution of granitoid suites after major orogenesis. *Geol. J.*, **25**, 261-271.
- _____. 1991: The enclaves of alkaline anorogenic granites: an overview. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (179-189).
- _____ & GIRET, A. 1985a: Clinopyroxene compositional trends in oversaturated and undersaturated alkaline ring complexes. *J. Afr. Earth Sci.* **3**, 175-183.
- _____ & _____ 1985b: Contrasting roles of rock-forming minerals in alkaline ring complexes.

J. Afr. Earth Sci. **3**, 41-49.

BORLEY, G.D. 1963: Amphibole from the Younger Granites of Nigeria. Part 1, Chemical classification. *Mineral. Mag.* **33**, 358-376.

_____. 1976: Ferromagnesian mineralogy and temperatures of formation of the Younger Granites of Nigeria. In: **Geology of Nigeria** (Kogbe, C.A., ed.). Elizabethan Publ. Co., Lagos, Nigeria (159-176).

BOWDEN, P. 1982: Magmatic evolution and mineralisation in the Nigerian Younger Granite province. In: **Metallization Associated with Acid Magmatism** (A.M. Evans, ed.). John Wiley, London (51-62).

_____. 1985: The geochemistry and mineralisation of alkaline ring complexes in Africa (a review). *J. Afr. Earth Sci.* **3**, 17-29.

_____. & KINNAIRD, J.A. 1992. A-type granites in Africa. *EOS, Trans. Amer. Geophys. Union* **73**(14), 348.

_____, BENNETT, J.N., KINNAIRD, J.A., WHITLEY, J.E., ABAA, S.I. & HADZIGEORIOU-STAVRAKIS, P.K. 1981: Uranium in the Niger-Nigeria Younger Granite province. *Mineral. Mag.* **44**, 379-389.

_____, BLACK, R., MARTIN, R.F., IKE, E.C., KINNAIRD, J.A. & BATCHELOR, R.A. 1987: Niger-Nigerian alkaline ring complexes: a classic example of African anorogenic mid-plate magmatism. In: **Alkaline Igneous Rocks** (J.G. Fitton & B.J.G. Upton, eds.). *Geol. Soc. Am., Spec. Publ.* **30**, 357-379.

_____, KINNAIRD, J.A., ABAA, S.I., IKE, E.C. & TURAKI, U.M. 1984: Geology and mineralization of the Nigerian anorogenic ring complexes. *Geol. Jb.* **B56**, 3-65.

_____, WHITLEY, J.E. & VAN BREEMEN, O. 1976a: Recent geochemical studies on the Younger Granites of northern Nigeria. In: **Geology of Nigeria** (C.A. Kogbe, ed.).

- Elizabethan Publ. Comp., Lagos, Nigeria (177-193).
- _____, VAN BREEMEN, O., HUTCHINSON, J. & TURNER, D.C. 1976b: Palaeozoic and Mesozoic age trends for some ring complexes in Niger and Nigeria. *Nature* **259**, 297-299.
- BOWEN, N.L. & SCHAIRER, J.F. 1935: Grunerite from Rockport, Massachusetts, and a series of synthetic fluor-amphibole. *Am. Mineral.* **47**, 78-82.
- BRANDEIS, G. & JAUPART, C. 1987: Crystal sizes in intrusions of different dimensions: constraints on the cooling regime and crystallization kinetics. In: **Magmatic processes: physicochemical principles** (B.O. Mysen, ed.). *The Geochem. Soc., Spec. Publ.* **1**, 307-318.
- BROOKS, C.K., HENDERSON, P. & RØNSBØ, J.G. 1981: Rare-earth partition between allanite and glass in the obsidian of Sandy Braes, Northern Ireland. *Mineral. Mag.* **44**, 157-160.
- BROWN, G.C. & BOWDEN, P. 1973: Experimental studies concerning the genesis of the Nigerian Younger Granites. *Contrib. Mineral. Petrol.* **40**, 131-139.
- BROWN G.E. Jr. 1982: Olivines and silicate spinels. In: **Orthosilicates** (P.H. Ribbe, ed.). *Rev. Mineral.* **5**, 275-281.
- BROWN, W.L. & PARSONS, I. 1989: Alkali feldspars: ordering rates, phase transformations and behaviour diagrams for igneous rocks. *Min. Mag.* **53**, 25-42.
- BROWN W.L, MOREAU, C. & DEMAÏFFE, D. 1989: An anorthosite suite in a ring complex: crystallization and emplacement of an anorogenic type from Aborontok, Aïr, Niger. *J. Petrol.* **30**, 1501-1540.
- BUCHANAN, M.S. 1971: The Kwandonkaya complex. In: **The Geology of Jos Plateau: Younger Granite complexes** (M.S. Buchanan, W.N. MacLeod, D.C. Turner, R. Black, & N.G. Berridge, eds.). *Bull. Geol. Soc. Nigeria* **32(2)**, 35-40.
- CAMERON, M. & PAPIKE, J.J. 1980: Crystal chemistry of silicate pyroxenes. In: **Pyroxenes**

(C.T. Prewitt, ed.). *Rev. Mineral.* **17**, 5-92.

CANTAGREL, J.-M., DIDIER, J., & GOURGAUD, A. 1984. Magma mixing: origin of intermediate rocks and "enclaves" from volcanism to plutonism. *Phys. Earth & Planet. Inter.* **35**, 63-76.

CARMICHAEL, I.S.E., TURNER, F.J., & VERHOOGEN, J. 1974. **Igneous Petrology**. McGraw Hill, New York.

CARSTENS, H. 1983: Simultaneous crystallization of quartz-feldspar intergrowths from granitic magmas. *Geology* **11**, 339-341.

CASTRO, A., DE LA ROSA J.D. & STEPHENS, W.E. 1990: Magma mixing in the subvolcanic environment: petrology of the Gerena interaction zone near Seville, Spain. *Contrib. Mineral. Petrol.* **105**, 9-26.

_____, MORENO-VENTAS, I. & DE LA ROSA, J.D. 1991: Multistage crystallization of tonalitic enclaves in granitoid rocks (Hercynian belt, Spain): implications for magma mixing. *Geol. Rund.* **80**, 109-120.

CATHELINEAU, M. 1987: U, Th, and U mobility during albitization and quartz dissolution in granitoids: evidence from southeast French Massif Central. *Bull. Minéral.* **110**, 249-259.

CAWTHORN, R.G. 1976: Some controls on igneous amphibole compositions. *Geochim. Cosmochim. Acta* **40**, 1319-1328.

ČERNÝ, P. & BURT, D.M. 1984: Paragenesis, crystallochemical characteristics, and geochemical evolution of micas in granite pegmatites. In: **Micas** (S. W. Bailey, ed.), *Rev. Mineral.* **13**, 257-297.

CHAPMAN, D. S. & FURLONG, K. P. 1992. Thermal state of the continental lower crust. In: **Continental lower crust** (D. M. Fountain, R. Arculus, & R. W. Kay, eds.). Elsevier, Amsterdam (179-199).

- CHAPPELL, B.W., WHITE, A.J.R. & WYBORN, D. 1987: The importance of residual source material (restite) in granite petrogenesis. *J. Petrol.* **28**, 1111-1138.
- _____ & WHITE, A.J.R. 1991: Restite enclaves and the restite model. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (375-281).
- CHRISTIANSEN, E.H., SHERIDAN, M.F. & BURT, D.M. 1986: The geology and geochemistry of Cenozoic topaz rhyolites from Western United States. *Geol. Soc. Am., Spec. Pap.* **205**, 82p.
- CLEMENS, J.D., WALL, V.J., & CLARKE, B.D. 1983: Models for granitoid evolution and source compositions: restite - "R. I. P.". *Geol. Soc. Am., Abstr. Prog.* **9**, 178-179.
- _____, HOLLOWAY, J.R. & WHITE, A.J.R. 1986: Origin of A-type granites: experimental constraints. *Am. Mineral.* **71**, 317-324.
- COLLINS, W.J., BEAMS, S.D., WHITE, A.S.R. & CHAPPELL, B.W. 1982: Nature and origin of A-type granites with particular reference to southeastern Australia. *Contrib. Mineral. Petrol.* **80**, 189-200.
- CONDIE, K.C. 1991: Precambrian granulites and anorogenic granites: are they related? *Precamb. Res.* 161-172.
- CREASER, R.A., PRICE, R.C. & WORMALD, R.J. 1991: A-type granites revisited: assessment of a residual source model. *Geology* **19**, 163-166.
- CREASY, J.W. & EBY, G.N. 1993: Ring dyke and plutons: a deeper view of calderas as illustrated by the White Mountain igneous province, New Hampshire. In: **Field Trip Guidebook for the northeastern United States** (J.T. Cheney & J.C. Hepburn, eds.). 1993 Boston GSA, vol. 1 (N1-N25).
- CRISCI, G.M., DE FRANCESCO, A.M., MAZZOULI, R., POLI, G. & STANZONE, D. 1989: Geochemistry of Recent volcanics of Ischia island, Italy: evidence for fractional

- crystallization and magma mixing. *Chem. Geol.* **78**, 15-33.
- CZAMANSKE, G.K. 1988: Interplay between REE-bearing accessory minerals and silicic magma, Questa, New Mexico, U.S.A. Fifth international symposium on tin/tungsten granites in southeast Asia and western Pacific (Matsue). *IGCP 220, Abstr. with Prog.*, 25-29.
- _____ & WONES, D.R. 1973: Oxidation during magmatic differentiation, Finnmarka complex, Oslo area, Norway. 2. The mafic silicates. *J. Petrol.* **14**, 349-380.
- _____, ISHIHARA, S., & ATKIN, S. A. 1981: Chemistry of rock-forming minerals of the Cretaceous - Paleocene batholith in southwestern Japan and implications for magma genesis. *J. G. R.* **86**, 10431-10470.
- DADA, S.S., LANCELOT, J.R. & BRIQUEU, L. 1991: Age and origin of the annular charnockitic complex at Toro, northern Nigeria: U-Pb and Rb-Sr evidence. *J. Afr. Earth Sci.* **9(2)**, 227-234.
- DE ALBUQUERQUE, C.A.R. 1973: Geochemistry of biotites from granitic batholiths, northern Portugal. *Geochim. Cosmochim. Acta* **37**, 1799-1802.
- DICKIN, A.P., HALLIDAY, A.N. & BOWDEN, P. 1983: A lead, strontium, and neodymium isotope study of the Nigerian Mesozoic ring complexes. *Abs. EUG*, Strasbourg
- DIDIER, J. 1973: **Granites and their enclaves: The bearing of enclaves on the origin of granites.** Elsevier, Amsterdam, The Netherlands.
- _____ 1991: The various types of enclaves in the Hercynian granitoids of the Massif Central, France. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (47-61).
- _____ & BARBARIN, B. 1991: The different types of enclaves in granites - nomenclature. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (19-23).

- DINGWELL, D.B. 1988: The structure and properties of fluorine-rich magmas: a review of experimental studies: In: **Recent Advances in the Geology of Granite-related Mineral Deposits** (R.P. Taylor & D.F. Strong, eds.). *Can. Inst. Min. Metall., Spec. vol. 39*, 1-12.
- DODGE, F.C. & MOORE, J.G. 1968: Occurrence and composition of biotite from granitic rocks of the Sierra Nevada batholith, California. *U.S. Geol. Surv., Prof. Pap.*, **600-B**, 6-10.
- _____, SMITH, V.C. & MAYS, R.E. 1969: Biotites from granitic rocks of the Sierra Nevada batholith, California. *J. Petrol.* **10**, 250-271.
- DROOP, G.T.R. 1987: A general equation for estimating Fe^{3+} concentrations for ferromagnesian silicates and oxides from microprobe analyses, using stoichiometric criteria. *Mineral. Mag.* **51**, 431-435.
- DRYSDALL, A.R., JACKSON, N.J., RAMSAY, C.R., DOUCH, C.J. & HACKETT, D. 1984: Rare element mineralization related to Precambrian alkali granites in the Arabian Shield. *Econ. Geol.* **79**, 1366-1377.
- EBADI, A. & JOHANNES, W. (1991): Beginning of melting and composition of first melts in the system quartz-albite-orthoclase- H_2O - CO_2 . *Contrib. Mineral. Petrol.* **106**, 286-295.
- EBORALL, M.I. 1976: Intermediate rocks from the Older Granite complex of the Bauchi area, northern Nigeria. In: **Geology of Nigeria** (C. A. Kogbe, ed.). Elizabethan Publishing Co., Lagos, Nigeria (65-74).
- EBY, N.G. 1990: The A-type granitoids: A review of their occurrence and speculation on their petrogenesis. *Lithos* **26**, 115-134.
- _____, KRUEGER, H.W. & CREASEY, J.W. 1992: Geology, geochronology, and geochemistry of the White Mountain batholith, New Hampshire. *Geol. Soc. Am., Spec. Pap.* **268**, 379-397.
- EI BOUSEILY, A.M. & EI SOKKARY, A.A. 1975: The relation between Rb, Ba, and Sr in

- granitic rocks. *Chem. Geol.* **16**, 207-219.
- ELLISON, A. J. & HESS, P. G. 1986: Solution behaviour of +4 cations in high silica melts: petrologic and geochemical implications. *Contrib. Mineral. Petrol.* **94**, 343-351.
- EUGSTER, H.P. & WONES, D.R. 1962: Stability relations of the ferruginous biotite, annite. *J. Petrol.* **3**, 82-125.
- EXLEY, R.A. 1980: Microprobe studies of REE-rich accessory minerals: implications for Skye granite petrogenesis and REE mobility in hydrothermal systems. *Earth Planet. Sci. Lett.* **48**, 97-100.
- FABRIÈS, J. & ROCCI, G. 1965: Le massif granitique du Tarraouadji (Républic du Niger): étude et signification pétrogénétique des principaux minéraux. *Bull. Soc. Franc., Minéral. Cristallogr.* **88**, 319-340.
- FALCONER, J.D. 1911: **The geology and geography of Northern Nigeria**. MacMillan, London.
- FARMER, G. L. 1992: Magmas as tracers of lower crustal composition: an isotopic approach. In: **Continental lower crust** (D.M. Fountain, R. Arculus, and R. W. Kay, eds.). Elsevier, Amsterdam (363-390).
- FENN, P.M. 1977: The nucleation and growth of alkali feldspars from hydrous melts. *Can. Mineral.* **15**, 135-161.
- FERGUSON, A.K. 1978: The crystallization of pyroxenes and amphiboles in some alkaline rocks and the presence of a pyroxene compositional gap. *Contrib. Mineral. Petrol.* **67**, 11-15.
- FERNANDEZ, A.N. & BARBARIN, B. 1991: Relative rheology of coeval mafic and felsic magmas: nature of resulting interaction processes and shape and mineral fabrics of mafic microgranular enclaves. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (263-275).
- FERRÉ, E., GLEIZES, G., BOUCHEZ, J. L., & NNABO, P. N. 1995: Internal fabric and strike-

- slip emplacement of the Pan-African granite of Solli Hills, northern Nigeria. *Tectonics* **14**, 1205-1219
- FERRY, J.M. 1985: Hydrothermal alteration of Tertiary igneous rocks from the Isle of Skye, northwest Scotland. *Contrib. Mineral. Petrol.* **91**, 283-304.
- FOUNTAIN, D. M. 1989: Growth and modification of the lower continental crust in extended terrains: the role of extension and magmatic underplating. In: **Properties of the Earth's lower crust** (R.I. Mereu, Meuller, S., & Fountain D.M., eds.). *Geophysical Monograph* **51**, IUGG v. 6 (287-299).
- FOSTER, M.D. 1960: Interpretation of the composition of lithium micas. *U.S. Geol. Surv., Prof. Pap.* **354-E**, 115-147.
- GILBERT, M.C. 1966: Stability and synthesis of the hornblende, ferropargasite. *Am. J. Sci.* **264**, 698-742.
- _____, HELTZ, R.T., POPP, R.K. & SPEAR, F.S. 1982: Experimental studies of amphibole stability. In: **Amphiboles: Petrology and experimental phase relations** (D.R. Veblen & P.H. Ribbe, eds.). *Rev. Mineral.* **9B**, 229-353.
- GIRET, A., BONIN, B. & LEGER, J.-M. 1980: Amphibole compositional trends in oversaturated and undersaturated alkaline plutonic ring-complexes. *Can. Mineral.* **18**, 481-495.
- GOURGAUD, A. 1991: Comagmatic enclaves in lavas from the Monte-Dore composite volcano, Massif Central, France. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (221-233).
- _____ & MAURY, R.C. 1984: Magma mixing in alkaline series: an example from Sancy volcano (Mont-Dore, Massif Central, France). *Bull. Volcanol., Spec. Issue on magma mixing* (M.P. Semet & N. Vatin-Perignon, eds.), **47-4**, 827-847.
- GREEN, T.H. & PEARSON, N.J. 1986: Rare-earth element partitioning between sphene and

- coexisting silicate liquid at high pressure and temperature. *Chem. Geol.* **55**, 105-119.
- _____ & _____ 1988: Experimental crystallization of chevkinite/perrierite from REE-enriched silicate liquids at high pressure and temperature. *Mineral. Mag.* **52**, 113-120.
- GREENWOOD, R. 1951: Younger intrusive rocks of Plateau province, Nigeria, compared with the alkaline rocks of New England. *Bull. Geol. Soc. Am.* **62**, 1151 - 1178.
- GUNOW, A.J., LUDINGTON, S. & MUNOZ, J.L. 1980: Fluorine in micas from the Henderson molybdenite deposit, Colorado. *Econ. Geol.* **75**, 1127-1137.
- HAASE, C.J., CHADAM, J., FEINN, D. & ORTOLEVA, P. 1980: Oscillatory zoning in plagioclase feldspar. *Science* **209**, 272-274.
- HAGGERTY, S.E. 1991: Oxide minerals - a mini-atlas. In: **Oxide minerals: petrologic and magnetic significance** (D.H. Lindsley, ed.). *Rev. Mineral.* **25**, 129-219.
- _____ & MARIANO, A. N. 1983: Strontian loparite and strontio-chevkinite: two new minerals in reomorphic fenites from the Paran  basin carbonatites, south America. *Mineral. Petrol.*
- HALDEN, N.M., HAWTHORNE, F.C., CAMPBELL, J.L., TEESDALE, W.J., MAXWELL, J. A. & HIGUCHI, D. 1993: Chemical characterization of oscillatory zoning and overgrowth in zircon using 3 MeV μ -PIXE. *Can. Mineral.* **31**, 637-647.
- HANSON, G. 1978: The application of trace elements to the petrogenesis of igneous rocks of granitoid composition. *Earth Planet. Sci. Lett.* **38**, 26-43.
- HARDING, R.R., MERRIMAN, R.J. & NANCARROW, P.H.A. 1982: A note on the occurrence of chevkinite, allanite, and zirkelite on St. Kilda, Scotland. *Mineral. Mag.* **46**, 445-448.
- HARTE, B. 1987: Metasomatic events recorded in mantle xenoliths: an overview. In: **Mantle Xenoliths** (P. H. Nixon, ed.). John Wiley & Sons Ltd., New York (625-640).
- HARRISON, T.M. & WATSON, E.B. 1984: The behaviour of apatite during crustal anatexis:

- equilibrium and kinetic considerations. *Geochim. Cosmochim. Acta* **48**, 1467-1477.
- HASKIN, L.A., & HASKIN, M.A. 1968: Relative and absolute abundance of the rare-earths. In: **Origin and distribution of elements** (L.H. Ahrens, ed.). Pergamon, New York (889-912).
- HASLAM, H.W. 1968: The crystallization of intermediate and acid magmas at Ben Nevis, Scotland. *J. Petrol.* **9**, 84-104.
- HAWTHORNE, F.C. 1981: Crystal chemistry of the amphiboles. *Rev. Mineral.* **9A**, 1-102.
- _____ 1983: The crystal chemistry of the amphiboles. *Can. Mineral.* **21**, 173-480.
- _____ & ČERNÝ, P. 1982: The mica group. In: **Granitic pegmatites in science and industry** (P. Černý, ed.). *Mineral. Assoc. Can., Short Course Handbook* **8**, 63-98.
- _____, GROAT, L.E., RAUDSEPP, M., BALL, N.A., KIMATA, M., SPIKE, F.D., GABA, R., HALDEN, N.M., LUMPKIN, G.R., EWING, R.C., GREGOR, R.B., LYTLE, F.W., ERCIT, T.S., ROSSMAN, G.R., WICKS, R.J., RAMIK, R.A., SHERRIFF, B.L., FLEET, M.E., & McCAMMON, C. 1991: Alpha-decay damage in titanite. *Am. Mineral.* **76**, 370-396.
- HELZ, R.T. 1979: Alkali exchange between hornblende and melt: a temperature-sensitive reaction. *Am. Mineral.* **64**, 953-965.
- HEWITT, D.A. & WONES, D.R. 1984: Experimental phase relations of the micas. In: **Micas** (S.W. Bailey, ed.), *Rev. Mineral.* **13**, 201-256.
- HIBBARD, M.J. 1981: The magma mixing origin of mantled feldspars. *Am. Mineral.* **76**, 158-170.
- HILDRETH, W. 1981: Gradients in silicic magma chambers: implications for lithospheric magmatism. *J. Geophys. Res.* **86**, 10153-10192.
- HOLTZ, F., BARBEY, P., JOHANNES, W. & PICHAVANT, M. 1989: Composition and temperatures of minimum melt in the quartz-albite-orthoclase system for H₂O-

- undersaturated conditions: Experimental investigations. *Terra cognita* **1**, 271-272.
- _____, JOHANNES, W. & PICHAVANT, M. 1992: Effect of excess aluminum on phase relations in the system quartz-albite-orthoclase: experimental investigations at 2 kbar and reduced H₂O-activity. *Eur. J. Mineral.* **4**, 137-152.
- _____, PICHAVANT, M., BARBEY, P. & JOHANNES, W. 1992: Effects of H₂O on liquidus phase relations in the haplogranite system at 2 and 5 kbar. *Am. Mineral.* **77**, 1223-1241.
- HOOVER, J.D. & IRVINE, T.N. 1978: Liquidus relations and Mg-Fe partitioning in part of the system $\text{Mg}_2\text{SiO}_4\text{-Fe}_2\text{SiO}_4\text{-CaMgSi}_2\text{O}_6\text{-CaFeSi}_2\text{O}_6\text{-KAlSi}_3\text{O}_8\text{-SiO}_2$. *Carnegie Inst. Washington Year Book* **77**, 774-784.
- HOSSAIN, M.T. & TURAKI, U.M. 1983: Bibliography on the Younger Granite ring complexes. *J. African Earth Sci.* **1**, 73-81.
- HUEBNER, J.S. 1980: Phase equilibria at low pressure. In: **Pyroxenes** (C.T. Prewitt, ed.). *Rev. Mineral.* **7**, 213-288.
- HUPPERT, H.E. & SPARKS, R.S.J. 1988: The generation of granitic magmas by intrusion of basalt into continental crust. *J. Petrol.* **29**, 599-624.
- HUPPERT, H.E., SPARKS, R.S.J. & TURNER, J.S. 1984: Some effects of viscosity on the dynamics of replenished magma chamber. *J. Geophys. Res.* **89**, 6857-6877.
- IKE, E.C. 1979: *The structure, petrology, and geochemistry of the Tibchi Younger Granite ring complex, Nigeria*. Unpublished Ph.D. thesis, University of St. Andrews, St. Andrews, U.K.
- _____, 1983: The structural evolution of the Tibchi ring-complex: a case study for the Nigerian Younger granite province. *J. Geol. Soc. Lond.* **140**, 781-788.
- _____, BOWDEN, P. & MARTIN, R.F. 1984: Fayalite and clinopyroxene in the porphyries of the Tibchi ring complex, Nigeria: postmagmatic initiation of a peralkaline trend. *Can. Mineral.* **22**, 401-409.

- _____, _____ & _____ 1985: Amphibole in the porphyries of the Tibchi anorogenic ring-complex, Nigeria: product of deuteric adjustments. *Can. Mineral.* **23**, 447-456.
- _____ & SAKOMA, E.M. 1985: Field geology of the Dago ring complex. *J. Afr. Earth Sci.* **3**, 97-100.
- IMEOKPARIA, E.G. 1982: Tin content of biotite from Afu Younger Granite complex, central Nigeria. *Econ. Geol.* **77**, 1710-1724
- _____ 1984: Geochemistry of granitic rocks from the Kwandonkaya complex, northern Nigeria. *Lithos* **17**, 103-115.
- ITO, J. 1967: A study of chevkinite and perrierite. *Am. Mineral.* **52**, 1094-1104.
- _____ & AREM, J.E. 1971: Chevkinite and perrierite: synthesis, crystal growth and polymorphism. *Am. Mineral.* **56**, 307-319.
- IXER, R.A., ASHWORTH, J.R. & POINTER, C.M. 1987: Accessory mineralogy of the Ririwai biotite granite, Nigeria, and its albitized and greisenized facies. *Geol. J.* **22**, 403-427.
- IZZETT, G.A. & WILCOX, R.E. 1968: Perrierite, chevkinite, and allanite in upper Cenozoic ash beds in the western United States. *Am. Mineral.* **53**, 1558-1567.
- JACKSON, K.J. & HELGESON, H.C. 1985: Chemical and thermodynamic constraints in the hydrothermal transport and deposition of tin: II. Interpretation of phase relations in the southeast Asian Tin belt. *Econ. Geol.* **80**, 1365-1378.
- JACOBSON, R.R. & JACQUES, R.H. 1944: Reports on wolframite investigation. *Rep. Geol. Surv. Nigeria.*, 5-13.
- _____, MACLEOD, W.N. & BLACK, R. 1958: Ring complexes in the Younger Granite of northern Nigeria. *Geol. Soc. London Mem.* **1**.
- _____, SNELLING, N.J. & TRUSWELL, J.F. 1963: Age determinations in the geology of Nigeria with special reference to the Older and Younger Granites of Nigeria. *Overseas*

Geol. Mineral. Res. **9**, 168-182.

JAFFE, H.W., EVANS, H.T. & CHAPMAN, R.W. 1958: Occurrence and age of chevkinite from the Devil's Slide fayalite quartz syenite near Stark, New Hampshire. *Am. Mineral.* **41**, 474-484.

JANECZNEK, J. 1989: Manganoan fayalite and products of its alteration from the Strzegom pegmatites, Poland. *Mineral. Mag.* **53**, 315-325.

KAYODE, A.A. 1976: On the genesis of small and large feldspar porphyritic Older Granites in the Igbo Ora complex, southwestern Nigeria. In: *Geology of Nigeria* (C.A. Kogbe, ed.). Elizabethan Publ. Co., Lagos, Nigeria (75-84).

KEPPLER, H. 1993: Influence of fluorine on the enrichment of high field strength trace element in granitic rocks. *Contrib. Mineral. Petrol.* **114**, 479-488.

KINNAIRD, J.A. 1976: *A geological investigation of some Nigerian Jurassic granites and their mineralization*. Unpubl. M.Sc. Thesis, University of St. Andrews, Scotland, U. K.

_____ 1981: *Geology of the Nigerian anorogenic ring complexes*. Map scale: 1:500,000. John Bartholomew, Edinburgh.

_____ 1985: Hydrothermal alteration and mineralization of the alkaline anorogenic complexes of Nigeria. *J. Afr. Earth Sci.* **3**, 229-251.

_____ 1987: *Hydrothermal alteration and mineralization of the Nigerian anorogenic ring complexes: with special reference to the Saiya-Shokobo complex*. Unpubl. Ph.D. thesis, University of St. Andrews, St. Andrews, U. K.

_____ & BOWDEN, P. 1987: African anorogenic alkaline magmatism and mineralization - a discussion with reference to the Niger-Nigerian Province. *Geol. J.* **22**, 297-340.

_____, ABERNETHY, C.A. & BOWDEN, P. 1992: The significance of mica compositional variations in Nigerian A-type granites: is A for annite? *EOS, Trans. Amer. Geophys.*

Union 73(14), 348.

- _____, BATCHELOR, R.A., WHITLEY, J.E. & MacKENZIE, A.B. 1985a: Geochemistry, mineralization, and hydrothermal alteration of the Nigerian high heat producing granites. In: **High heat producing (HHP) granites, hydrothermal circulation and ore genesis** (C. Halls, ed.). Inst. Min. Metall., London, U.K. (169-195).
- _____, BOWDEN, P., IXER, R.A. & ODLING, N.W.A. 1985b: Mineralogy, geochemistry, and mineralization of the Ririwai complex, northern Nigeria. *J. Afr. Earth Sci.* 3, 185-222.
- KIRKPATRICK, R.J. 1975: Crystal growth from a melt: a review. *Am. Mineral.* 60, 798-814.
- KLEEMAN, G.J. & TWIST, D. 1989. The compositionally zoned sheet-like granite pluton of the Bushveld complex: evidence bearing on the nature of A-type magmatism. *J. Petrol.* 30, 1383-1414.
- KOGBE, C.A. & OBIALO, A.U. 1976: Statistics of mineral production in Nigeria (1946 to 1967) and the contribution of the mineral industry to the Nigerian economy. In: **Geology of Nigeria** (C.A. Kogbe, ed.). Elizabethan Publishing Co., Lagos, Nigeria (391-426).
- KONTAK, D.J. 1991: The East Kemptville topaz-muscovite leucogranite, Nova Scotia. II. Mineral chemistry. *Can. Mineral.* 29, 37-60.
- KOYAGUCHI, T. 1991: Enclaves in volcanic rocks from Japan. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.), Elsevier, Amsterdam (235-250).
- _____ & BLAKE, S. 1991: Origin of mafic enclaves: constraints on the magma mixing model from fluid dynamic experiments. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (415-429).
- KRETZ, R. 1983: Symbols for rock-forming minerals. *Am. Mineral.* 68, 277-279.
- KROLL, H. & RIBBE, P.H. 1983: Lattice parameters, composition, and Al,Si order in alkali feldspars. *Rev. Mineral.* 2, 57-99.

- KRÖNER, A. 1981: Precambrian plate tectonics. In: **Precambrian Plate Tectonics** (A. Kröner, ed). Elsevier, New York.
- LAHTI, S.I. 1987: Zoning in columbite-tantalite crystals from the granite pegmatites of the Eräjärvi area, southern Finland. *Geochim. Cosmochim. Acta* **51**, 509-517.
- LALONDE, A.E. & BERNARD, P. 1993: Composition and color of biotite from granites: two useful properties in the characterization of plutonic suites from the Hepburn internal zone of Wopmay orogen, Northwest Territories. *Can. Mineral.* **31**, 203-217.
- _____ & MARTIN, R.F. 1983a: The Baie-des-Moutons syenitic complex, La Tabatière, Québec. 1. Petrography and feldspar mineralogy. *Can. Mineral.* **21**, 65-79.
- _____ & _____ 1983b: The Baie-des-Moutons syenitic complex, La Tabatière, Quebec. II. The ferromagnesian minerals. *Can. Mineral.* **21**, 81-91.
- LANDOLL, J.D. & FOLAND, K.A. 1996: The formation of quartz-syenite by crustal contamination at Mt. Shefford and other Monteregian complexes, Quebec. *Can. Mineral.* **34**, 301-304.
- LANGMUIR, C.H., VOCKE, R.D., HANSON, G.N., HART, S.R. 1977: A general mixing equation with applications to Icelandic basalts. *Earth and Planet. Sci. Lett.* **37**, 380-392.
- LEAKE, B.E. 1978: Nomenclature of amphiboles. *Can. Mineral.* **16**, 501-520.
- LE BAS, M.J. 1977: **Carbonatite-nephelinite volcanism**. John Wiley & Sons, New York, N.Y.
- LE BAS, M.J., LE MAITRE, R.W., STRECKEISEN, A. & ZANETTI, B. 1986: A chemical classification of volcanic rocks based on the total alkali silica diagram. *J. Petrol.* **27**, 745-750.
- LIANKUN, S. & KUIRONG, Y. 1991: A two-stage crust-mantle interaction model for mafic microgranular enclaves in the Daning granodiorite pluton, Guangxi, China. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (95-112).

- LINDSLEY, D.H. 1980: Phase equilibria of pyroxenes at pressures > 1 atmosphere. In: **Pyroxenes** (C.T. Prewitt, ed.). *Rev. Mineral.* **7**, 289-307.
- LOISELLE, M. G. & WONES, D. R. 1979: Characteristics and origin of anorogenic granites. *Geol. Soc. Am, Abstr. Prog.* **11**, 468.
- LONG, P.E. 1978: Experimental determination of partition coefficients for Rb, Sr, and Ba between alkali feldspar and silicate liquids. *Geochim. Cosmochim. Acta* **42**, 833-846.
- LUTH, W.C., JAHNS, R.H. & TUTTLE, O.F. 1964: The granite system at pressures of 4 to 10 kilobars. *J. Geophys. Res.* **69**, 759-773.
- MACLEOD, W.N., TURNER, D.C. & WRIGHT, E. P. 1971: The geology of Jos Plateau. I. General geology. *Bull. Geol. Surv. Nigeria* **32**.
- MAHOOD, G. & HILDRETH, W. 1983: Large partition coefficients for trace elements in high-silica rhyolites. *Geochim. Cosmochim. Acta* **47**, 11-30.
- MANNING, D.A.C. 1981: The effect of fluorine on the liquidus phase relationships in the system Qz-Ab-Or with excess water at 1 kbar. *Contrib. Mineral. Petrol.* **76**, 206-215.
- _____ & PICHAVANT, M. 1988: Volatiles and their bearing on the behaviour of metals in granitic systems. In: **Recent Advances in the Geology of granite-related Mineral deposits** (R.P. Taylor & D.F. Strong, eds.). *Can. Inst. Min. Metall., Spec. Vol.* **39**, 13 - 24.
- MARTIN, R.F. 1984: Trends of albitization in the Montgenèvre ophiolite, Western Alps. *Bull. Minéral.* **107**, 345-356.
- _____ & LALONDE, A.E. 1979: Turbidity in K-feldspars: causes and implications. *Geol. Soc. Am. Abstr. Programs.* **11**, 472-3.
- _____ & BOWDEN, P. 1981: Peraluminous rocks produced by rock-fluid interaction in the Ririwai anorogenic ring-complex, Nigeria: mineralogical evidence. *Can. Mineral.* **19**, 65-

- _____ & MOROGAN, V. 1988: Partial melting of fenitized crustal xenoliths in the Oldoinyo Lengai carbonatitic volcano, Tanzania: Reply. *Am. Mineral.* **73**, 1468-1471.
- MASON, R.A. 1992: Models of order and iron-fluorine avoidance in biotite. *Can. Mineral.* **30**, 343-354.
- McCARTHY, T.S. & HASTY, R.A. 1976: Trace element distribution patterns and their relationship to the crystallization of granitic melts. *Geochim. Cosmochim. Acta* **40**, 1351-1358.
- McDOWELL, S.D. 1979: Chevkinite from the Little Chief granite porphyry stock, California. *Am. Mineral.* **64**, 721-727.
- MERILAINEN, K. 1976: The granulite complex and adjacent rocks in Lapland, northern Finland. *Geol. Surv. Finland Bull.* **281**.
- MICHAEL, P.J. 1988: Partition coefficients for rare earth elements in mafic minerals of high silica rhyolites: the importance of accessory mineral inclusions. *Geochim. Cosmochim. Acta* **52**, 275-282.
- _____ 1990: A review of the compositional variations of amphibole in alkaline plutonic complexes. *Lithos* **26**, 135-156.
- MITCHELL, R.H. & PLATT, R.G. 1978: Mafic mineralogy of ferroaugite syenite from Coldwell alkalic complex, Ontario, Canada. *J. Petrol.* **19**, 627-651.
- MONIER, G. & ROBERT, J.-L. 1986: Evolution of the miscibility gap between muscovite and biotite solid solutions with increasing Li content: an experimental study in the system $K_2O-Li_2O-MgO-FeO-Al_2O_3-SiO_2-H_2O-HF$ at 600°C, 2 kbar P_{H_2O} : comparison with natural lithium micas. *Mineral. Mag.* **50**, 641-651.
- MORA, C.I. & VALLEY, J.W. 1989: Halogen-rich scapolite and biotite: implications for

- metamorphic fluid-rock interaction. *Am. Mineral.* **74**, 721-737.
- MOREAU, C., BROWN, W.L. & KARCHE, J.-P. 1987: Monzo-anorthosite from the Taguëï ring-complex, Aïr, Niger: a hybrid rock with cumulus plagioclase and an infiltrated granitic intercumulus liquid? *Contrib. Mineral. Petrol.* **95**, 32-43.
- MORIMOTO, N. 1989: The nomenclature of pyroxenes. *Can. Mineral.* **27**, 143-156.
- MOROGAN, V. 1988: *Fenitization, hallmark of ijolite-carbonatite magmatic association*. Doctoral Dissertation, University of Stockholm, Stockholm, Sweden.
- _____ & MARTIN, R.F. 1985: Mineralogy and partial melting of fenitized crustal xenoliths in the Oldoinyo Lengai carbonatitic volcano, Tanzania. *Am. Mineral.* **70**, 1114-1126.
- MUELLER, R.F. 1972: Stability of biotite: a discussion. *Am. Mineral.* **57**, 300-316.
- MUMPTON, F.A. & ROY, R. 1981: Hydrothermal stability studies of the zircon-thorite group. *Geochim. Cosmochim. Acta* **21** 217-238.
- MUNGALL, J.E. 1993: *Compositional effects of magma mixing and diffusive mass transport on the basalt - pantellerite suite, Terceira, Azores*. Unpubl. Ph.D. thesis, McGill University, Montreal, Canada.
- MUNOZ, J.L. 1984: F-OH and Cl-OH exchange in micas with applications to hydrothermal ore deposits. In: **Micas** (S.W. Bailey, ed). *Rev. Mineral.* **13**, 469-493.
- _____ 1992: Calculation of HF and HCl fugacities from biotite compositions: revised equation. *Geol. Soc. Am., Abstr. Programs* **24**, A221.
- _____ & LUDINGTON, S.D. 1974: Fluoride-hydroxyl exchange in biotite. *Am. J. Sci.* **274**, 396-413.
- _____ & SWENSEN, A. 1981: Chlorine-hydroxyl exchange in biotite and estimation of relative HCl/HF activities in hydrothermal fluids. *Econ. Geol.* **76**, 2212-2221.
- NAGASAWA, H. & SCHNETZLER, C.C. 1971: Partitioning of rare earth, alkali and alkali earth

- elements between phenocrysts and acidic igneous magma. *Geochim. Cosmochim. Acta* **35**, 953-968.
- NAKADA, S. 1991: Magmatic processes in titanite-bearing dacites, central Andes of Chile and Bolivia. *Am. Mineral.* **76**, 548-560.
- NASH, W.P. & WILKINSON, J.F.G. 1970: Shonkin Sag lacolith, Montana. 1. Mafic minerals and estimation of temperature, pressure, oxygen fugacity, and silica activity. *Contrib. Mineral. Petrol.* **25**, 241-269
- _____ & CRECRAFT, H.R. 1985: Partition coefficients for trace elements in silicic magmas. *Geochim. Cosmochim. Acta* **49**, 2309-2322.
- NASSIF, G.J. 1993: *The Strange Lake peralkaline complex, Quebec-Labrador: the hypersolvus - subsolvus granite transition and the feldspar mineralogy*. Unpubl. M.Sc. thesis, McGill University, Montreal, Canada.
- NELSON, S.T. & MONTANA, A. 1992: Sieve-textured plagioclase in volcanic rocks produced by rapid decompression. *Am. Mineral.* **77**, 1242-1249.
- NIXON, G.T. 1988a: Petrology of the Younger andesites and dacites of Iztaccíhuatl volcano, Mexico: 1. Disequilibrium phenocryst assemblages as indicators of magma chamber processes. *J. Petrol.* **29**, 213-264.
- _____ 1988b: Petrology of the Younger andesites and dacites of Iztaccíhuatl volcano, Mexico: II. Chemical stratigraphy, magma mixing, and the composition of basaltic magma influx. *J. Petrol.* **29**, 265-303.
- _____ & PEARCE, T.H. 1987: Laser-interferometry study of oscillatory zoning in plagioclase: the record of magma mixing and phenocryst recycling in calc-alkaline magma chambers, Iztaccíhuatl volcano, Mexico. *Am. Mineral.* **72**, 1144-1162.
- O'BRIEN, H.E., IRVING, A.J. & McCALLUM, I.S. 1988: Complex zoning and resorption of

- phenocrysts in mixed potassic mafic magmas of the Highwood Mountains, Montana. *Am. Mineral.* **73**, 1007-1024.
- O'HALLORAN, D.A. 1985: Ras ed Dom migrating ring complex: A-type granites and syenites from the Bayuda Desert, Sudan. *J. Afr. Earth Sci.* **3**, 61-75.
- ORSINI, J.-B., COCIRTA, C. & ZORPI, M.-J. 1991: Genesis of mafic microgranular enclaves through differentiation of basic magmas, mingling, and chemical exchange with their host granitoid magmas. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (445-464).
- ORT, M.H. 1992: Orbicular volcanic rocks of Cerro Panizos: their origin and implications for orb formation. *Geol. Soc. Am. Bull.* **104**, 1048-1058.
- PARSONS, I. 1978: Feldspars and fluids in cooling plutons. *Mineral. Mag.* **42**, 1-18.
- _____ 1981: The Klokken gabbro-syenite complex, south Greenland: quantitative interpretation of mineral chemistry. *J. Petrol.* **22**, 233-260.
- _____, MASON, R.A., BECKER, S.M. & FINCH, A.A. 1991: Biotite equilibria and fluid circulation in the Klokken intrusion. *J. Petrol.* **32**, 1299-1333.
- PEARCE, J.A., HARRIS, N.B.W. & TINDLE, A.G. 1984: Trace element discrimination diagrams for tectonic interpretation of granitic rocks. *J. Petrol.* **25**, 956-983.
- PEREZ, J.P. 1985: *Nouvelles données sur le complexe granitique anorogénique de Taghouaji (République du Niger): influence des fluides au cours de la cristallisation*. Thèse de l'Université de Nancy, Nancy, France.
- PITCHER, W.S. 1991: Synplutonic dykes and mafic enclaves. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (383-391).
- PLATT, R.G. 1996: Nepheline syenite complexes - an overview. In: **Undersaturated alkaline rocks: mineralogy, petrogenesis, and economic potential** (R. H. Mitchell, ed.). *Min.*

- POINTER, C.M., ASHWORTH, J.R. & IXER, R.A. 1988a: The zircon-thorite mineral group in metasomatized granite, Ririwai, Nigeria. 1. Geochemistry and metastable solid solution of thorite and coffinite. *Mineral. Petrol.* **38**, 245-262.
- _____, _____, _____ 1988b: The zircon-thorite mineral group in metasomatized granite, Ririwai, Nigeria, 2. Zoning, alteration, and exsolution in zircon. *Mineral. Petrol.* **39**, 21-37.
- RAHAMAN, M.A., VAN BREEMEN, O., BOWDEN, P. & BENNETT, J.N. 1984: Age and migration of anorogenic ring complexes, northern Nigeria. *J. Geol.* **92**, 173-184.
- RAMBERG, H. 1952: Chemical bonds and the distribution of cations in silicates. *J. Geol.* **69**, 331-355.
- RÄMÖ, O.T. 1991: Petrogenesis of the Proterozoic rapakivi granites and related basic rocks of the southeastern Fennoscandia: Nd and Pb isotopic and general geochemical constraints. *Geol. Surv. Finland Bull.* **355**.
- RIBBE, P.H. 1983: Aluminum-silicon order in feldspars, domain textures and diffraction patterns. *Rev. Mineral.* **2**, 21-55.
- ROBINSON, P. 1980: The compositional space of terrestrial pyroxenes - internal and external limits. In: **Pyroxenes** (C.T. Prewitt, ed.). *Rev. Mineral.* **7**, 419-494.
- _____, SPEAR, F.S., SCHUMAKER, J.C., LAIRD, J., KLEIN, C., EVANS, B.W. & DOOLAN, B.L. 1982: Phase relations of metamorphic amphiboles: Natural occurrence and theory. In: **Amphiboles: Petrology and experimental phase relations**. *Rev. Mineral.* **9B**, 1-227.
- ROCK, N.M.S. 1987: The need for standardization of normalized multi-element diagrams. *Geochem. J.* **21**, 75-84.

- ROSENBERG, P.E. & FOIT, F.F., Jr. 1977: Fe²⁺-F avoidance in silicates. *Geochim. Cosmochim. Acta* **41**, 345-436.
- ROWINS, S.M., LALONDE, A.E. & CAMERON, E.M. 1991: Magmatic oxidation in the syenitic Murdork Creek intrusion, Kirkland Lake area, Ontario: evidence from the ferromagnesian silicates. *J. Geol.* **99**, 395-414.
- RUBIN, J.N., HENRY, C.D. & PRICE, J.G. 1989: Hydrothermal zircons and zircon overgrowths, Sierra Blanca Peaks, Texas. *Am. Mineral.* **74**, 865-869.
- RUDNICK, R.L. & PRESPEL, T. 1990: Geochemistry of intermediate- to high-pressure granulites. In: **Granulites and crustal evolution**. (D. Vielzeuf & Ph. Vidal, eds.). NATO ASI Series C: Mathematical Physical Sciences 311 (523-550).
- SAKOMA, E.M. 1986: *A comparative study of the geology and mineralization in Diko, Niger state and Kalato, Bauchi state*. Unpubl. M.Sc. thesis, Ahmadu Bello University, Zaria, Nigeria.
- _____ & MARTIN, R.F. 1991: The Kwandonkaya anorogenic ring complex, Younger Granite Province of Nigeria. *Geol. Assoc. Can./Mineral. Assoc. Can. Prog. with Abstr.* **16**, A109.
- _____ & _____ 1992: Metaluminous-peraluminous transition in the Kwandonkaya complex, Mesozoic A-type granites of Nigeria. *Eos, Trans. Am. Geophys. Union* **73(14)**, 354.
- SATO, H. 1975: Diffusion coronas around quartz xenocrysts in andesite and basalt from Tertiary volcanic region in Northeastern Shikoku, Japan. *Contrib. Mineral. Petrol.* **50**, 48-64.
- SAWKA, W.N., CHAPPELL, B.W. & NORRISH, K. 1984. Light-rare-earth element zoning in sphene and allanite during granitoid fractionation. *Geology* **12**, 131-134.
- SCHANDL, E.S., MARTIN, R.F. & STEVENSON, J.S. 1986: Feldspar mineralogy of the Sudbury igneous complex and the Onaping Formation, Sudbury, Ontario. *Can. Mineral.* **24**, 747-759.

- SCHNEIDERMAN, J.S. 1991: Petrology and mineral chemistry of the Ascutney Mountain igneous complex. *Am. Mineral.* **76**, 218-229.
- SCHNETZLER, C.C. & PHILPOTTS, T.A. 1970: Partition coefficients of rare earth elements between igneous matrix and rock-forming mineral phenocrysts. II. *Geochim. Cosmochim. Acta* **34**, 331-340.
- SHAW, D.M. 1970: Trace element fractionation during anatexis. *Geochim. Cosmochim. Acta* **34**, 237-243.
- SKJERLIE, K.P. & JOHNSTON, A.D. 1992: Vapour-absent melting at 10 kbar of a biotite- and amphibole-bearing tonalitic gneiss: implications for the generation of A-type granites. *Geol.* **20**, 263-266.
- _____ & _____ 1994: Fluid-absent melting behaviour of a F-rich tonalitic gneiss at mid-crustal pressures: implications for the generation of anorogenic granites. *J. Petrol.* **34**, 785-851.
- SMITH, J.V. 1974: **Feldspar minerals. 2. Chemical and textural properties.** Springer Verlag, New York.
- SMITH, P. & PARSONS, I. 1974: The alkali feldspar solvus at 1 kbar water-vapour pressure. *Mineral. Mag.* **39**, 747-767.
- SOO, S.H. 1967: **Fluid dynamics of multiphase systems.** Blaisdell, Waltham, Mass.
- SPARKS, R.S.J., HUPPERT, H.E. & TURNER, J.S. 1984: The fluid dynamics of evolution of magma chambers. *Phil. Trans. R. Soc. London A* **310**, 511-534.
- _____, SIGURDSSON, H. & WILSON, L. 1977: Magma mixing: a mechanism for triggering acid explosive eruptions. *Nature* **267**, 315-318.
- SPEER, J.A. 1984: Micas in igneous rocks. In: Micas (S. W. Bailey, ed). *Rev. Mineral.* **13**, 299-356.

- _____ 1987: Evolution of magmatic AFM mineral assemblages in granitic rocks: the hornblende + melt $\rightleftharpoons$ biotite reaction in the Liberty Hill pluton, South Carolina. *Am. Mineral.* **72**, 863-878.
- _____ & GIBBS, G.V. 1976: The crystal structure of synthetic titanite (CaTiOSiO_4), and the domain textures of natural titanites. *Am. Mineral.* **61**, 238-247.
- SPERA, F.J. 1980: Aspects of magma transport. In: **Physics of magmatic processes** (R.B. Hargraves. ed.), Princeton Univ. Press, Princeton, N.J. (265-323).
- STEPHENSON, D. & UPTON, B.G.J. 1982: Ferromagnesian silicates in a differentiated alkaline complex: Kûngnât Fjeld, south Greenland. *Mineral. Mag.* **46**, 283-300.
- STEWART, D.B. & WRIGHT, T.L. 1974: Al/Si order and symmetry of natural alkali feldspars and the relationship of strained cell parameters to bulk composition. *Bull. Soc. Franc., Mineral. Cristallogr.* **97**, 356-377.
- STONE, M., EXLEY, C.S. & GEORGE, M.C. 1988: Composition of trioctahedral micas in the Cornubian batholith. *Mineral. Mag.* **2**, 175-192.
- STRECKEISEN, A.L. 1973: Plutonic rocks - classification and nomenclature recommended by the I.U.G.S. subcommission on the systematics of igneous rocks. *Geotimes* **18**, 26-30.
- _____ 1976: To each plutonic rocks its proper name. *Earth Sci. Rev.* **12**, 1-33.
- STULL, R.J. 1979: Mantled feldspars and synneusis. *Am. Mineral.* **64**, 514-518.
- SWANSON, S.E. 1977: Relation of nucleation and crystal growth rate to the development of granitic textures. *Am. Mineral.* **62**, 966-978.
- TAYLOR, R.P. 1992: Petrological and geochemical characteristics of the Pleasant Ridge zinnwaldite-topaz granite, southern New Brunswick, and comparison with other topaz-bearing felsic rocks. *Can. Mineral.* **30**, 895-921.
- TAYLOR, S.R. & McLENNAN, S.M. 1985: **The continental crust: its composition and**

- evolution.** Blackwell Scientific Publication, Boston, Massachussetts.
- THOMAS, W.M. 1982: Stability relations of the amphibole hastingsite. *Am. J. Sci.* **282**, 136-164.
- THOMPSON, J.B. Jr. 1947: Role of aluminum in the rock-forming silicates. *Geol. Soc. Am. Bull.* **58**, 1232
- TINDLE, A.G. 1991: Trace element behaviour in microgranular enclaves from granitic rocks. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (313-331).
- _____ & PEARCE, J.A. 1981: Petrologic modelling of in situ fractional crystallization in the zoned Loch Doon pluton, Scotland. *Contrib. Mineral. Petrol.* **78**, 196-206.
- _____ & WEBB, P.C. 1990: Estimation of Li contents in trioctahedral micas using microprobe data; application to micas from granitic rocks. *Eur. J. Mineral.* **2**, 595-610.
- TORAMARU, A. 1991: Model of nucleation and growth of crystals in cooling body of magma. *Contrib. Mineral. Petrol.* **108**, 106-117.
- TSUCHIYAMA, A. 1985: Dissolution kinetics of plagioclase in the melt of the system diopside-albite-anorthite, and origin of dusty plagioclase in andesites. *Contrib. Mineral. Petrol.* **89**, 1-16.
- TURNER, D.C. 1963: Ring structures in the Sara-Fier complex, northern Nigeria. *Q. J. Geol. Soc. London* **119**, 34.
- _____ & BOWDEN, P. 1979: The Ningi-Burra complex, Nigeria: dissected calderas and migrating magmatic centers. *J. Geol. Soc. London* **136**, 105-119.
- TUTTLE, O.F. & BOWEN, N.L. 1958: Origin of granite in the light of experimental studies in the system $\text{NaAlSi}_3\text{O}_8$ - KAlSi_3O_8 - SiO_2 - H_2O . *Geol. Soc. Am., Mem.* **74**.
- VAN BREEMEN, O. & BOWDEN, P. 1973: Sequential age trends from some Nigerian Mesozoic granites. *Nature* **242**, 9-11.

- _____, HUTCHINSON, J. & BOWDEN, P. 1975: Age and origin of the Nigerian Mesozoic granites: a Rb-Sr isotopic study. *Contrib. Mineral. Petrol.* **50**, 157-172.
- VAN MIDDELAAR, W.T. & KEITH, J.D. 1990: Mica chemistry as an indicator of oxygen and halogen fugacities in the Cantung and other W-related granitoids in the North American Cordillera. In: **Ore-bearing granite system; petrogenesis and mineralizing processes** (H.J. Stein & J.L. Hannah, eds.). *Geol. Soc. Am., Spec. Pap.* **246**, 205-220.
- VERNON, R.H. 1991: Interpretation of microstructures of microgranitoid enclaves. In: **Enclaves and granite petrology** (J. Didier & B. Barbarin, eds.). Elsevier, Amsterdam (277-291).
- WAGMAN, D.D., EVANS, W.H., PARKER, V.B., SCHUMM, R.H., HALOW, I., BAILEY, S.M., CHURNEY, K.L. & NUTTALL, R.L. 1982: The NBS tables of chemical thermodynamic properties. Selected values for inorganic and C1 and C2 organic substances in SI units. *J. Phys. Chem. Ref. Data* **11**, Suppl. No. 2, 392p.
- WATSON, E.B. 1979: Zircon saturation in felsic liquids: experimental results and application to trace element geochemistry. *Contrib. Mineral. Petrol.* **70**, 407-419.
- _____ & HARRISON, T.M. 1984: Accessory mineral and the geochemical evolution of crustal magmatic systems: a summary of experimental approaches. *Phys. Earth Planet. Int.* **35**, 19-30.
- WEBB, P.K. 1974: Roof modification of a high level granite, Jos-Bukuru Younger Granite ring complex, Nigeria. *Geol. Mag.* **111**, 43-47.
- WEBSTER, J.D. & HOLLOWAY, J.R. 1990: Partitioning of F and Cl between magmatic hydrothermal fluids and highly evolved granitic magmas. In: **Ore-bearing granite system; petrogenesis and mineralizing processes** (H.J. Stein & J.L. Hannah, eds.). *Geol. Soc. Am. Spec. Pap.* **246**, 21-34.
- WHALEN, J.B. & CURRIE, K.L. 1984: The Topsails igneous terrane, Western Newfoundland:

- evidence for magma mixing. *Contrib. Mineral. Petrol.* **87**, 319-327.
- _____ & _____ 1990: The Topsails igneous suite, Western Newfoundland; fractionation and magma mixing in an "anorogenic" A-type granite suite. In: **Ore bearing granite systems; petrogenesis and mineralizing processes** (H.J. Stein & J.L. Hannah, eds.). *Geol. Soc. Am., Spec. Pap.* **246**, 287-299.
- _____, _____ & CHAPPELL, B.W. 1987: A-type granites: geochemical characteristics, discrimination, and petrogenesis. *Contrib. Mineral. Petrol.* **95**, 407-419.
- WHITE, A.J.R. & CHAPPELL, B.W. 1983: Granitoid types and their distribution in the Lachlan Fold Belt, Southeastern Australia. *Geol. Soc. Am. Mem.* **159**, 21-34.
- WICKHAM, S.M. 1987: The segregation and emplacement of granitic magmas. *J. Geol. Soc. London* **144**, 281-297.
- WILCOX, R.E. 1984: Optical properties of micas under the polarizing microscope. In: **Micas**. (S.W. Bailey, ed.). *Rev. Mineral.* **13**, 183-200.
- WILLIAMS, H., TURNER, F.J. & GILBERT, C.M. 1982: **An introduction to the study of rocks in thin sections**. W. H. Freeman & Co. New York (42-53).
- WONES, D.R. 1972: Stability of biotite: reply. *Am. Mineral.* **57**, 316-317.
- _____ 1989: The significance of the assemblage titanite-magnetite-quartz in granitic rocks. *Am. Mineral.* **74**, 744-749.
- _____ & EUGSTER, H.P. 1965: Stability of biotite: experiment, theory, and applications. *Am. Mineral.* **50**, 1228-1278.
- _____ & GILBERT, M.C. 1982: Amphiboles in the igneous environment. In: **Amphiboles: petrology and experimental phase relations**. *Rev. Mineral.* **9B**, 355-390.
- WOOLEY, A. R. 1991: The Chilwa alkaline province of Malawi. In: **Magmatism in extensional settings: the Phanerozoic African plate** (A.B. Kampunzu & R.T. Lubala, eds.). Springer-

Verlag, Berlin, 377-409.

WORDEN, R.H., WALKER, F.D.L., PARSONS, I. & BROWN, W.L. 1990: Development of microporosity, diffusion channels and deuteric coarsening in perthitic alkali feldspars.

Contrib. Mineral. Petrol. **104**, 507-515.

WRIGHT, J.B. 1975: Anorthosite - first occurrence in Nigeria and relevance to Younger Granite genesis. *Mineral. Mag.* **40**, 193-196.

WYLLIE, P.J. 1977: Crustal anatexis: an experimental review. *Tectonophys.* **43**, 41-71.

YUND, R.A. & TULLIS, J. 1983: Subsolidus phase relations in the alkali feldspars with emphasis on coherent phases. *Rev. Mineral.* **2**, 141-176.

ZEN, E-an 1986: Aluminium enrichment in silicate melts by fractional crystallization: some mineralogical and petrographic constraints. *J. Petrol.* **27**, 1095-1117.

_____ 1988: Phase relations of granitic rocks and their petrogenetic implications. *Ann. Rev. Earth Planet. Sci.* **16**, 21-51.

LIST OF APPENDICES

#	Description	
II.1:	Minimum compositions and temperatures at various Ps and volatile contents in granites	321
II.2:	A ternary plot of normative quartz-albite-orthoclase composition with data plotted from Appendix II.1	322
III.1:	Results of EMPA (spot) analyses of fayalite in subvolcanic rocks	323
III.2:	Results of EMPA (spot) analyses of clinopyroxene in subvolcanic rocks	324
III.3:	Results of EMPA (spot) analyses of amphibole in subvolcanic rocks	325
IV.1a:	SEM - BSE image of annitic biotite (Bt) in hornblende biotite granite porphyry . .	326
IV.1b:	SEM - BSE image of zoned zinnwaldite-trilithionite in drusy biotite granite	326
IV.2:	Results of EMPA of mica in subvolcanic rocks	327
IV.3:	Results of EMPA (spot) analyses of mica in plutonic rocks	328
IV.4:	Results of EMPA (spot) analyses of mica in altered granites	329
IV.5:	a) A schematic illustration of the structure of mica (modified from Mason 1991), and illustrating a cluster of 3 octahedral M-sites (M_3). b) Plot of mole fraction of F (X_F) vs. $X_{Mg(\frac{1}{2}Ti, Li)}$ for mica from Kwadonkaya complex. The area shaded m is the compositional field in which X_F is plotted against X_{Mg} , and the area shaded grey is the same field when the mole fraction of Ti and Li are added to that of Mg.	330
V.1:	Unit-cell parameters determined for alkali feldspars in representative samples	331
V.2:	Schematic illustration of structure, site occupancy, and monoclinic - triclinic inversion of alkali feldspar	332
V.3:	T-X phase diagram for alkali feldspars	333
VI.1:	SEM - EDS spectra of: a) chevkinite, b) allanite, c) fluocerite, and d) bastnäsité . .	334
VI.2:	SEM - EDS spectra of: a) fergusonite, b) xenotime, c) columbite, and d) unidentified REE-Th mineral	335
VI.3:	Photomicrographs of accessory minerals (chevkinite: a-c, allanite: d-e, and fluocerite: f)	336

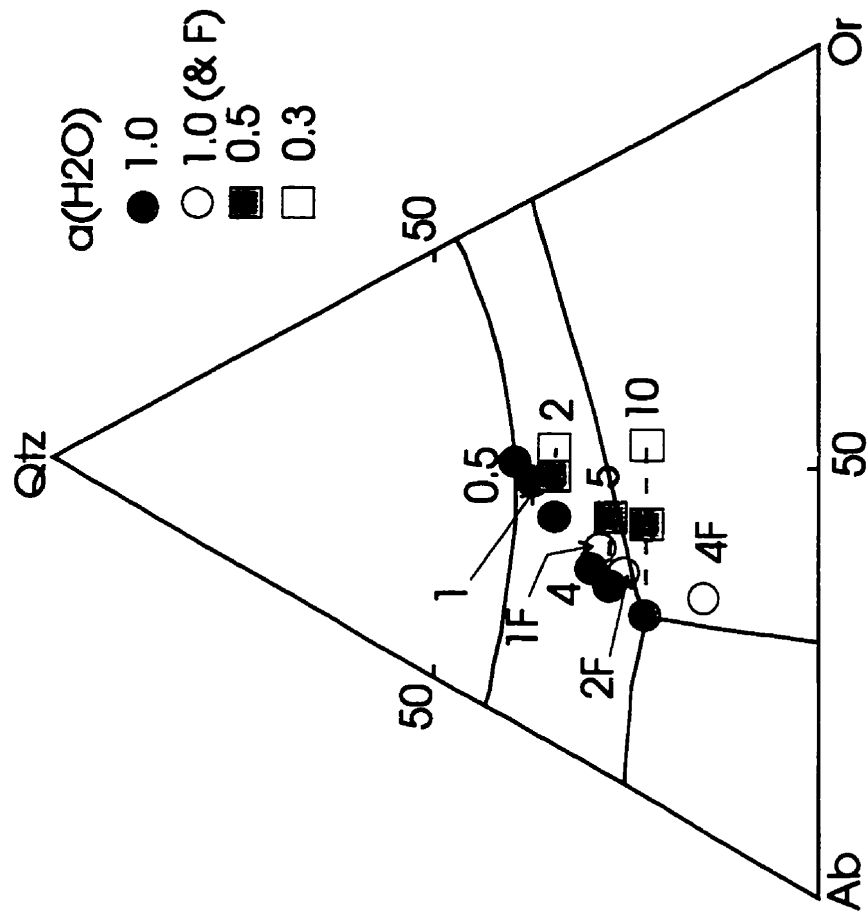
VI.4: Photograph of samples of beryl (Brl), quartz (Qtz), and topaz (Toz)	337
VI.5: SEM - BSE image of columbite (Col) - cassiterite (Cst) intergrowth	338
VII.1: Thin section photograph (a, b, e, and f), photomicrograph (c), and SEM-BSE image (f) of enclaves	339
VII.2: Results of EMPA (spot) analyses of plagioclase in enclaves within plutonic rocks	340
VII.3: Results of EMPA (spot) analyses of pyroxenes in enclaves and their host rocks . .	341
VII.4: Results of EMPA (spot) analyses of amphibole in enclaves (+host rock)	342
VII.5: Results of EMPA (spot) analyses of mica in enclaves and their host rocks	343
VII.6: Calculated proportions of mixing in enclaves assuming two end-member mixing . .	344
VIII.1: Melt compositions generated by partial melting model: a) melt compositions generated by (i) modal, and (ii) nonmodal melting of non-enriched trial source- rocks (Data plotted on Figs. 8.5a - d, and 8.7a - d). b) Enrichments of trial source-rock compositions for 20 and 60% (i) modal, and (ii) nonmodal melting to yield target melt composition (Data plotted on Figs. 8.6a - d, and 8.8a - d) . . .	345
VIII.2: a) p_i values compiled by Rämö (1991), b) calculated D and adjusted P values (i.e., bulk melt distribution coefficients) for nonmodal melting, and c) inferred "appropriate" p_i values that will yield target melt composition from selected source rock compositions	346

Appendix II.1: Minimum compositions and temperatures at various P_s and volatile contents in granites. a) Water saturated conditions at different pressures, i.e., $a(H_2O) = 1.00$, b) water-saturated conditions, $P_{Total} = 1.0$ kbar, and different fluorine contents, and c) water-undersaturated conditions

P_{Total} (kbar)	Quartz	Albite	Orthoclase	T, °C	Source
0.5	40	30	30	725	Tuttle & Bowen 1958
1.0	38	33	29		''
2.0	35	39	26		''
4.0	30	47	23		''
5.0	28	50	22		Luth et al. 1964
10.0	23	56	21		''
4 (dry)	35	30	35		''
b) F content					
1 wt. %	29	45	26	690	Manning 1981
2 wt. %	25	50	25	670	''
4 wt. %	15	58	27	680	''
c) i: $a(H_2O) = 0.5$					
P_{Total} , kbar					
2.0	35	34	31		Holtz et al. 1989, 1991
5.0	28	42	30		Ebadi and Johannes 1991
10.0	23	45	32		''
c). ii: $a(H_2O) = 0.3$					
2.0	35	31	34		Holtz et al. 1989, 1991
10.0	23	36	41		Ebadi and Johannes 1991

Appendix II.2: A ternary plot of normative quartz-albite-orthoclase composition with data plotted from Appendix II.1 (this is the same as inset in Fig. 2d).

Appendix II.2



Appendix III.1: Results of EMPA (spot) analyses of fayalite in subvolcanic rocks. *: indicates Fe_2O_3 calculated after the equation derived by Droop (1987); F/FM (in this and other Tables) = $(\text{FeO}_{\text{tot}})/(\text{FeO}_{\text{tot}} + \text{MgO})$; Fo = $\text{MgO}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{MnO})$, Fa = $\text{FeO}_{\text{tot}}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{MnO})$, Te = $\text{MnO}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{MnO})$; Rhyolite porphyry (49B), and fayalite syenogranite (57).

	49B.1	49B.2	49B.3	49B.4	57.1	57.2	57.3	57.4	57.5	57.6	57.7
SiO ₂	30.91	30.45	30.44	30.76	28.81	28.77	29.03	29.15	29.14	29.01	28.97
TiO ₂	0.04	0.03	0.01	0.02	0.02	0.03	-	-	0.04	0.00	0.04
Al ₂ O ₃	0.01	-	0.01	-	-	0.02	-	0.02	0.01	0.01	0.01
Cr ₂ O ₃	-	0.03	-	-	-	-	-	-	-	-	-
Fe ₂ O ₃ *	-	-	-	-	1.26	0.75	0.40	0.27	1.09	0.13	0.50
FeO	66.45	67.07	64.97	65.79	67.64	67.48	67.98	67.98	68.55	67.87	67.93
MnO	1.82	1.90	1.65	1.81	1.41	1.28	1.16	1.26	1.16	1.25	1.25
NiO	-	-	-	0.01	-	-	-	-	-	-	-
MgO	1.66	1.11	1.15	1.14	0.11	0.10	0.13	0.15	0.13	0.11	0.12
CaO	0.35	0.37	0.31	0.29	0.10	0.09	0.07	0.06	0.09	0.06	0.06
Na ₂ O	0.04	-	0.01	-	0.02	0.02	0.02	0.04	0.03	0.01	0.01
K ₂ O	-	0.00	0.01	-	-	0.01	-	0.01	-	-	0.01
Total	101.27	100.94	98.57	99.83	99.37	98.54	98.80	98.92	100.26	98.46	98.89
Structural formula on the basis of 4 cations											
Si ⁴⁺	1.02	1.01	1.03	1.03	0.98	0.99	1.00	1.00	0.99	1.00	0.99
Ti ⁴⁺	0.00	0.00	0.00	0.00	0.00	0.00	-	-	0.00	0.00	0.00
Al ³⁺	0.00	-	0.00	-	-	0.00	-	0.00	0.00	0.00	0.00
Cr ³⁺	-	0.00	-	-	-	-	-	-	-	-	-
Fe ³⁺	-	-	-	-	0.03	0.02	0.01	0.01	0.03	0.00	0.01
Fe ²⁺	1.83	1.86	1.85	1.85	1.93	1.94	1.95	1.95	1.94	1.95	1.95
Mn ²⁺	0.05	0.05	0.05	0.05	0.04	0.04	0.03	0.04	0.03	0.04	0.04
Ni ²⁺	-	-	-	0.00	-	-	-	-	-	-	-
Mg ²⁺	0.08	0.05	0.06	0.06	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Ca ²⁺	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na ⁺	0.00	-	0.00	-	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K ⁺	-	0.00	0.00	-	-	0.00	-	0.00	-	-	0.00
F/FM	0.96	0.97	0.97	0.97	1.00	1.00	1.00	1.00	1.00	1.00	1.00
Fo	4.14	2.78	2.99	2.92	0.27	0.25	0.34	0.38	0.34	0.29	0.31
Fa	93.28	94.51	94.58	94.45	97.70	97.88	97.98	97.79	98.00	97.89	97.89
Te	2.58	2.71	2.44	2.63	2.03	1.87	1.68	1.83	1.66	1.83	1.81

Appendix III.2: Results of EMPA (spot) analyses of clinopyroxene in subvolcanic rocks. Fs = $\text{FeO}_{\text{tot}}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{CaO})$, Wo = $\text{CaO}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{CaO})$, En = $\text{MgO}/(\text{FeO}_{\text{tot}} + \text{MgO} + \text{CaO})$; * indicates Fe_2O_3 calculated after Droop (1987); Rhyolite porphyry (49B, & 87), hedenbergite fayalite granite porphyry (43, 67, & 64).

	87.1c	87.2r	87.3c	87.41	87.5r	87.6	87.7	49B.1	49B.2	43.1	43.2	43.3	43.4	43.5	43.6	43.7	43.8	43.9	43.10	43.11	43.12
SiO ₂	47.88	47.62	47.34	47.69	47.55	47.15	46.78	49.40	48.74	49.83	49.42	49.29	49.10	49.50	48.99	49.43	48.37	48.80	49.12	48.70	48.99
TiO ₂	0.18	0.17	0.47	0.38	0.26	0.27	0.64	0.12	0.14	0.13	0.15	0.12	0.33	0.43	0.08	0.05	0.11	0.14	0.21	0.08	0.15
Al ₂ O ₃	0.23	0.29	0.64	0.39	0.41	0.41	0.73	0.26	0.51	0.17	0.16	0.18	0.57	0.42	0.22	0.11	0.21	0.18	0.35	0.23	0.16
Cr ₂ O ₃	-	-	-	-	-	-	-	-	-	-	0.02	-	0.03	-	0.03	-	-	-	-	0.05	-
Fe ₂ O ₃	0.78	2.17	1.29	0.84	1.13	1.66	0.64	1.47	0.91	-	-	-	-	-	0.15	-	-	-	-	0.20	2.59
FeO	29.45	28.84	29.71	30.49	30.29	29.56	29.40	26.50	27.13	28.31	29.24	28.96	28.14	27.92	28.98	27.98	27.91	28.19	29.04	29.07	27.59
MnO	0.47	0.45	0.84	0.79	0.84	0.76	0.85	0.74	0.64	0.56	0.71	0.56	0.65	0.79	0.61	0.68	0.55	0.63	0.73	0.63	0.69
NiO	-	-	-	-	-	-	-	0.05	0.04	0.04	-	-	0.04	-	-	0.09	0.05	0.08	-	0.03	-
MgO	0.03	0.00	0.45	0.52	0.50	0.44	0.09	2.26	2.06	0.79	0.74	0.80	1.52	2.40	0.77	0.69	0.78	0.77	0.78	0.73	0.75
CaO	19.40	19.49	18.77	18.40	18.36	18.65	19.18	19.55	18.61	20.14	19.33	19.89	19.04	18.43	19.52	20.03	19.87	19.08	18.92	19.18	18.86
Na ₂ O	0.56	0.61	0.34	0.34	0.32	0.34	0.33	0.60	0.65	0.39	0.63	0.49	0.45	0.37	0.58	0.44	0.46	0.78	0.41	0.57	1.07
K ₂ O	-	0.00	0.01	-	0.02	0.01	-	0.01	0.03	-	0.01	0.02	0.08	0.01	-	-	-	-	-	0.01	-
Tot	98.98	99.65	99.87	99.85	99.68	99.26	98.65	100.95	99.45	100.35	100.42	100.31	99.96	100.25	99.93	99.50	98.32	98.64	99.57	99.48	100.84
FeOt	30.15	30.79	30.87	31.25	31.30	31.06	29.97	27.82	27.95	28.31	29.24	28.96	28.14	27.92	29.11	27.98	27.91	28.19	29.04	29.25	29.92
Structural formula on the basis of 6 oxygens																					
IVSi	2.00	1.99	1.97	1.98	1.98	1.98	1.97	2.00	2.00	2.03	2.02	2.02	2.00	2.01	2.01	2.03	2.01	2.02	2.02	2.01	2.01
IVAl	-	0.01	0.03	0.02	0.02	0.02	0.03	0.00	-	-	-	-	-	-	-	-	-	-	-	-	-
T	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.03	2.02	2.02	2.00	2.01	2.01	2.03	2.01	2.02	2.02	2.01	2.01
VIAl	0.01	0.00	0.00	0.00	0.00	-	0.00	0.01	0.02	0.01	0.01	0.01	0.03	0.02	0.01	0.01	0.01	0.01	0.02	0.01	0.01
Ti	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00
Cr	-	-	-	-	-	-	-	-	-	-	0.00	-	0.00	-	0.00	-	-	-	-	0.00	-
Fe ³⁺	0.02	0.07	0.04	0.03	0.04	0.05	0.02	0.04	0.03	-	-	-	-	-	0.00	-	-	-	-	0.01	0.08
Fe ²⁺	1.03	1.00	1.03	1.06	1.05	1.03	1.03	0.89	0.93	0.96	1.00	0.99	0.96	0.95	1.00	0.96	0.97	0.98	1.00	1.00	0.94
Mn ²⁺	0.02	0.02	0.03	0.03	0.03	0.03	0.03	0.03	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.03	0.02	0.02
Ni	-	-	-	-	-	-	-	0.00	0.00	0.00	-	-	0.00	-	-	0.00	0.00	0.00	-	0.00	-
Mg	0.00	0.00	0.03	0.03	0.03	0.03	0.01	0.14	0.13	0.05	0.04	0.05	0.09	0.14	0.05	0.04	0.05	0.05	0.05	0.04	0.05
Ca	0.87	0.87	0.83	0.82	0.82	0.83	0.86	0.84	0.82	0.88	0.85	0.87	0.83	0.80	0.86	0.88	0.89	0.85	0.83	0.85	0.82
Na	0.05	0.05	0.03	0.03	0.03	0.03	0.03	0.05	0.05	0.03	0.05	0.04	0.04	0.03	0.05	0.03	0.04	0.06	0.03	0.05	0.08
K	-	0.00	0.00	-	0.00	0.00	-	0.00	0.00	-	0.00	0.00	0.00	0.00	-	-	-	-	-	0.00	-
M1,M2	2.00	2.01	2.01	2.00	2.01	2.01	2.00	2.01	2.00	1.95	1.98	1.98	1.99	1.98	1.99	1.95	1.98	1.97	1.96	1.99	2.01
F/FM	1.00	1.00	0.97	0.97	0.97	0.98	0.99	0.87	0.88	0.95	0.96	0.95	0.91	0.87	0.96	0.96	0.95	0.95	0.95	0.96	0.96
Fs	54.19	53.58	54.45	55.44	55.36	54.50	54.32	47.68	49.65	50.98	52.86	51.84	50.94	50.04	52.35	50.99	50.97	52.20	53.12	52.92	51.98
Wo	45.72	46.40	44.07	42.87	43.00	44.05	45.40	45.06	43.65	46.47	44.77	45.61	44.15	42.31	45.18	46.76	46.49	45.26	44.33	44.73	45.51
En	0.09	0.02	1.48	1.69	1.64	1.46	0.28	7.26	6.71	2.55	2.37	2.55	4.91	7.65	2.47	2.25	2.55	2.54	2.55	2.35	2.52

	43.13	43.14	43.15	43.16	43.17	43.18	43.19	43.20c	43.21r	67.1	67.2r	67.3c	67.4c	67.51	67.6r	67.7	64.1	64.2	64.3	64.4	64.5
SiO2	49.90	49.78	50.54	47.99	47.80	47.39	47.82	47.67	47.25	48.16	48.57	48.53	48.29	48.17	48.15	47.73	48.20	48.49	48.06	48.17	48.34
TiO2	0.09	0.10	-	0.07	0.13	0.26	0.18	0.14	0.11	0.08	0.03	0.06	0.14	0.16	0.14	0.08	0.09	0.02	0.10	0.10	0.05
Al2O3	0.19	0.38	0.03	0.14	0.18	0.23	0.19	0.18	0.23	0.08	0.14	0.13	0.21	0.26	0.17	0.17	0.53	0.60	0.59	0.57	0.36
Cr2O3	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.01	-	0.01	-	0.01
Fe2O3	-	-	-	1.70	2.56	3.95	2.00	2.79	3.14	-	1.39	0.93	1.98	1.78	0.80	1.68	1.00	1.08	1.47	0.94	0.31
FeO	29.13	29.03	26.55	27.59	27.19	25.74	27.02	27.45	26.69	27.64	26.90	26.57	27.13	26.99	27.22	27.39	27.85	27.70	27.28	27.30	27.58
MnO	0.72	0.60	0.60	0.63	0.69	0.68	0.68	0.72	0.63	0.99	0.81	0.90	0.64	0.66	0.68	0.71	0.85	0.78	0.79	0.90	0.82
NiO	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.05	0.00	0.00	0.01	0.08
MgO	0.57	0.37	0.69	0.76	0.68	0.72	0.74	0.55	0.50	0.81	1.05	1.00	1.36	1.31	1.32	0.52	0.05	0.06	0.05	0.06	0.05
CaO	19.09	18.59	19.89	20.35	19.56	18.81	19.82	19.92	20.33	20.50	20.74	20.71	19.43	19.79	19.91	20.34	21.08	21.54	21.29	21.18	21.58
Na2O	0.55	0.41	0.34	0.38	0.66	1.09	0.63	0.52	0.50	0.13	0.42	0.49	0.60	0.52	0.41	0.44	0.40	0.39	0.46	0.47	0.35
K2O	0.01	0.06	0.01	0.00	0.01	0.01	-	0.01	0.01	0.09	-	0.01	-	-	0.01	-	0.01	0.00	0.00	0.01	0.00
Tot	100.25	99.32	98.65	99.60	99.47	98.88	99.08	99.96	99.39	98.48	100.05	99.32	99.78	99.64	98.82	99.06	100.13	100.66	100.10	99.72	99.53
FeOt	29.14	29.03	26.55	29.11	29.50	29.29	28.82	29.96	29.51	27.64	28.15	27.40	28.91	28.59	27.95	28.90	28.76	28.67	28.60	28.15	27.86
Structural formula on the basis of 6 oxygens																					
IVSi	2.04	2.04	2.07	1.99	1.99	1.99	1.99	1.98	1.98	2.01	2.00	2.01	1.99	1.99	2.00	1.99	1.99	1.99	1.99	1.99	2.00
IVAl	-	-	-	0.01	0.01	0.01	0.01	0.01	0.01	-	0.00	-	0.01	0.01	0.00	0.01	0.01	0.01	0.01	0.01	-
T	2.04	2.04	2.07	2.00	2.00	2.00	2.00	1.99	1.99	2.01	2.00	2.01	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
VIA1	0.01	0.02	0.00	-	0.00	0.00	0.00	-	-	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.02	0.02	0.01	0.02	0.02
Ti	0.00	0.00	-	0.00	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	0.00	0.00	0.00	0.00
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.00	-	0.00	-	0.00
Fe3+	-	-	-	0.05	0.08	0.12	0.06	0.09	0.10	-	0.04	0.03	0.06	0.06	0.03	0.05	0.03	0.03	0.05	0.03	0.01
Fe2+	0.99	1.00	0.91	0.95	0.94	0.89	0.94	0.95	0.93	0.96	0.92	0.92	0.93	0.93	0.94	0.95	0.96	0.95	0.94	0.94	0.95
Mn2+	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.03	0.03	0.03	0.02	0.02	0.02	0.03	0.03	0.03	0.03	0.03	0.03
Ni	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.00	0.00	0.00	0.00	0.00
Mg	0.03	0.02	0.04	0.05	0.04	0.04	0.05	0.03	0.03	0.05	0.06	0.06	0.08	0.08	0.08	0.03	0.00	0.00	0.00	0.00	0.00
Ca	0.83	0.82	0.87	0.90	0.87	0.84	0.88	0.88	0.90	0.92	0.91	0.91	0.86	0.87	0.88	0.91	0.93	0.94	0.94	0.94	0.96
Na	0.04	0.03	0.03	0.03	0.05	0.09	0.05	0.04	0.04	0.01	0.03	0.04	0.05	0.04	0.03	0.04	0.03	0.03	0.04	0.04	0.03
K	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	0.00	0.00	0.00
M1,M2	1.94	1.92	1.87	2.01	2.01	2.02	2.01	2.02	2.03	1.99	2.01	2.00	2.01	2.01	2.00	2.01	2.01	2.01	2.01	2.00	2.00
F/FM	0.97	0.98	0.96	0.96	0.96	0.96	0.96	0.97	0.97	0.95	0.94	0.94	0.92	0.92	0.92	0.97	1.00	1.00	1.00	1.00	1.00
Fs	53.35	54.26	49.84	50.15	50.86	50.35	50.29	50.87	49.76	49.94	48.62	48.40	49.82	49.36	49.41	50.38	50.69	50.00	49.92	50.06	49.86
Wo	44.78	44.51	47.84	47.40	46.87	47.13	47.26	47.30	48.58	47.45	48.02	48.34	45.73	46.37	46.31	47.93	49.14	49.81	49.92	49.75	49.99
En	1.87	1.24	2.33	2.45	2.27	2.51	2.45	1.83	1.65	2.61	3.37	3.26	4.45	4.27	4.28	1.69	0.17	0.18	0.16	0.19	0.15

Appendix III.3: Results of EMPA (spot) analyses of amphibole in subvolcanic rocks. Rhyolite porphyry (49A), hedenbergite fayalite granite porphyry (41 & 67), hedenbergite hastingsite granite porphyry (64), hastingsite granite porphyry (40A), and hornblende biotite granite porphyry (51, 55, 56, 68A, 68B, & 79).

	49A.1	49A.2	49A.3	41.1	41.2	41.3	43.1	67.1	67.2	64.1	64.2	64.3	64.4	64.5	64.6	64.7	40A.1	40A.2	40A.3	40A.4
SiO2	49.32	49.85	48.72	45.50	46.76	45.63	50.76	43.55	47.63	37.42	37.12	38.24	37.61	36.16	37.11	37.95	38.43	38.82	38.63	37.64
TiO2	0.43	0.31	0.39	0.26	0.55	0.96	0.01	1.12	0.19	0.82	0.80	1.19	0.57	0.49	0.57	0.71	0.47	0.50	0.51	0.65
Al2O3	3.51	3.24	3.44	8.57	7.11	7.88	0.63	4.63	1.53	8.53	8.75	7.93	8.60	9.58	8.55	8.04	9.08	9.00	8.99	8.64
FeO	23.83	24.64	25.48	23.33	24.06	23.73	34.36	30.21	34.85	33.53	34.37	34.47	34.38	33.82	34.48	34.22	34.83	33.75	34.35	33.92
MnO	0.61	0.54	0.60	0.63	0.74	0.72	0.71	0.47	0.69	0.55	0.54	0.59	0.62	0.50	0.62	0.60	0.64	0.87	0.81	0.84
MgO	7.61	7.47	7.09	7.95	7.95	7.39	0.59	4.11	0.95	0.04	0.04	0.03	0.03	0.02	0.02	0.04	0.18	0.16	0.22	0.23
ZnO	0.20	0.19	0.15	0.04	0.14	0.05	-	-	-	-	-	-	-	-	-	-	0.03	0.16	0.16	-
CaO	8.64	9.40	8.85	10.94	10.83	10.79	10.19	9.11	10.92	10.27	10.56	10.41	10.62	10.47	10.62	10.27	10.46	10.56	10.60	10.67
Na2O	2.26	1.60	2.38	1.39	1.35	1.55	0.17	2.30	0.37	1.74	1.72	1.80	1.70	1.61	1.66	1.82	1.33	1.44	1.46	1.25
K2O	0.78	0.74	0.80	0.96	0.81	0.96	0.14	0.89	0.26	1.86	2.00	1.54	1.73	2.09	2.02	1.67	2.30	2.18	1.97	2.20
H2O#	1.50	1.58	1.26	1.68	1.78	1.77	1.84	1.14	1.81	0.90	0.94	1.19	1.07	0.81	0.79	1.12	0.98	0.96	1.05	1.26
F	0.85	0.76	1.39	0.52	0.40	0.33	0.06	1.30	0.04	0.48	0.33	0.51	0.52	0.47	0.44	0.53	0.11	0.22	0.21	0.16
Cl	0.16	0.10	0.10	0.21	0.11	0.22	0.04	0.35	0.08	2.44	2.57	1.31	1.72	2.73	2.92	1.49	3.01	2.86	2.55	1.68
O=F+Cl	-0.40	-0.30	-0.60	-0.30	-0.20	-0.20	-0.04	-0.60	-0.04	-0.80	-0.70	-0.50	-0.60	-0.80	-0.80	-0.60	-0.70	-0.70	-0.70	-0.40
Total	99.33	100.07	100.03	101.71	102.39	101.79	99.47	98.54	99.29	97.82	99.01	98.71	98.58	97.95	98.96	97.90	101.12	100.75	100.85	98.71
Structural formula on the basis of 24(O,OH,F,Cl)																				
IVSi	7.59	7.63	7.53	6.88	7.03	6.91	8.10	7.09	7.73	6.44	6.35	6.49	6.42	6.26	6.38	6.50	6.43	6.48	6.45	6.41
IVAl	0.41	0.37	0.47	1.12	0.97	1.09	-	0.89	0.27	1.56	1.65	1.51	1.58	1.74	1.62	1.50	1.57	1.52	1.55	1.59
IVTi	-	-	-	-	-	-	-	0.02	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.10	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.23	0.21	0.15	0.40	0.29	0.32	0.12	-	0.03	0.17	0.12	0.07	0.15	0.22	0.11	0.13	0.22	0.25	0.22	0.15
Ti	0.05	0.04	0.04	0.03	0.06	0.11	0.00	0.11	0.02	0.11	0.10	0.15	0.07	0.06	0.07	0.09	0.06	0.06	0.06	0.08
Mg	1.75	1.70	1.63	1.79	1.78	1.67	0.14	1.00	0.23	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.04	0.04	0.06	0.06
Fe2+	2.97	3.05	3.17	2.77	2.87	2.90	4.59	3.89	4.72	4.71	4.77	4.77	4.77	4.72	4.81	4.77	4.68	4.64	4.66	4.71
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	-	-	0.10	-	-	-	-	-	-	-	-	-	-	-	-	-
Ca	-	-	-	-	-	-	0.05	-	-	-	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Fe2+	0.10	0.10	0.12	0.18	0.15	0.10	-	0.22	0.02	0.12	0.15	0.13	0.14	0.18	0.14	0.13	0.19	0.07	0.13	0.12
Zn	0.02	0.02	0.02	0.00	0.02	0.01	-	-	-	-	-	-	-	-	-	-	0.00	0.02	0.02	-
Mn	0.08	0.07	0.08	0.08	0.09	0.09	-	0.06	0.09	0.08	0.08	0.08	0.09	0.07	0.09	0.09	0.09	0.12	0.11	0.12
Ca	1.43	1.54	1.46	1.74	1.74	1.75	1.69	1.59	1.89	1.80	1.78	1.79	1.77	1.75	1.77	1.78	1.72	1.79	1.74	1.76
Na	0.38	0.26	0.32	-	-	0.05	0.05	0.12	-	-	-	-	-	-	-	-	-	-	-	-
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	1.74	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	0.03	0.01	-	-	-	0.01	0.09	0.16	0.10	0.17	0.20	0.19	0.11	0.16	0.10	0.16	0.19
Na	0.30	0.21	0.39	0.41	0.39	0.41	-	0.60	0.12	0.58	0.57	0.59	0.56	0.54	0.55	0.60	0.43	0.47	0.47	0.41
K	0.15	0.15	0.16	0.18	0.16	0.19	0.03	0.18	0.05	0.41	0.44	0.33	0.38	0.46	0.44	0.37	0.49	0.46	0.42	0.48
A site	0.45	0.35	0.55	0.62	0.56	0.60	0.03	0.78	0.18	1.08	1.17	1.03	1.11	1.20	1.18	1.08	1.08	1.03	1.05	1.08
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.54	1.61	1.29	1.69	1.78	1.79	1.96	1.23	1.96	1.03	1.08	1.35	1.22	0.94	0.91	1.28	1.09	1.07	1.17	1.43
F	0.42	0.37	0.68	0.25	0.19	0.16	0.03	0.67	0.02	0.26	0.18	0.27	0.28	0.26	0.24	0.29	0.06	0.12	0.11	0.09
Cl	0.04	0.02	0.03	0.05	0.03	0.06	0.01	0.10	0.02	0.71	0.75	0.38	0.50	0.80	0.85	0.43	0.85	0.81	0.72	0.48
F/FM	0.64	0.65	0.67	0.62	0.63	0.64	0.97	0.80	0.95	1.00	1.00	1.00	1.00	1.00	1.00	1.00	0.99	0.99	0.99	0.99

	40A.5	40A.6	40A.7	40A.8	40A.9	40A.10	51.1	51.2	51.3	51.4	51.5	51.6	55.1	55.2	55.3	55.4	55.5	55.6	56.1	56.2
SiO2	36.22	36.61	36.79	37.87	37.51	37.70	41.83	42.71	40.51	41.46	40.64	41.57	39.67	39.81	38.65	38.98	39.73	39.52	30.31	39.38
TiO2	0.43	0.43	0.31	0.33	0.52	0.43	0.80	0.71	1.48	0.87	0.94	0.94	1.41	1.38	0.54	0.50	1.75	1.60	1.65	1.73
Al2O3	8.77	9.02	9.11	8.36	8.37	7.83	4.93	4.80	6.45	5.71	6.41	5.71	7.81	7.82	8.74	8.47	7.72	7.71	7.78	7.61
FeO	33.92	33.85	34.05	34.15	34.02	34.14	35.58	35.82	33.88	35.16	34.94	33.86	32.82	32.10	33.39	33.56	31.88	32.19	30.43	30.42
MnO	0.86	0.78	0.72	0.71	0.71	0.84	0.78	0.80	0.68	0.79	0.65	0.70	0.51	0.47	0.90	0.89	0.61	0.48	0.48	0.48
MgO	0.15	0.16	0.14	0.17	0.14	0.19	0.03	0.05	0.05	0.05	0.03	0.03	1.31	1.35	0.34	0.36	1.36	1.04	1.24	1.24
ZnO	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
CaO	10.56	10.35	10.68	10.69	10.65	10.48	9.49	9.60	9.75	9.60	9.77	9.67	10.16	10.16	10.05	10.09	10.10	10.15	10.14	10.09
Na2O	1.25	1.36	1.55	1.40	1.39	1.42	1.56	1.61	1.78	1.69	1.85	1.74	2.00	2.02	1.83	1.86	2.13	2.04	2.03	2.10
K2O	2.26	2.21	2.17	1.94	1.89	2.04	0.94	0.86	1.15	0.96	1.17	1.01	1.41	1.40	1.59	1.46	1.36	1.45	1.48	1.44
H2O#	0.88	0.95	0.94	1.09	1.01	0.99	1.50	1.53	1.53	1.47	1.40	1.48	1.25	1.26	1.28	1.29	1.30	1.34	0.94	1.19
F	0.17	0.27	0.32	0.34	0.33	0.23	0.43	0.40	0.26	0.48	0.57	0.46	0.86	0.82	0.47	0.53	0.75	0.65	0.85	0.76
Cl	2.99	2.59	2.55	1.96	2.26	2.51	0.31	0.30	0.50	0.35	0.47	0.33	0.62	0.61	1.06	0.96	0.60	0.61	0.80	0.87
O=F+Cl	-0.70	-0.70	-0.70	-0.60	-0.70	-0.70	-0.20	-0.20	-0.20	-0.30	-0.30	-0.30	-0.50	-0.50	-0.40	-0.40	-0.50	-0.40	-0.50	-0.50
Total	97.71	97.88	98.61	98.44	98.17	98.33	97.92	98.96	97.80	98.32	98.48	97.23	99.34	98.71	98.41	98.50	98.84	98.35	87.58	96.80
Structural formula on the basis of 24(O,OH,F,Cl)																				
IVS1	6.32	6.34	6.33	6.48	6.45	6.50	7.06	7.11	6.82	6.96	6.83	7.02	6.57	6.61	6.52	6.56	6.59	6.60	5.89	6.65
IVal	1.68	1.66	1.67	1.52	1.55	1.50	0.94	0.89	1.18	1.04	1.17	0.98	1.43	1.39	1.48	1.44	1.41	1.40	1.78	1.35
IVT1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.24	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	7.91	8.00
VIA1	0.12	0.18	0.17	0.16	0.14	0.08	0.04	0.06	0.11	0.09	0.10	0.15	0.10	0.14	0.26	0.24	0.10	0.12	-	0.16
Ti	0.06	0.06	0.04	0.04	0.07	0.06	0.10	0.09	0.19	0.11	0.12	0.12	0.18	0.17	0.07	0.06	0.22	0.20	-	0.22
Mg	0.04	0.04	0.04	0.04	0.04	0.05	0.01	0.01	0.01	0.01	0.01	0.01	0.32	0.33	0.09	0.09	0.34	0.26	0.36	0.31
Fe2+	4.78	4.72	4.75	4.75	4.75	4.81	4.85	4.84	4.69	4.79	4.77	4.72	4.40	4.35	4.59	4.60	4.35	4.42	4.64	4.29
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.02
Ca	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Fe2+	0.16	0.18	0.14	0.13	0.14	0.14	0.17	0.15	0.08	0.15	0.14	0.06	0.14	0.11	0.13	0.12	0.07	0.07	0.31	-
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	0.13	0.11	0.10	0.10	0.10	0.12	0.11	0.11	0.10	0.11	0.09	0.10	0.07	0.07	0.13	0.13	0.09	0.07	0.08	0.05
Ca	1.71	1.71	1.75	1.76	1.76	1.74	1.71	1.71	1.76	1.73	1.76	1.75	1.78	1.81	1.75	1.75	1.79	1.82	1.62	1.82
Na	-	-	-	-	-	-	0.01	0.03	0.07	0.01	0.01	0.10	-	0.02	-	-	0.05	0.04	-	0.12
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	0.26	0.21	0.21	0.19	0.20	0.19	-	-	-	-	-	-	0.02	-	0.07	0.07	-	-	0.50	-
Na	0.42	0.46	0.52	0.46	0.46	0.47	0.50	0.49	0.51	0.54	0.59	0.47	0.64	0.63	0.60	0.61	0.64	0.62	0.76	0.57
K	0.50	0.49	0.48	0.42	0.41	0.45	0.20	0.18	0.25	0.21	0.25	0.22	0.30	0.30	0.34	0.31	0.29	0.31	0.37	0.31
A site	1.19	1.16	1.21	1.08	1.08	1.12	0.71	0.68	0.76	0.74	0.84	0.69	0.96	0.92	1.01	0.99	0.93	0.93	1.63	0.88
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.02	1.09	1.08	1.25	1.16	1.14	1.68	1.70	1.72	1.65	1.57	1.66	1.38	1.40	1.45	1.44	1.44	1.49	1.21	1.34
F	0.09	0.15	0.18	0.19	0.18	0.12	0.23	0.21	0.14	0.25	0.30	0.24	0.45	0.43	0.25	0.28	0.39	0.34	0.52	0.41
Cl	0.88	0.76	0.74	0.57	0.66	0.73	0.09	0.08	0.14	0.10	0.13	0.09	0.17	0.17	0.30	0.27	0.17	0.17	0.26	0.25
F/FH	0.99	0.99	0.99	0.99	0.99	0.99	1.00	1.00	1.00	1.00	1.00	1.00	0.93	0.93	0.98	0.98	0.93	0.93	0.93	0.93

	56.3	56.4	56.5	56.6	56.7	56.8	56.9	56.10	56.11	68A.1	68A.2	68A.3	68B.1	68B.2	68B.3	68B.4	79.1	79.2	79.3	79.4
SiO2	39.45	39.25	39.22	39.90	39.67	39.56	39.89	39.65	34.14	39.29	39.18	39.50	40.21	39.48	38.14	42.46	39.98	39.99	39.48	40.08
TiO2	1.58	1.81	1.88	1.61	1.71	1.60	1.06	1.60	2.19	1.72	1.91	1.87	1.41	0.28	0.27	0.16	1.64	1.73	1.52	1.38
Al2O3	7.69	7.83	7.83	7.28	7.57	7.79	7.75	7.77	11.49	7.39	7.28	7.36	6.97	9.22	10.32	6.01	7.52	7.52	7.29	7.67
FeO	32.50	32.10	32.58	32.16	32.50	31.46	34.07	32.80	33.24	34.90	34.11	33.94	33.70	33.93	33.42	35.01	31.56	30.90	31.67	30.71
MnO	0.44	0.47	0.47	0.49	0.59	0.65	0.74	0.82	0.78	0.75	0.71	0.70	0.89	0.86	0.91	1.14	0.68	0.60	0.53	0.52
MgO	1.29	1.14	1.14	1.19	1.10	1.10	0.79	0.97	0.64	0.09	0.06	0.06	0.13	0.04	0.04	0.03	1.47	1.54	1.98	1.56
ZnO	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
CaO	10.01	10.13	10.04	9.93	9.88	10.04	10.10	10.09	10.01	9.78	9.30	9.84	9.58	10.63	10.19	9.68	10.23	10.08	10.14	10.01
Na2O	2.12	2.07	2.11	2.09	2.12	2.01	1.99	1.90	2.03	1.54	1.81	1.58	1.48	1.22	1.23	0.79	2.02	1.98	1.97	1.96
K2O	1.37	1.48	1.48	1.33	1.40	1.34	1.41	1.41	1.13	1.32	1.22	1.38	1.14	1.19	1.48	0.72	1.42	1.33	1.41	1.39
H2O#	1.24	1.27	1.29	1.27	1.28	1.31	1.41	1.39	1.36	1.47	1.44	1.47	1.62	1.71	1.61	1.75	1.29	1.32	1.16	1.33
F	0.78	0.59	0.54	0.61	0.63	0.53	0.42	0.43	0.45	0.24	0.37	0.26	0.10	0.11	0.24	0.03	0.66	0.63	0.99	0.53
Cl	0.75	0.96	1.00	0.92	0.85	0.89	0.81	0.83	0.70	0.79	0.63	0.76	0.42	0.17	0.23	0.13	0.81	0.71	0.63	0.82
O=F+Cl	-0.50	-0.50	-0.50	-0.50	-0.50	-0.40	-0.40	-0.40	-0.30	-0.30	-0.30	-0.30	-0.10	-0.08	-0.20	-0.04	-0.50	-0.40	-0.60	-0.40
Total	98.71	98.64	99.12	98.33	98.84	97.85	100.07	99.28	97.82	99.00	97.71	98.43	97.52	98.75	97.92	97.86	98.82	97.90	98.23	97.56
Structural formula on the basis of 24(O,OH,F,Cl)																				
IVSi	6.58	6.55	6.53	6.66	6.61	6.62	6.60	6.58	5.83	6.60	6.63	6.64	6.78	6.57	6.42	7.09	6.63	6.66	6.59	6.69
IVAl	1.42	1.45	1.47	1.34	1.39	1.38	1.40	1.42	2.17	1.40	1.37	1.36	1.22	1.43	1.58	0.91	1.37	1.34	1.41	1.31
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	0.09	0.09	0.06	0.10	0.09	0.16	0.11	0.10	0.14	0.06	0.09	0.10	0.16	0.38	0.46	0.28	0.10	0.14	0.03	0.20
Ti	0.20	0.23	0.24	0.20	0.21	0.20	0.13	0.20	0.28	0.22	0.24	0.24	0.18	0.03	0.03	0.02	0.20	0.22	0.19	0.17
Mg	0.32	0.28	0.28	0.30	0.27	0.27	0.20	0.24	0.16	0.02	0.01	0.02	0.03	0.01	0.01	0.01	0.36	0.38	0.49	0.39
Fe2+	4.40	4.40	4.42	4.40	4.42	4.36	4.56	4.46	4.41	4.70	4.65	4.65	4.62	4.58	4.50	4.70	4.33	4.26	4.29	4.24
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Ca	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Fe2+	0.13	0.08	0.12	0.09	0.10	0.04	0.15	0.09	0.33	0.20	0.18	0.12	0.13	0.15	0.20	0.20	0.04	0.04	0.14	0.05
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	0.06	0.07	0.07	0.07	0.08	0.09	0.10	0.11	0.11	0.11	0.10	0.10	0.13	0.12	0.13	0.16	0.10	0.08	0.08	0.07
Ca	1.79	1.81	1.79	1.78	1.76	1.80	1.74	1.79	1.56	1.69	1.69	1.77	1.73	1.73	1.67	1.64	1.82	1.80	1.79	1.79
Na	0.02	0.04	0.03	0.06	0.05	0.06	-	-	-	-	0.03	0.01	0.01	-	-	-	0.04	0.08	-	0.08
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	-	-	-	0.05	0.00	0.27	0.06	-	-	0.17	0.17	0.09	-	-	-	0.03	-
Na	0.67	0.63	0.65	0.61	0.63	0.59	0.64	0.61	0.67	0.50	0.56	0.51	0.47	0.39	0.40	0.25	0.61	0.56	0.64	0.55
K	0.29	0.32	0.31	0.28	0.30	0.29	0.30	0.30	0.25	0.28	0.26	0.30	0.24	0.25	0.32	0.15	0.30	0.28	0.30	0.30
A site	0.96	0.95	0.97	0.90	0.93	0.87	0.98	0.91	1.19	0.85	0.82	0.80	0.71	0.81	0.89	0.50	0.91	0.84	0.97	0.85
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.38	1.42	1.43	1.42	1.43	1.47	1.55	1.54	1.55	1.65	1.62	1.64	1.82	1.90	1.81	1.95	1.43	1.47	1.30	1.49
F	0.41	0.31	0.28	0.32	0.33	0.28	0.22	0.22	0.24	0.13	0.20	0.14	0.05	0.06	0.13	0.02	0.35	0.33	0.52	0.28
Cl	0.21	0.27	0.28	0.26	0.24	0.25	0.23	0.23	0.20	0.22	0.18	0.22	0.12	0.05	0.06	0.04	0.23	0.20	0.18	0.23
F/FH	0.93	0.94	0.94	0.94	0.94	0.94	0.96	0.95	0.97	1.00	1.00	1.00	0.99	1.00	1.00	1.00	0.92	0.92	0.90	0.92

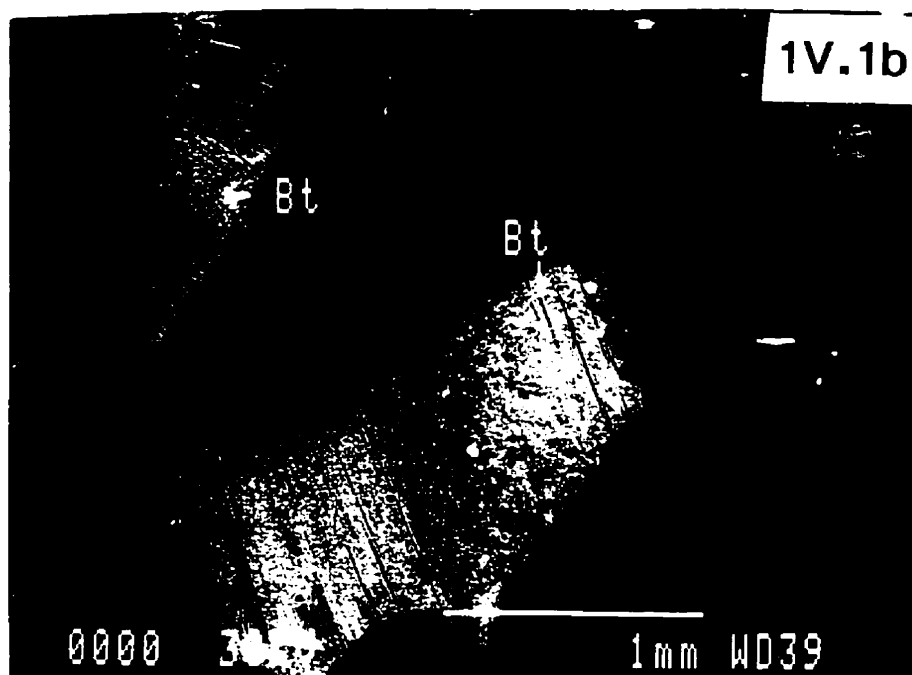
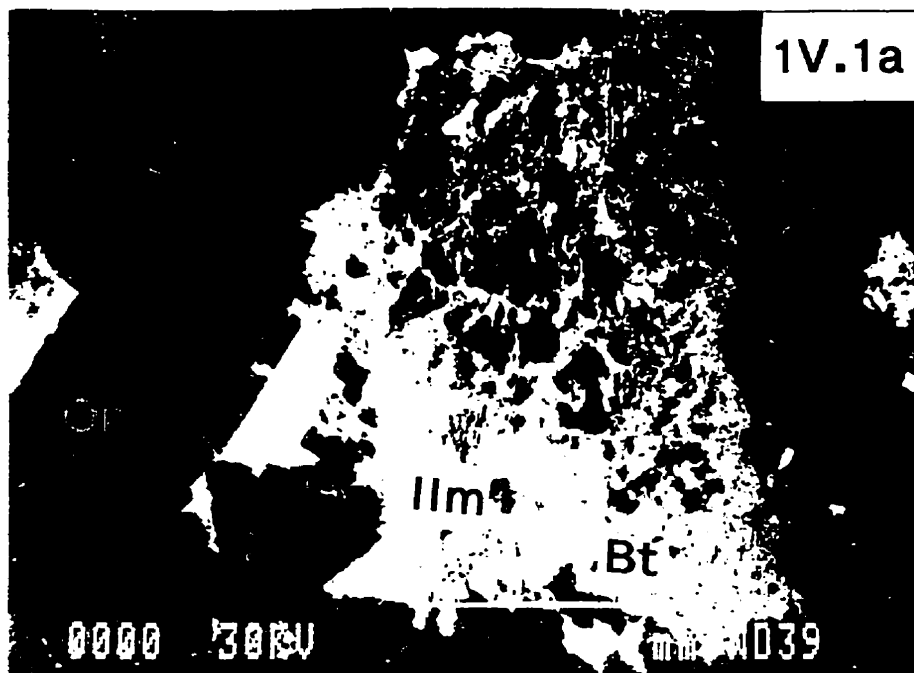
	79.5	79.6	79.7	79.8	79.9	79.10
SiO2	39.77	39.77	40.05	40.04	40.81	40.56
TiO2	1.63	1.57	1.75	1.76	1.61	1.58
Al2O3	7.53	7.51	7.65	7.62	7.15	7.15
FeO	30.39	30.95	29.02	29.96	28.82	29.76
MnO	0.44	0.51	0.48	0.64	0.52	0.54
MgO	1.71	1.65	2.06	1.85	2.34	2.28
ZnO	-	-	-	-	-	-
CaO	9.99	9.99	10.14	10.02	9.99	10.17
Na2O	2.08	1.99	2.00	2.01	2.07	2.05
K2O	1.42	1.43	1.39	1.38	1.25	1.32
H2O	1.12	1.14	1.20	1.21	1.16	1.16
F	0.95	0.90	0.95	0.94	1.09	1.10
Cl	0.85	0.87	0.59	0.58	0.50	0.52
O=F+Cl	-0.60	-0.60	-0.50	-0.50	-0.60	-0.60
Total	97.31	97.71	96.75	97.49	96.72	97.61

Structural formula on the basis of 24(O,OH,F,Cl)

IVS1	6.66	6.65	6.69	6.67	6.79	6.73
IVAl	1.34	1.35	1.31	1.33	1.21	1.27
T site	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.15	0.13	0.20	0.16	0.19	0.13
Ti	0.21	0.20	0.22	0.22	0.20	0.20
Mg	0.43	0.41	0.51	0.46	0.58	0.56
Fe2+	4.22	4.26	4.05	4.16	4.01	4.11
Zn	-	-	-	-	-	-
Mn	-	-	0.02	-	0.01	-
Ca	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00
Fe2+	0.04	0.07	-	0.01	-	0.02
Zn	-	-	-	-	-	-
Mn	0.06	0.07	0.05	0.09	0.06	0.08
Ca	1.79	1.79	1.82	1.79	1.78	1.81
Na	0.10	0.07	0.13	0.11	0.16	0.10
M4 site	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	-	-	-
Na	0.57	0.58	0.51	0.54	0.51	0.56
K	0.30	0.30	0.30	0.29	0.26	0.28
A site	0.88	0.89	0.81	0.84	0.77	0.84
O	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.25	1.28	1.33	1.34	1.29	1.28
F	0.51	0.48	0.50	0.50	0.57	0.58
Cl	0.24	0.25	0.17	0.16	0.14	0.15
F/FM	0.91	0.91	0.89	0.90	0.87	0.88

IV.1a: SEM - BSE image of annitic biotite (Bt) in hornblende biotite granite porphyry (sample 68). The subhedral mica on the lower half is interstitial (late-magmatic) and bears inclusions of ilmenite. The upper half of the mica grain is comprised of anastomosing mica filling fractures in brecciated orthoclase microperthite (Or), suggesting continued growth of annitic mica at subsolidus stages.

IV.1b: SEM - BSE image of zoned zinnwaldite-trilithionite in drusy biotite granite (sample 37). The grain to the top left corner (enclosed in alkali feldspar, not discernible on picture) preserves distinct zoning from low BSE (darker) core of trilithionite to a higher BSE (lighter) margin. This suggests an earlier Li and F-rich fluid which escaped to yield a drusy granite and led to the crystallization of zinnwaldite (lower F and Li) in marginal parts. Another grain of mica (Bt) in the lower part is in the matrix and is characterized by patches of low-BSE mica (trilithionite) within predominantly brighter zinnwaldite. This suggests homogenization of zoning in mica grains not "locked-up" in other minerals.



Appendix IV.2: Results of EMPA of mica in subvolcanic rocks. Rhyolite porphyry (49A), fayalite syenogranite (47A, 47B, & 58), hornblende biotite granite porphyry (51, 55, 56, 68A, 68B, & 79), and biotite granite porphyry (38).

	49A.1	49A.2	49A.3	47A.1	47A.2	47A.3	47A.4	47A.5	47B.1	47B.2	47B.3	47B.4	47B.5	58.1	58.2	51.1	55.1	55.2	55.3	55.4
SiO2	41.92	36.51	36.43	34.68	34.97	35.22	35.10	34.86	34.45	34.10	34.56	33.50	34.48	34.26	24.47	35.02	31.97	33.94	33.79	33.60
TiO2	-	2.52	2.71	2.75	2.30	2.64	2.50	2.53	3.23	2.97	1.73	3.05	3.02	3.52	0.15	1.44	2.22	2.67	2.80	3.06
Al2O3	6.61	10.69	10.74	12.12	12.17	11.94	12.10	11.92	12.13	12.55	12.37	11.34	11.67	12.18	18.81	12.51	12.13	11.65	11.65	11.76
FeO	36.46	33.09	33.38	35.96	35.94	36.34	36.94	36.45	35.74	35.57	36.58	37.56	36.74	35.03	45.39	36.87	37.80	37.26	36.80	36.25
MnO	0.60	0.22	0.23	0.29	0.27	0.28	0.33	0.23	0.27	0.34	0.24	0.22	0.35	0.57	0.11	0.63	0.58	0.56	0.54	0.51
MgO	1.65	3.17	2.48	0.82	0.85	0.91	1.08	1.22	0.09	0.24	0.08	0.06	0.14	0.89	0.06	0.05	0.67	0.47	0.48	0.52
CaO	0.23	0.09	0.02	-	-	0.01	0.01	0.00	0.06	-	0.04	0.04	0.04	0.00	0.02	0.01	0.29	0.02	-	0.01
Na2O	0.09	0.06	0.04	0.01	0.04	0.07	0.01	0.05	0.06	0.04	0.00	0.02	0.06	0.10	0.01	0.04	0.04	0.08	0.06	0.06
K2O	2.60	7.64	8.52	9.09	8.79	8.96	8.84	8.95	9.19	9.39	8.94	8.74	9.30	9.11	0.10	8.64	7.82	9.15	9.32	9.70
H2O#	3.84	3.46	3.50	3.13	3.18	2.99	2.99	3.01	3.09	2.96	3.06	3.04	2.91	3.37	3.50	3.55	3.27	3.14	3.14	3.13
F	0.00	0.18	0.18	0.41	0.38	0.70	0.74	0.71	0.34	0.47	0.16	0.31	0.39	0.06	0.06	0.03	0.05	0.40	0.43	0.44
Cl	0.03	0.65	0.45	1.08	1.02	1.13	1.03	1.02	1.36	1.55	1.71	1.39	1.83	0.87	0.08	0.25	0.95	0.90	0.87	0.86
O=F+Cl	-0.01	-0.20	-0.20	-0.40	-0.40	-0.50	-0.50	-0.50	-0.40	-0.50	-0.50	-0.40	-0.60	-0.20	-0.04	-0.07	-0.20	-0.40	-0.40	-0.40
Total	94.02	98.06	98.50	99.94	99.51	100.65	101.15	100.42	99.56	99.63	99.03	98.82	100.35	99.74	92.73	98.95	97.57	99.87	99.49	99.52
Li2O*	2.48	0.93	0.90	0.40	0.48	0.56	0.52	0.45	0.33	0.24	0.37	0.06	0.34	0.28	-	0.50	-	0.19	0.15	0.09
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	6.78	5.94	5.94	5.73	5.78	5.76	5.73	5.73	5.74	5.69	5.80	5.71	5.74	5.67	4.46	5.82	5.53	5.69	5.69	5.66
IVAl	1.22	2.05	2.06	2.27	2.22	2.24	2.27	2.27	2.26	2.31	2.20	2.28	2.26	2.33	3.54	2.18	2.47	2.30	2.31	2.34
IVTi	-	0.01	0.00	-	-	-	-	-	-	-	-	0.01	-	-	-	-	-	0.00	-	0.00
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.04	-	-	0.09	0.15	0.07	0.05	0.04	0.12	0.16	0.25	-	0.03	0.05	0.50	0.27	0.00	-	0.00	-
VITi	-	0.30	0.33	0.34	0.29	0.33	0.31	0.31	0.40	0.37	0.22	0.38	0.38	0.44	0.02	0.18	0.29	0.33	0.35	0.39
Fe2+	4.93	4.50	4.55	4.97	4.97	4.97	5.04	5.01	4.98	4.97	5.14	5.35	5.12	4.85	6.92	5.13	5.47	5.23	5.18	5.11
Mn2+	0.08	0.03	0.03	0.04	0.04	0.04	0.05	0.03	0.04	0.05	0.03	0.03	0.05	0.08	0.02	0.09	0.09	0.08	0.08	0.07
Mg	0.40	0.77	0.60	0.20	0.21	0.22	0.26	0.30	0.02	0.06	0.02	0.02	0.03	0.22	0.02	0.01	0.17	0.12	0.12	0.13
Li	1.61	0.61	0.59	0.27	0.32	0.37	0.34	0.30	0.22	0.16	0.25	0.04	0.23	0.19	-	0.33	-	0.13	0.10	0.06
Y site	7.07	6.21	6.10	5.92	5.98	5.99	6.05	6.00	5.78	5.77	5.91	5.82	5.84	5.82	7.48	6.01	6.02	5.89	5.84	5.76
Ca	0.04	0.02	0.00	-	-	0.00	0.00	0.00	0.01	-	0.01	0.01	0.01	0.00	0.00	0.00	0.05	0.00	-	0.00
Na	0.03	0.02	0.01	0.00	0.01	0.02	0.00	0.02	0.02	0.01	0.00	0.01	0.02	0.03	0.01	0.01	0.01	0.03	0.02	0.02
K	0.54	1.59	1.77	1.92	1.85	1.87	1.84	1.88	1.95	2.00	1.92	1.90	1.98	1.92	0.02	1.83	1.72	1.96	2.00	2.09
A site	0.60	1.62	1.79	1.92	1.86	1.90	1.85	1.90	1.98	2.01	1.92	1.92	2.00	1.96	0.03	1.85	1.79	1.99	2.02	2.11
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.99	3.73	3.78	3.48	3.52	3.33	3.33	3.35	3.44	3.32	3.43	3.43	3.28	3.73	3.94	3.91	3.69	3.53	3.53	3.52
F	0.00	0.09	0.09	0.22	0.20	0.36	0.38	0.37	0.18	0.25	0.08	0.17	0.20	0.03	0.03	0.02	0.03	0.21	0.23	0.23
Cl	0.01	0.18	0.12	0.30	0.29	0.31	0.28	0.28	0.38	0.44	0.49	0.40	0.52	0.24	0.02	0.07	0.28	0.26	0.25	0.25
X(F)	0.00	0.02	0.02	0.05	0.05	0.09	0.10	0.09	0.04	0.06	0.02	0.04	0.05	0.01	0.01	0.00	0.01	0.05	0.06	0.06
X(Cl)	0.00	0.04	0.03	0.08	0.07	0.08	0.07	0.07	0.10	0.11	0.12	0.10	0.13	0.06	0.01	0.02	0.07	0.06	0.06	0.06
X(F/Cl)	0.32	0.53	0.74	0.71	0.70	1.15	1.35	1.30	0.46	0.56	0.17	0.42	0.39	0.13	1.35	0.24	0.10	0.83	0.92	0.95
F/FM	0.93	0.85	0.88	0.96	0.96	0.96	0.95	0.94	1.00	0.99	1.00	1.00	0.99	0.96	1.00	1.00	0.97	0.98	0.98	0.98
X(Phl)	0.06	0.12	0.10	0.03	0.04	0.04	0.04	0.05	0.00	0.01	0.00	0.00	0.01	0.04	0.00	0.00	0.03	0.02	0.02	0.02
X(Ann)	0.94	0.82	0.84	0.65	0.65	0.69	0.66	0.67	0.66	0.60	0.64	0.72	0.71	0.63	-	0.64	0.55	0.68	0.68	0.66
X(Srp)	-	0.05	0.06	0.32	0.31	0.27	0.29	0.28	0.34	0.39	0.36	0.28	0.28	0.34	1.08	0.36	0.42	0.30	0.30	0.32

	55.5	55.6	55.7	55.8	55.9	56.1	56.2	56.3	56.4	56.5	56.6	68A.1	68A.2	68A.3	68A.4	68A.5	68B.1	68B.2	68B.3	79.1
SiO2	33.92	33.69	32.90	33.39	34.26	34.04	34.16	33.39	33.42	34.14	32.23	32.90	32.98	32.71	32.92	33.04	32.75	32.58	33.37	34.28
TiO2	-	2.98	2.19	2.12	2.25	3.52	3.08	2.93	2.73	2.19	2.45	3.52	3.36	2.95	3.01	3.40	3.22	3.13	3.46	2.02
Al2O3	11.60	11.36	12.28	12.38	12.18	12.49	11.57	11.36	11.86	11.49	12.19	12.10	12.00	12.21	11.95	12.20	12.35	13.24	12.09	11.63
FeO	36.67	37.07	38.00	35.18	35.03	35.89	35.33	36.52	36.22	35.99	35.50	36.38	35.50	37.27	37.63	34.76	36.49	35.87	35.91	35.34
MnO	0.48	0.49	0.55	0.48	0.57	0.31	0.39	0.70	0.67	0.45	0.36	0.51	0.52	0.42	0.51	0.43	0.46	0.40	0.51	0.42
MgO	0.57	0.56	0.69	0.74	0.89	1.29	1.38	0.93	0.14	0.79	1.06	0.04	0.03	0.06	0.10	0.08	0.06	0.13	0.07	2.53
CaO	0.02	0.04	0.28	0.01	0.00	0.19	0.20	0.08	0.14	0.05	0.12	-	-	0.09	-	0.09	0.13	0.08	0.01	0.06
Na2O	0.23	0.08	0.05	0.06	0.10	0.05	0.04	0.04	0.03	0.04	0.03	0.02	0.02	0.04	0.05	0.05	0.03	0.03	0.02	0.02
K2O	8.97	9.08	8.06	8.84	9.11	8.20	8.39	8.81	8.36	8.64	7.80	9.23	9.36	8.66	8.41	8.40	8.68	8.60	8.76	8.93
H2O#	3.26	3.10	3.24	3.33	3.36	3.37	3.32	3.26	3.33	3.35	3.33	3.22	3.20	3.21	3.24	3.34	3.25	3.35	3.35	3.24
F	0.14	0.50	0.14	-	0.06	0.19	0.21	0.32	0.18	0.07	0.05	-	-	-	0.05	-	0.06	0.00	-	0.40
Cl	0.90	0.85	0.98	1.00	0.87	0.67	0.76	0.60	0.64	0.81	0.89	1.29	1.40	1.30	1.13	1.00	1.12	0.92	0.93	0.70
O=F+Cl	-0.30	-0.40	-0.30	-0.20	-0.20	-0.20	-0.30	-0.30	-0.20	-0.20	-0.20	-0.30	-0.30	-0.30	-0.30	-0.20	-0.30	-0.20	-0.20	-0.30
Total	96.50	99.40	99.08	97.29	98.46	99.96	98.58	98.67	97.50	97.81	95.78	98.93	98.05	98.64	98.72	96.57	98.33	98.11	98.27	99.23
Li2O*	0.18	0.12	-	0.03	0.28	0.22	0.25	0.03	0.04	0.25	-	-	-	-	-	-	-	-	0.03	0.29
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.89	5.69	5.59	5.71	5.75	5.61	5.72	5.67	5.72	5.79	5.61	5.60	5.65	5.59	5.62	5.69	5.59	5.54	5.67	5.70
IVAl	2.11	2.26	2.41	2.29	2.25	2.39	2.28	2.27	2.28	2.21	2.39	2.40	2.35	2.41	2.38	2.31	2.41	2.46	2.33	2.28
IVTi	-	0.05	-	-	-	-	0.00	0.06	-	-	-	-	-	-	-	-	-	-	-	0.01
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	0.27	-	0.04	0.21	0.16	0.03	-	-	0.11	0.09	0.11	0.02	0.07	0.05	0.02	0.16	0.07	0.19	0.09	-
VIT1	-	0.33	0.28	0.27	0.28	0.44	0.39	0.31	0.35	0.28	0.32	0.45	0.43	0.38	0.39	0.44	0.41	0.40	0.44	0.24
Fe2+	5.33	5.23	5.40	5.04	4.92	4.94	4.94	5.18	5.18	5.11	5.17	5.18	5.09	5.33	5.37	5.00	5.21	5.10	5.10	4.92
Mn2+	0.07	0.07	0.08	0.07	0.08	0.04	0.06	0.10	0.10	0.06	0.05	0.07	0.08	0.06	0.07	0.06	0.07	0.06	0.07	0.06
Mg	0.15	0.14	0.17	0.19	0.22	0.32	0.34	0.24	0.04	0.20	0.27	0.01	0.01	0.02	0.03	0.02	0.02	0.03	0.02	0.63
Li	0.13	0.08	-	0.02	0.19	0.14	0.17	0.02	0.03	0.17	-	-	-	-	-	-	-	-	0.02	0.19
Y site	5.95	5.85	5.97	5.80	5.86	5.91	5.90	5.86	5.80	5.91	5.93	5.73	5.68	5.83	5.87	5.69	5.78	5.78	5.74	6.04
Ca	0.00	0.01	0.05	0.00	0.00	0.03	0.04	0.01	0.03	0.01	0.02	-	-	0.02	-	0.02	0.02	0.02	0.00	0.01
Na	0.08	0.03	0.02	0.02	0.03	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01
K	1.99	1.95	1.75	1.93	1.95	1.72	1.79	1.91	1.83	1.87	1.73	2.00	2.05	1.89	1.83	1.84	1.89	1.87	1.90	1.90
X site	2.07	1.99	1.81	1.95	1.99	1.77	1.84	1.94	1.86	1.89	1.76	2.01	2.05	1.92	1.84	1.88	1.92	1.89	1.90	1.91
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.66	3.49	3.64	3.71	3.72	3.71	3.67	3.66	3.71	3.73	3.71	3.63	3.59	3.62	3.65	3.71	3.64	3.73	3.73	3.59
F	0.08	0.26	0.08	-	0.03	0.10	0.11	0.17	0.10	0.04	0.03	-	-	-	0.03	-	0.03	0.00	-	0.21
Cl	0.27	0.24	0.28	0.29	0.25	0.19	0.22	0.17	0.19	0.23	0.26	0.37	0.41	0.38	0.33	0.29	0.32	0.26	0.27	0.20
X(F)	0.02	0.07	0.02	-	0.01	0.03	0.03	0.04	0.02	0.01	0.01	-	-	-	0.01	-	0.01	0.00	-	0.05
X(Cl)	0.07	0.06	0.07	0.07	0.06	0.05	0.05	0.04	0.05	0.06	0.07	0.09	0.10	0.09	0.08	0.07	0.08	0.07	0.07	0.05
X(F/Cl)	0.29	1.09	0.27	-	0.13	0.54	0.52	0.98	0.54	0.17	0.10	-	-	-	0.09	-	0.11	0.01	-	1.07
F/FM	0.97	0.97	0.97	0.96	0.96	0.94	0.93	0.96	0.99	0.96	0.95	1.00	1.00	1.00	1.00	1.00	1.00	0.99	1.00	0.89
X(Phl)	0.02	0.02	0.03	0.03	0.04	0.05	0.06	0.04	0.01	0.03	0.05	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.10
X(Ann)	0.69	0.71	0.57	0.57	0.63	0.57	0.68	0.68	0.65	0.70	0.54	0.60	0.62	0.58	0.62	0.60	0.57	0.48	0.62	0.64
X(Srp)	0.29	0.27	0.40	0.39	0.34	0.37	0.27	0.28	0.35	0.26	0.41	0.39	0.38	0.41	0.38	0.40	0.43	0.52	0.37	0.26

	79.2	38-1	38-2	38-3	38-4	38-5
SiO2	34.06	35.47	35.48	34.81	34.76	35.81
TiO2	2.31	2.21	2.58	2.21	2.33	2.38
Al2O3	12.03	14.72	14.32	15.23	14.84	15.31
FeO	32.65	31.87	31.92	32.15	33.44	29.99
MnO	0.37	0.42	0.34	0.49	0.25	0.40
MgO	1.91	1.08	1.33	1.07	0.15	1.41
CaO	-	0.01	0.04	0.00	0.02	0.04
Na2O	0.03	0.13	0.14	0.09	0.09	0.25
K2O	8.71	9.31	9.24	9.27	9.09	8.90
H2O#	3.25	3.13	2.89	3.08	3.45	3.12
F	0.38	0.82	1.31	0.97	0.31	0.97
Cl	0.79	0.70	0.69	0.56	0.32	0.65
O=F+Cl	-0.30	-0.50	-0.70	-0.50	-0.20	-0.60
Total	96.16	99.37	99.59	99.41	98.85	98.67
Li2O*	0.22	0.63	0.63	0.44	0.42	0.72

Structural formula on the basis of 24 (O,OH,F,Cl)

IVSi	5.79	5.73	5.72	5.65	5.69	5.74
IVAl	2.21	2.27	2.28	2.35	2.31	2.26
IVTi	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	0.20	0.53	0.44	0.56	0.55	0.64
VITi	0.30	0.27	0.31	0.27	0.29	0.29
Fe2+	4.64	4.30	4.30	4.36	4.58	4.02
Mn2+	0.05	0.06	0.05	0.07	0.03	0.05
Mg	0.48	0.26	0.32	0.26	0.04	0.34
Li	0.15	0.41	0.41	0.29	0.28	0.47
Y site	5.83	5.83	5.83	5.80	5.77	5.80
Ca	-	0.00	0.01	0.00	0.00	0.01
Na	0.01	0.04	0.05	0.03	0.03	0.08
K	1.89	1.92	1.90	1.92	1.90	1.82
A site	1.90	1.96	1.95	1.95	1.93	1.90
X	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.57	3.39	3.14	3.35	3.75	3.33
F	0.20	0.42	0.67	0.50	0.16	0.49
Cl	0.23	0.19	0.19	0.15	0.09	0.18
X(F)	0.05	0.10	0.17	0.12	0.04	0.12
X(Cl)	0.06	0.05	0.05	0.04	0.02	0.04
X(F/Cl)	0.89	2.18	3.55	3.20	1.85	2.77
F/FM	0.91	0.94	0.93	0.94	0.99	0.92
X(Phl)	0.08	0.04	0.05	0.04	0.01	0.06
X(Ann)	0.60	0.43	0.46	0.38	0.42	0.40
X(Srp)	0.31	0.52	0.49	0.58	0.58	0.55

Appendix IV.3: Results of EMPA (spot) analyses of mica in plutonic rocks. Coarse-grained biotite granite (26A, 26B, 33, & 39C), medium-grained biotite granite (13B, 14, 29, 2A1, & 2B1), fine-grained biotite granite (83, 28B, 30A, & 37), and porphyritic biotite granite (1A1, 1A2, & 1B2).

	26A-1	26A-2	26A-3	26A-4	26A-5	26A-6	26B-1	26B-2	26B-3	26B-4	26B-5	26B-6	26B-7	33-1	33-2	33-3	33-4	33-5	33-6	39C-1
SiO ₂	34.74	35.25	35.83	36.27	35.01	35.96	33.00	35.04	36.19	35.72	35.71	35.83	35.02	35.91	35.85	36.69	36.38	36.09	36.24	37.71
TiO ₂	0.99	1.32	2.41	2.09	1.97	2.55	1.97	2.18	2.36	2.58	2.53	2.55	1.44	1.98	2.42	2.81	2.46	2.76	2.62	2.16
Al ₂ O ₃	16.65	16.18	14.09	14.09	15.63	14.65	15.81	15.56	14.40	14.35	14.38	14.02	15.96	16.68	15.98	14.85	16.28	14.72	15.52	18.19
FeO	32.16	32.26	31.37	31.47	31.62	30.88	33.29	29.57	30.48	30.14	31.59	30.98	30.88	30.40	29.83	30.05	29.15	30.64	29.40	24.05
MnO	0.46	0.36	0.32	0.35	0.33	0.35	0.41	0.33	0.35	0.33	0.40	0.38	0.38	0.51	0.56	0.53	0.46	0.48	0.51	0.23
MgO	1.96	1.97	2.98	3.25	2.12	2.12	2.23	1.93	2.59	2.50	2.00	2.20	2.32	1.63	1.88	2.07	1.78	1.79	1.80	2.26
CaO	0.04	0.01	0.03	0.01	0.01	-	0.01	0.02	-	0.02	0.02	-	0.01	-	0.01	-	0.01	0.03	-	0.02
Na ₂ O	0.09	0.15	0.11	0.09	0.13	0.10	0.10	0.10	0.06	0.09	0.12	0.07	0.09	0.13	0.12	0.10	0.17	0.25	0.12	0.11
K ₂ O	9.21	9.26	9.38	9.11	9.14	9.34	6.97	9.27	8.79	9.03	9.08	9.10	9.42	9.41	9.45	9.45	9.24	9.01	9.46	9.78
H ₂ O ^W	2.47	2.39	2.37	2.32	2.38	2.34	2.83	2.92	2.42	2.43	2.42	2.42	2.73	2.70	2.58	2.55	2.57	2.57	2.70	3.30
F	2.25	2.45	2.58	2.74	2.43	2.59	1.51	1.38	2.52	2.44	2.38	2.42	1.75	2.00	2.32	2.34	2.25	2.13	1.97	1.12
Cl	0.63	0.63	0.48	0.47	0.66	0.60	0.45	0.57	0.48	0.51	0.60	0.53	0.59	0.42	0.26	0.42	0.54	0.60	0.48	0.25
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.70	-0.70	-1.00	-1.00	-1.00	-1.00	-0.90	-0.90	-1.00	-1.00	-1.00	-1.00	-0.90	-0.50
Total	100.57	101.05	100.75	100.99	100.27	100.27	97.85	98.16	99.47	99.02	100.08	99.38	99.73	100.81	100.23	100.77	100.23	100.04	99.88	98.64
Li ₂ O*	0.42	0.56	0.73	0.86	0.50	0.77	-	0.50	0.84	0.70	0.70	0.73	0.50	0.75	0.74	0.98	0.89	0.81	0.85	1.27
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.56	5.60	5.68	5.71	5.60	5.71	5.46	5.68	5.75	5.73	5.71	5.75	5.62	5.63	5.66	5.74	5.70	5.72	5.72	5.78
IVAl	2.44	2.40	2.32	2.29	2.40	2.29	2.54	2.32	2.25	2.27	2.29	2.25	2.38	2.37	2.34	2.26	2.30	2.28	2.28	2.22
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAl	0.70	0.63	0.31	0.33	0.55	0.45	0.54	0.65	0.45	0.44	0.41	0.40	0.64	0.72	0.63	0.48	0.70	0.48	0.61	1.07
VITi	0.12	0.16	0.29	0.25	0.24	0.30	0.25	0.27	0.28	0.31	0.30	0.31	0.17	0.23	0.29	0.33	0.29	0.33	0.31	0.25
Fe ²⁺	4.30	4.29	4.16	4.14	4.23	4.10	4.61	4.01	4.05	4.04	4.22	4.16	4.14	3.99	3.93	3.93	3.82	4.06	3.88	3.09
Mn ²⁺	0.06	0.05	0.04	0.05	0.05	0.05	0.06	0.05	0.05	0.05	0.05	0.05	0.05	0.07	0.08	0.07	0.06	0.06	0.07	0.03
Mg	0.47	0.47	0.71	0.76	0.51	0.50	0.55	0.47	0.61	0.60	0.48	0.53	0.56	0.38	0.44	0.48	0.41	0.42	0.42	0.52
Li	0.27	0.36	0.47	0.54	0.32	0.49	-	0.33	0.53	0.45	0.45	0.47	0.32	0.48	0.47	0.61	0.56	0.51	0.54	0.78
Y site	5.92	5.95	5.98	6.07	5.89	5.90	6.00	5.76	5.98	5.89	5.91	5.91	5.88	5.87	5.83	5.91	5.84	5.87	5.83	5.74
Ca	0.01	0.00	0.01	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	0.05	0.03	0.03	0.04	0.03	0.03	0.03	0.02	0.03	0.04	0.02	0.03	0.04	0.03	0.05	0.08	0.04	0.03	0.03
K	1.88	1.88	1.90	1.83	1.87	1.89	1.47	1.91	1.78	1.85	1.85	1.86	1.93	1.88	1.90	1.89	1.85	1.82	1.90	1.91
X site	1.92	1.92	1.94	1.86	1.91	1.92	1.50	1.95	1.80	1.88	1.89	1.88	1.96	1.92	1.94	1.92	1.90	1.91	1.94	1.95
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.69	2.60	2.58	2.51	2.59	2.54	3.08	3.14	2.61	2.62	2.63	2.63	2.95	2.90	2.77	2.73	2.74	2.77	2.89	3.39
F	1.14	1.23	1.29	1.36	1.23	1.30	0.79	0.71	1.26	1.24	1.20	1.23	0.89	0.99	1.16	1.16	1.11	1.07	0.99	0.54
Cl	0.17	0.17	0.13	0.13	0.18	0.16	0.13	0.16	0.13	0.14	0.16	0.14	0.16	0.11	0.07	0.11	0.14	0.16	0.13	0.07
X(F)	0.28	0.31	0.32	0.34	0.31	0.32	0.20	0.18	0.32	0.31	0.30	0.31	0.22	0.25	0.29	0.29	0.28	0.27	0.25	0.14
X(Cl)	0.04	0.04	0.03	0.03	0.04	0.04	0.03	0.04	0.03	0.04	0.03	0.04	0.04	0.03	0.02	0.03	0.04	0.04	0.03	0.02
X(F/Cl)	6.65	7.31	10.03	10.84	6.88	7.99	6.29	4.56	9.86	8.83	7.37	8.50	5.53	8.86	16.66	10.40	7.78	6.68	7.66	8.22
F/PM	0.90	0.90	0.86	0.84	0.89	0.89	0.89	0.90	0.87	0.87	0.90	0.89	0.88	0.91	0.90	0.89	0.90	0.91	0.90	0.86
X(Phl)	0.08	0.08	0.12	0.13	0.09	0.09	0.09	0.08	0.10	0.10	0.08	0.09	0.09	0.06	0.08	0.08	0.07	0.07	0.07	0.09
X(Ann)	0.27	0.31	0.46	0.47	0.34	0.44	0.27	0.35	0.45	0.44	0.45	0.48	0.32	0.31	0.35	0.44	0.34	0.44	0.39	0.26
X(Srp)	0.65	0.61	0.42	0.41	0.57	0.48	0.64	0.57	0.44	0.46	0.47	0.43	0.59	0.63	0.58	0.47	0.59	0.49	0.54	0.65

	39C-2	39C-3	39C-4	39C-5	39C-6	39C-7	83.1	83.2	83.3	83.4	83.5	83.6	13B1-1	13B1-2	13B1-3	13B1-4	13B1-5	13B1-6	13B1-7	14-1
SiO2	36.01	36.92	36.85	36.78	44.80	36.08	33.88	33.83	33.78	33.96	34.04	34.35	36.94	37.00	36.47	36.47	36.45	36.62	36.06	37.22
TiO2	1.19	2.13	2.66	2.73	0.29	0.50	1.84	2.64	2.73	3.39	3.07	2.15	2.24	2.15	2.55	2.64	2.38	3.20	3.35	0.44
Al2O3	18.94	17.69	17.03	17.29	28.02	19.34	11.19	11.30	11.10	11.16	11.56	10.99	14.20	15.26	14.68	14.79	15.05	13.54	14.28	20.25
FeO	26.91	25.26	26.19	25.79	8.37	25.04	37.28	36.82	36.63	35.19	36.01	36.78	29.06	27.99	29.88	29.25	29.43	29.52	27.94	21.80
MnO	0.46	0.43	0.53	0.44	0.16	0.50	0.39	0.38	0.35	0.32	0.37	0.33	0.62	0.52	0.52	0.38	0.58	0.48	0.50	0.63
MgO	1.08	1.03	1.24	1.11	0.34	1.30	0.06	0.05	0.05	0.05	0.04	0.04	3.69	3.58	2.66	2.51	2.41	2.74	2.36	0.53
CaO	-	0.01	-	-	-	-	-	-	0.02	-	-	-	-	0.03	0.06	0.02	0.03	0.03	-	0.02
Na2O	0.10	0.17	0.16	0.10	0.23	0.03	0.04	0.06	0.07	0.09	0.09	0.06	0.09	0.06	0.09	0.07	0.07	0.13	0.13	0.11
K2O	9.74	9.49	9.60	9.61	10.78	9.48	8.49	8.52	8.60	8.22	8.75	8.24	10.11	10.25	9.88	9.55	9.83	9.65	9.66	10.02
H2O#	2.79	2.58	2.44	2.55	3.23	2.37	2.79	2.74	2.82	2.93	2.83	2.86	2.44	2.58	2.55	2.60	2.71	2.57	2.87	2.34
F	1.98	2.44	2.68	2.48	2.30	2.87	1.29	1.41	1.24	1.22	1.41	1.31	2.56	2.34	2.21	2.18	1.98	2.12	1.59	3.12
Cl	0.29	0.32	0.35	0.34	0.03	0.28	0.55	0.55	0.54	0.31	0.31	0.33	0.45	0.38	0.57	0.49	0.41	0.66	0.49	0.14
O=F+Cl	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-0.70	-0.70	-0.60	-0.60	-0.70	-0.60	-1.00	-1.00	-1.00	-1.00	-0.90	-1.00	-0.80	-1.00
Total	98.61	97.37	98.52	98.11	97.58	96.52	97.13	97.59	97.29	96.27	97.81	96.80	101.23	101.08	101.05	99.91	100.40	100.22	98.47	95.26
Li2O*	0.78	1.04	1.02	1.01	3.31	0.80	0.17	0.16	0.14	0.19	0.22	0.31	1.05	1.07	0.91	0.91	0.91	0.96	0.80	1.13
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.67	5.82	5.78	5.78	6.12	5.75	5.84	5.79	5.80	5.82	5.77	5.89	5.75	5.73	5.71	5.75	5.73	5.78	5.76	5.88
IVAl	2.33	2.18	2.22	2.22	1.88	2.25	2.16	2.21	2.20	2.18	2.23	2.11	2.25	2.27	2.29	2.25	2.27	2.22	2.24	2.12
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	1.19	1.10	0.92	0.98	2.64	1.38	0.11	0.06	0.04	0.08	0.08	0.11	0.36	0.51	0.43	0.49	0.52	0.30	0.46	1.65
VTi	0.14	0.25	0.31	0.32	0.03	0.06	0.24	0.34	0.35	0.44	0.39	0.28	0.26	0.25	0.30	0.31	0.28	0.38	0.40	0.05
Fe2+	3.54	3.33	3.43	3.39	0.96	3.34	5.37	5.27	5.26	5.05	5.11	5.27	3.78	3.62	3.92	3.86	3.87	3.90	3.74	2.88
Mn2+	0.06	0.06	0.07	0.06	0.02	0.07	0.06	0.05	0.05	0.05	0.05	0.05	0.08	0.07	0.07	0.05	0.08	0.06	0.07	0.08
Mg	0.25	0.24	0.29	0.26	0.07	0.31	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Li	0.50	0.66	0.65	0.63	1.82	0.51	0.12	0.11	0.10	0.13	0.15	0.21	0.66	0.66	0.58	0.58	0.57	0.61	0.51	0.72
Y site	5.68	5.64	5.68	5.64	5.53	5.67	5.91	5.85	5.82	5.76	5.79	5.92	6.00	5.94	5.91	5.88	5.88	5.89	5.74	5.51
Ca	-	0.00	-	-	-	-	-	-	0.00	-	-	-	-	0.00	0.01	0.00	0.00	0.00	-	0.00
Na	0.03	0.05	0.05	0.03	0.06	0.01	0.01	0.02	0.02	0.03	0.03	0.02	0.03	0.02	0.03	0.02	0.02	0.04	0.04	0.03
K	1.96	1.91	1.92	1.92	1.88	1.93	1.87	1.86	1.88	1.80	1.89	1.80	2.01	2.02	1.98	1.92	1.97	1.94	1.97	2.02
X site	1.99	1.96	1.97	1.96	1.94	1.94	1.88	1.88	1.91	1.83	1.92	1.82	2.04	2.05	2.01	1.94	2.00	1.99	2.01	2.06
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.94	2.70	2.58	2.68	3.00	2.48	3.13	3.08	3.17	3.25	3.15	3.19	2.62	2.76	2.75	2.78	2.91	2.77	3.06	2.41
F	0.99	1.21	1.33	1.23	1.00	1.44	0.71	0.76	0.67	0.66	0.76	0.71	1.26	1.14	1.10	1.09	0.99	1.06	0.80	1.56
Cl	0.08	0.09	0.09	0.09	0.01	0.08	0.16	0.16	0.16	0.09	0.09	0.09	0.12	0.10	0.15	0.13	0.11	0.18	0.13	0.04
X(F)	0.25	0.30	0.33	0.31	0.25	0.36	0.18	0.19	0.17	0.17	0.19	0.18	0.31	0.29	0.27	0.27	0.25	0.26	0.20	0.39
X(Cl)	0.02	0.02	0.02	0.02	0.00	0.02	0.04	0.04	0.04	0.02	0.02	0.02	0.03	0.03	0.04	0.03	0.03	0.04	0.03	0.01
X(F/Cl)	12.62	14.04	14.19	13.68	143.25	18.89	4.39	4.77	4.27	7.26	8.49	7.48	10.71	11.36	7.28	8.23	9.14	5.98	6.01	42.43
F/FM	0.93	0.93	0.92	0.93	0.93	0.92	1.00	1.00	1.00	1.00	1.00	1.00	0.82	0.81	0.86	0.87	0.87	0.86	0.87	0.96
X(Phl)	0.04	0.04	0.05	0.05	0.01	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.14	0.14	0.11	0.10	0.10	0.11	0.10	0.02
X(Ann)	0.20	0.28	0.32	0.30	0.06	0.18	0.75	0.74	0.76	0.76	0.71	0.80	0.47	0.40	0.44	0.43	0.42	0.53	0.46	0.17
X(Srp)	0.76	0.67	0.63	0.65	0.93	0.77	0.25	0.26	0.24	0.24	0.29	0.20	0.39	0.46	0.46	0.47	0.49	0.36	0.44	0.80

	14-2	14-3	14-4	14-5	14-6	14-7	29-1	29-2	29-3	29-4	29-5	29-6	29-7	29-8	28B-1	28B-2	28B-3	28B-4	28B-5	28B-6
SiO2	37.00	36.54	36.84	37.11	37.37	37.26	35.76	34.99	47.47	35.36	36.07	35.37	35.91	35.78	36.55	37.08	36.88	37.28	36.48	36.55
TiO2	1.75	1.07	1.68	1.62	1.95	1.78	1.83	2.12	0.09	2.33	2.57	1.88	2.68	2.63	2.70	2.18	2.38	1.63	1.75	2.25
Al2O3	17.87	18.49	18.06	17.76	17.48	17.69	17.91	16.98	27.92	15.87	15.91	16.81	15.93	15.87	17.66	17.74	17.74	18.74	19.53	17.50
FeO	24.62	25.69	23.83	28.71	28.26	29.11	29.43	29.60	6.82	28.89	27.22	27.23	25.99	26.69	27.25	27.79	27.89	26.24	26.92	28.53
MnO	0.60	0.49	0.52	0.48	0.54	0.51	0.74	0.74	0.55	0.70	0.51	0.59	0.74	0.76	0.47	0.47	0.50	0.43	0.50	0.49
MgO	0.65	0.64	0.71	0.64	0.60	0.68	0.96	0.86	0.53	0.95	1.01	0.90	0.96	0.94	0.56	0.55	0.56	0.62	0.57	0.60
CaO	-	0.01	-	-	-	-	-	0.02	0.03	0.02	0.02	0.01	0.02	-	0.00	-	-	0.00	0.02	-
Na2O	0.13	0.12	0.10	0.19	0.12	0.15	0.13	0.15	0.13	0.12	0.13	0.12	0.11	0.07	0.14	0.18	0.17	0.17	0.12	0.17
K2O	9.68	9.64	9.75	9.60	9.79	9.63	8.99	9.51	11.07	9.35	9.47	9.51	9.59	9.72	9.29	9.38	9.38	9.80	9.49	9.72
H2O#	2.49	2.64	2.43	2.37	2.48	2.53	2.59	2.82	3.32	2.71	2.74	2.83	2.87	2.83	2.14	1.89	2.14	2.27	2.17	2.07
F	2.71	2.35	2.83	2.84	2.69	2.56	2.20	1.60	2.28	1.89	1.86	1.67	1.68	1.81	3.28	3.82	3.26	3.16	3.37	3.37
Cl	0.15	0.20	0.17	0.35	0.20	0.25	0.52	0.58	0.01	0.47	0.55	0.46	0.40	0.28	0.38	0.43	0.46	0.26	0.22	0.47
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.80	-1.00	-0.90	-0.90	-0.80	-0.80	-0.80	-1.00	-2.00	-1.00	-1.00	-1.00	-2.00
Total	96.47	96.86	95.68	100.41	100.30	101.02	100.01	99.15	99.25	97.77	97.14	96.59	96.09	96.54	98.95	99.80	99.89	99.20	99.67	100.19
Li2O*	1.07	0.94	1.02	1.10	1.17	1.14	0.71	0.49	4.07	0.60	0.80	0.60	0.76	0.72	0.94	1.09	1.03	1.15	0.92	0.94
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.87	5.81	5.88	5.75	5.78	5.74	5.62	5.62	6.26	5.73	5.81	5.75	5.83	5.80	5.73	5.76	5.73	5.77	5.65	5.71
IVAl	2.13	2.19	2.12	2.25	2.22	2.26	2.38	2.38	1.74	2.27	2.19	2.25	2.17	2.20	2.27	2.24	2.27	2.23	2.35	2.29
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAl	1.21	1.27	1.27	1.00	0.96	0.95	0.94	0.83	2.60	0.76	0.83	0.97	0.87	0.84	0.99	1.01	0.98	1.19	1.22	0.93
VITi	0.21	0.13	0.20	0.19	0.23	0.21	0.22	0.26	0.01	0.28	0.31	0.23	0.33	0.32	0.32	0.25	0.28	0.19	0.20	0.26
Fe2+	3.26	3.42	3.18	3.72	3.65	3.75	3.87	3.97	0.75	3.92	3.67	3.70	3.53	3.62	3.57	3.61	3.63	3.40	3.49	3.73
Mn2+	0.08	0.07	0.07	0.06	0.07	0.07	0.10	0.10	0.06	0.10	0.07	0.08	0.10	0.10	0.06	0.06	0.07	0.06	0.07	0.07
Mg	0.15	0.15	0.17	0.15	0.14	0.16	0.22	0.21	0.10	0.23	0.24	0.22	0.23	0.23	0.13	0.13	0.13	0.14	0.13	0.14
Li	0.68	0.60	0.65	0.68	0.73	0.71	0.45	0.32	2.16	0.39	0.52	0.39	0.49	0.47	0.59	0.68	0.65	0.71	0.57	0.59
Y site	5.60	5.63	5.54	5.80	5.78	5.83	5.80	5.68	5.69	5.68	5.63	5.60	5.55	5.58	5.67	5.74	5.73	5.69	5.69	5.72
Ca	-	0.00	-	-	-	-	-	0.00	0.00	0.00	0.00	0.00	0.00	-	0.00	-	-	0.00	0.00	-
Na	0.04	0.04	0.03	0.06	0.04	0.05	0.04	0.05	0.03	0.04	0.04	0.04	0.04	0.02	0.04	0.05	0.05	0.05	0.04	0.05
K	1.96	1.95	1.98	1.90	1.93	1.89	1.80	1.95	1.86	1.93	1.95	1.97	1.99	2.01	1.86	1.86	1.86	1.94	1.88	1.94
X site	2.00	1.99	2.02	1.96	1.97	1.94	1.84	2.00	1.90	1.98	1.99	2.01	2.02	2.03	1.90	1.91	1.91	1.99	1.91	1.99
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.60	2.77	2.53	2.52	2.63	2.69	2.76	3.03	3.05	2.90	2.90	3.01	3.03	3.00	2.27	2.01	2.28	2.38	2.29	2.21
F	1.36	1.18	1.43	1.39	1.32	1.25	1.10	0.81	0.95	0.97	0.94	0.86	0.86	0.93	1.63	1.88	1.60	1.55	1.65	1.66
Cl	0.04	0.05	0.04	0.09	0.05	0.06	0.14	0.16	0.00	0.13	0.15	0.13	0.11	0.08	0.10	0.11	0.12	0.07	0.06	0.12
X(F)	0.34	0.30	0.36	0.35	0.33	0.31	0.27	0.20	0.24	0.24	0.24	0.22	0.22	0.23	0.41	0.47	0.40	0.39	0.41	0.42
X(Cl)	0.01	0.01	0.01	0.02	0.01	0.02	0.03	0.04	0.00	0.03	0.04	0.03	0.03	0.02	0.03	0.03	0.03	0.02	0.01	0.03
X(F/Cl)	33.70	22.15	31.81	15.27	25.14	19.29	7.88	5.18	708.19	7.50	6.25	6.75	7.77	12.04	16.08	16.54	13.33	22.36	28.88	13.37
F/FM	0.96	0.96	0.95	0.96	0.96	0.96	0.95	0.95	0.88	0.94	0.94	0.94	0.94	0.94	0.96	0.97	0.97	0.96	0.96	0.96
X(Phl)	0.03	0.03	0.03	0.03	0.02	0.03	0.04	0.04	0.02	0.04	0.04	0.04	0.04	0.04	0.02	0.02	0.02	0.03	0.02	0.02
X(Ann)	0.28	0.24	0.27	0.29	0.31	0.30	0.24	0.27	0.11	0.35	0.37	0.29	0.36	0.36	0.28	0.29	0.29	0.24	0.19	0.29
X(Srp)	0.69	0.74	0.70	0.68	0.66	0.67	0.72	0.69	0.87	0.61	0.59	0.67	0.60	0.60	0.69	0.69	0.69	0.73	0.79	0.68

	30A-1	30A-2	30A-3	30A-4	37-1	37-2	37-3	37-4	37-5	37-6	37-7	37-8	37-9	1A1-1	1A1-2	1A1-3	1A1-4	1A1-5	1A1-6	1A1-7
SiO2	39.74	40.10	37.57	38.97	39.03	39.75	40.72	39.23	40.67	38.91	43.35	40.56	38.29	52.00	37.67	37.97	38.09	38.08	37.66	37.53
TiO2	0.49	0.42	0.05	0.01	0.89	0.21	0.52	0.67	0.35	0.86	0.36	0.85	1.50	0.01	1.18	1.19	1.28	1.37	1.09	2.91
Al2O3	20.48	21.00	21.04	21.17	18.74	19.17	19.39	18.18	19.37	17.76	20.02	19.10	17.43	25.08	19.79	20.24	19.93	18.75	20.13	17.34
FeO	23.96	22.83	26.46	24.32	24.43	23.47	21.06	24.98	21.50	24.29	17.30	21.21	25.56	6.72	25.70	25.20	25.29	26.81	24.71	27.12
MnO	0.48	0.52	0.50	0.60	0.50	0.51	0.42	0.38	0.42	0.48	0.43	0.44	0.54	0.03	0.47	0.68	0.45	0.51	0.52	0.51
MgO	0.14	0.09	0.08	0.09	0.23	0.26	0.22	0.25	0.25	0.23	0.27	0.25	0.24	1.18	0.65	0.62	0.57	1.01	0.98	0.99
CaO	-	-	-	0.05	-	0.02	-	0.02	-	-	0.01	0.02	0.02	0.34	-	0.01	0.02	0.06	0.01	-
Na2O	0.18	0.20	0.09	0.10	0.20	0.16	0.17	0.15	0.11	0.16	0.11	0.16	0.08	0.02	0.09	0.11	0.09	0.15	0.18	0.11
K2O	9.76	9.88	9.88	10.02	10.03	10.07	10.08	9.68	10.00	9.89	10.24	9.93	9.84	5.40	9.85	9.99	9.87	9.73	9.96	9.82
H2O#	2.50	2.36	2.43	2.40	1.85	1.54	1.70	1.79	1.48	1.87	1.82	1.59	1.95	2.99	2.71	2.70	2.66	2.74	2.58	2.67
F	3.02	3.37	3.00	3.19	4.26	4.98	4.74	4.37	5.18	4.13	4.73	4.95	3.94	3.32	2.35	2.44	2.50	2.28	2.67	2.36
Cl	0.06	0.03	0.05	0.04	0.04	0.05	0.03	0.06	0.05	0.06	0.01	0.02	0.05	0.02	0.17	0.14	0.16	0.16	0.14	0.18
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00
Total	99.51	99.39	99.87	99.59	98.39	98.07	97.05	97.90	97.17	96.89	96.65	97.01	97.77	95.71	99.62	100.24	99.83	100.65	99.47	100.50
Li2O*	1.85	1.96	1.23	1.63	1.65	1.86	2.14	1.71	2.12	1.61	2.89	2.09	1.44	5.37	1.26	1.35	1.38	1.38	1.26	1.22
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.94	5.96	5.74	5.87	5.98	6.05	6.15	6.04	6.15	6.06	6.34	6.14	5.97	6.71	5.76	5.75	5.78	5.78	5.75	5.75
IVAl	2.06	2.04	2.26	2.13	2.02	1.95	1.85	1.96	1.85	1.94	1.66	1.86	2.03	1.29	2.24	2.25	2.22	2.22	2.25	2.25
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAl	1.55	1.64	1.53	1.62	1.36	1.49	1.60	1.33	1.60	1.32	1.80	1.55	1.17	2.53	1.33	1.36	1.35	1.14	1.37	0.89
VITi	0.05	0.05	0.01	0.00	0.10	0.02	0.06	0.08	0.04	0.10	0.04	0.10	0.18	0.00	0.14	0.14	0.15	0.16	0.12	0.34
Fe2+	2.99	2.84	3.38	3.06	3.13	2.99	2.66	3.21	2.72	3.16	2.12	2.68	3.33	0.73	3.29	3.19	3.21	3.40	3.15	3.48
Mn2+	0.06	0.07	0.06	0.08	0.06	0.07	0.05	0.05	0.05	0.06	0.05	0.06	0.07	0.00	0.06	0.09	0.06	0.07	0.07	0.07
Mg	0.03	0.02	0.02	0.02	0.05	0.06	0.05	0.06	0.06	0.05	0.06	0.06	0.06	0.23	0.15	0.14	0.13	0.23	0.22	0.23
Li	1.11	1.17	0.76	0.99	1.02	1.14	1.30	1.06	1.29	1.01	1.70	1.27	0.90	2.79	0.77	0.82	0.84	0.84	0.77	0.75
Y site	5.80	5.78	5.76	5.77	5.73	5.77	5.72	5.79	5.75	5.71	5.77	5.71	5.71	6.28	5.73	5.74	5.74	5.83	5.71	5.74
Ca	-	-	-	0.01	-	0.00	-	0.00	-	0.00	0.00	0.00	0.00	0.05	-	0.00	0.00	0.01	0.00	-
Na	0.05	0.06	0.03	0.03	0.06	0.05	0.05	0.05	0.03	0.05	0.03	0.05	0.02	0.00	0.03	0.03	0.03	0.05	0.05	0.03
K	1.86	1.87	1.93	1.92	1.96	1.96	1.94	1.90	1.93	1.96	1.91	1.92	1.96	0.89	1.92	1.93	1.91	1.88	1.94	1.92
X site	1.91	1.93	1.95	1.96	2.02	2.01	1.99	1.95	1.96	2.01	1.94	1.97	1.98	0.94	1.95	1.96	1.94	1.94	1.99	1.96
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.56	2.41	2.54	2.47	1.93	1.59	1.73	1.86	1.51	1.95	1.81	1.63	2.04	2.64	2.82	2.80	2.76	2.87	2.68	2.81
F	1.43	1.58	1.45	1.52	2.06	2.40	2.26	2.13	2.48	2.03	2.19	2.37	1.94	1.36	1.13	1.17	1.20	1.09	1.29	1.14
Cl	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.02	0.00	0.01	0.01	0.00	0.04	0.04	0.04	0.04	0.04	0.05
X(F)	0.36	0.40	0.36	0.38	0.52	0.60	0.57	0.53	0.62	0.51	0.55	0.59	0.49	0.34	0.28	0.29	0.30	0.27	0.32	0.29
X(Cl)	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	0.00	0.01	0.01	0.01	0.01	0.01	0.01
X(F/Cl)	100.60	185.13	116.51	145.01	184.66	197.61	340.35	140.50	201.54	118.45	882.67	384.65	138.83	326.47	25.46	31.35	28.49	26.92	36.35	24.61
F/Fe	0.99	0.99	0.99	0.99	0.98	0.98	0.98	0.98	0.98	0.97	0.98	0.98	0.98	0.76	0.96	0.96	0.96	0.94	0.93	0.94
X(Phl)	0.01	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.04	0.03	0.02	0.02	0.04	0.04	0.04
X(Ann)	0.23	0.21	0.15	0.18	0.29	0.29	0.30	0.33	0.30	0.34	0.33	0.31	0.35	0.28	0.20	0.19	0.21	0.26	0.19	0.32
X(Srp)	0.77	0.79	0.85	0.82	0.70	0.70	0.69	0.66	0.69	0.65	0.66	0.68	0.64	0.68	0.77	0.78	0.77	0.70	0.78	0.64

	1A1-8	1A1-9	1A1-10	1A1-11	1A2-1	1A2-2	1A2-3	1A2-4	1A2-5	1A2-6	1A2-7	1A2-8	1A2-9	1A2-10	1A2-11	1A2-12	1A2-13	1A2-14	1B2-1a	1B2-2a
SiO2	36.60	35.47	36.65	36.84	37.09	37.85	39.03	38.56	46.47	37.86	37.76	38.67	37.51	38.21	38.23	38.75	38.49	37.80	37.39	37.76
TiO2	2.26	2.34	2.03	2.28	1.24	1.10	1.01	0.94	0.34	0.77	1.61	1.12	0.91	1.25	0.95	0.77	0.97	1.70	2.55	2.06
Al2O3	18.27	17.74	18.67	18.65	19.43	19.22	19.30	19.56	27.51	19.67	18.59	19.78	19.57	20.18	20.09	20.36	20.29	20.03	17.19	18.09
FeO	27.93	29.48	25.96	27.43	25.73	25.54	24.69	25.11	8.56	23.62	25.26	23.50	23.92	23.13	22.18	21.74	22.29	22.64	25.43	24.47
MnO	0.49	0.50	0.55	0.47	0.59	0.57	0.48	0.44	0.16	0.42	0.56	0.50	0.54	0.48	0.47	0.43	0.45	0.46	0.35	0.27
MgO	0.99	0.96	1.00	1.04	1.05	0.93	0.97	0.96	0.40	1.01	1.04	1.04	0.95	0.95	1.03	1.01	0.99	0.97	1.68	1.75
CaO	0.02	0.03	0.01	0.03	0.02	0.01	-	-	-	0.01	-	-	-	0.00	0.01	0.01	-	0.02	0.02	0.04
Na2O	0.07	0.15	0.16	0.08	0.18	0.16	0.22	0.22	0.14	0.14	0.20	0.19	0.27	0.15	0.09	0.10	0.12	0.12	0.08	0.12
K2O	9.89	9.49	9.98	9.76	9.69	10.09	9.91	9.71	11.10	9.93	9.69	9.81	9.64	9.96	10.01	10.06	9.71	9.82	9.93	9.83
H2O#	2.76	2.50	2.55	2.59	2.49	2.67	2.45	2.50	3.24	2.44	2.32	2.27	2.31	2.50	2.34	2.26	2.37	2.57	2.78	2.74
F	2.12	2.50	2.60	2.51	2.78	2.44	3.02	2.87	2.35	2.95	3.18	3.39	3.19	2.89	3.21	3.42	3.17	2.71	2.10	2.26
Cl	0.19	0.30	0.18	0.20	0.15	0.14	0.14	0.15	0.04	0.13	0.14	0.11	0.12	0.14	0.14	0.14	0.15	0.16	0.26	0.21
O=F+Cl	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.90	-1.00
Total	100.66	100.34	99.19	100.77	99.23	99.67	99.92	99.77	99.31	97.70	98.96	98.94	97.56	98.59	97.37	97.58	97.65	97.82	98.82	98.61
Li2O*	0.95	0.63	0.97	1.02	1.09	1.31	1.65	1.51	3.79	1.31	1.28	1.55	1.21	1.41	1.42	1.57	1.49	1.30	1.18	1.29
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.65	5.58	5.70	5.65	5.73	5.79	5.88	5.84	6.20	5.85	5.81	5.86	5.82	5.82	5.87	5.90	5.87	5.80	5.80	5.81
IVAl	2.35	2.42	2.30	2.35	2.27	2.21	2.12	2.16	1.80	2.15	2.19	2.14	2.18	2.18	2.13	2.10	2.13	2.20	2.20	2.19
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.98	0.87	1.12	1.03	1.26	1.26	1.30	1.33	2.53	1.43	1.19	1.39	1.40	1.44	1.51	1.56	1.53	1.43	0.94	1.09
VITi	0.26	0.28	0.24	0.26	0.14	0.13	0.11	0.11	0.03	0.09	0.19	0.13	0.11	0.14	0.11	0.09	0.11	0.20	0.30	0.24
Fe2+	3.61	3.88	3.38	3.52	3.32	3.27	3.11	3.18	0.96	3.05	3.25	2.98	3.10	2.95	2.85	2.77	2.85	2.91	3.30	3.15
Mn2+	0.06	0.07	0.07	0.06	0.08	0.07	0.06	0.06	0.02	0.06	0.07	0.06	0.07	0.06	0.06	0.06	0.06	0.06	0.05	0.04
Mg	0.23	0.22	0.23	0.24	0.24	0.21	0.22	0.22	0.08	0.23	0.24	0.24	0.22	0.22	0.24	0.23	0.23	0.22	0.39	0.40
Li	0.59	0.40	0.61	0.63	0.68	0.81	1.00	0.92	2.03	0.82	0.80	0.94	0.76	0.87	0.88	0.96	0.92	0.80	0.73	0.80
Y site	5.73	5.72	5.64	5.74	5.72	5.74	5.81	5.80	5.65	5.68	5.73	5.74	5.66	5.67	5.64	5.67	5.68	5.61	5.70	5.72
Ca	0.00	0.01	0.00	0.01	0.00	0.00	-	-	-	0.00	-	-	-	0.00	0.00	0.00	-	0.00	0.00	0.01
Na	0.02	0.05	0.05	0.02	0.05	0.05	0.07	0.06	0.04	0.04	0.06	0.06	0.08	0.04	0.03	0.03	0.04	0.04	0.02	0.04
K	1.95	1.91	1.98	1.91	1.91	1.97	1.90	1.87	1.89	1.96	1.90	1.90	1.91	1.93	1.96	1.96	1.89	1.92	1.96	1.93
X site	1.97	1.96	2.03	1.94	1.96	2.02	1.97	1.94	1.93	2.00	1.96	1.95	1.99	1.98	1.99	1.99	1.93	1.96	1.99	1.97
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.91	2.68	2.68	2.73	2.61	2.78	2.53	2.59	3.00	2.52	2.42	2.34	2.40	2.57	2.40	2.32	2.43	2.64	2.90	2.84
F	1.04	1.24	1.28	1.22	1.36	1.18	1.44	1.38	0.99	1.44	1.55	1.63	1.57	1.39	1.56	1.65	1.53	1.32	1.03	1.10
Cl	0.05	0.08	0.05	0.05	0.04	0.04	0.03	0.04	0.01	0.03	0.04	0.03	0.03	0.03	0.04	0.04	0.04	0.04	0.07	0.05
X(F)	0.26	0.31	0.32	0.30	0.34	0.30	0.36	0.34	0.25	0.36	0.39	0.41	0.39	0.35	0.39	0.41	0.38	0.33	0.26	0.28
X(Cl)	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01
X(F/Cl)	20.62	15.69	27.05	23.37	35.27	32.51	41.38	36.24	125.03	42.69	42.64	55.56	48.82	39.93	43.14	46.63	39.73	32.27	15.10	20.12
F/FM	0.94	0.95	0.94	0.94	0.93	0.94	0.93	0.94	0.92	0.93	0.93	0.93	0.93	0.93	0.92	0.92	0.93	0.93	0.89	0.89
X(Phl)	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.01	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.07	0.07
X(Ann)	0.25	0.25	0.23	0.23	0.20	0.23	0.26	0.23	0.10	0.21	0.26	0.22	0.21	0.20	0.20	0.20	0.19	0.32	0.28	0.28
X(Srp)	0.71	0.72	0.73	0.73	0.76	0.73	0.71	0.73	0.88	0.75	0.70	0.74	0.75	0.77	0.76	0.76	0.76	0.77	0.62	0.65

	1B2-3a	1B2-4a	1B2-5a	1B2-6a	1B2-7a	1B2-8a	1B2-1b	1B2-2b	1B2-3b	1B2-4b	1B2-5b	1B2-6b	2A1-1	2A1-2	2A1-3	2A1-4	2A1-5	2A1-6	2A1-7	2A1-8
SiO2	37.38	37.38	37.33	36.89	36.76	36.64	37.31	37.74	36.87	37.19	37.74	36.42	37.13	37.04	35.83	37.18	37.41	37.78	37.37	37.71
TiO2	2.56	2.10	2.04	2.12	2.31	2.63	2.41	2.51	1.91	1.64	2.00	2.26	1.80	1.19	1.91	1.31	1.83	1.28	2.05	1.62
Al2O3	17.45	17.98	17.82	18.28	17.08	17.64	17.75	17.94	18.06	17.68	17.40	17.32	18.40	19.50	18.07	19.63	19.53	20.17	19.49	18.77
FeO	24.99	24.72	24.38	24.69	25.23	25.51	27.05	26.55	27.82	27.08	27.18	28.37	25.31	24.10	26.57	26.08	24.31	23.84	24.51	24.72
MnO	0.35	0.30	0.32	0.41	0.35	0.38	0.37	0.23	0.31	0.38	0.32	0.40	0.54	0.49	0.50	0.56	0.50	0.57	0.50	0.53
MgO	1.70	1.65	1.64	1.63	1.56	1.55	1.76	1.73	1.66	1.62	1.65	1.53	1.13	1.25	1.15	0.76	1.14	1.17	0.87	1.21
CaO	0.00	-	-	0.05	0.00	-	-	0.02	-	0.01	-	-	0.02	-	-	0.02	0.03	0.01	-	0.02
Na2O	0.14	0.06	0.11	0.06	0.19	0.07	0.13	0.09	0.12	0.08	0.10	0.08	0.20	0.11	0.14	0.15	0.06	0.13	0.17	0.13
K2O	9.95	9.89	9.94	9.83	9.90	9.92	9.94	10.01	9.89	9.73	9.92	9.64	9.52	9.55	9.06	9.56	9.66	9.46	9.66	9.35
H2O#	2.86	2.87	2.97	2.92	2.73	2.97	2.76	2.82	2.83	2.79	2.85	2.74	2.55	2.68	2.52	2.62	2.77	2.38	2.77	2.70
F	1.97	1.90	1.77	1.82	2.09	1.67	2.17	2.11	1.96	2.06	2.00	2.06	2.61	2.41	2.58	2.47	2.29	3.13	2.27	2.40
Cl	0.23	0.32	0.19	0.25	0.37	0.28	0.22	0.17	0.24	0.22	0.19	0.29	0.18	0.14	0.18	0.23	0.12	0.12	0.12	0.12
O=F+Cl	-0.90	-0.90	-0.80	-0.80	-1.00	-0.80	-1.00	0.90	-0.90	-0.90	-0.90	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00
Total	98.70	98.30	97.73	98.12	97.61	98.50	100.89	100.98	100.79	99.55	100.46	100.17	98.24	97.41	97.38	99.48	98.65	98.71	98.80	98.24
Li2O*	1.18	1.18	1.16	1.04	1.00	0.96	1.16	1.28	1.03	1.12	1.28	0.90	1.10	1.08	0.73	1.12	1.19	1.29	1.17	1.27
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.79	5.80	5.82	5.75	5.80	5.73	5.70	5.73	5.67	5.76	5.78	5.67	5.78	5.77	5.70	5.72	5.75	5.77	5.74	5.82
IVAl	2.21	2.20	2.18	2.25	2.20	2.27	2.30	2.27	2.33	2.24	2.22	2.33	2.22	2.23	2.30	2.28	2.25	2.23	2.26	2.18
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.97	1.09	1.09	1.11	0.98	0.98	0.90	0.93	0.94	0.99	0.92	0.85	1.16	1.36	1.09	1.28	1.29	1.40	1.27	1.23
VITi	0.30	0.24	0.24	0.25	0.27	0.31	0.28	0.29	0.22	0.19	0.23	0.26	0.21	0.14	0.23	0.15	0.21	0.15	0.24	0.19
Fe2+	3.24	3.21	3.18	3.22	3.33	3.33	3.46	3.37	3.58	3.51	3.48	3.70	3.30	3.14	3.54	3.36	3.12	3.04	3.15	3.19
Mn2+	0.05	0.04	0.04	0.05	0.05	0.05	0.05	0.03	0.04	0.05	0.04	0.05	0.07	0.06	0.07	0.07	0.06	0.07	0.06	0.07
Mg	0.39	0.38	0.38	0.38	0.37	0.36	0.40	0.39	0.38	0.37	0.38	0.35	0.26	0.29	0.27	0.18	0.26	0.27	0.20	0.28
Li	0.73	0.73	0.73	0.65	0.63	0.61	0.71	0.78	0.64	0.70	0.79	0.56	0.69	0.68	0.47	0.69	0.73	0.79	0.73	0.79
Y site	5.69	5.69	5.66	5.65	5.63	5.64	5.79	5.79	5.80	5.81	5.83	5.79	5.69	5.67	5.67	5.73	5.68	5.72	5.65	5.74
Ca	0.00	-	-	0.01	0.00	-	-	0.00	-	0.00	-	-	0.00	-	-	0.00	0.00	0.00	-	0.00
Na	0.04	0.02	0.03	0.02	0.06	0.02	0.04	0.03	0.04	0.02	0.03	0.02	0.06	0.03	0.04	0.04	0.02	0.04	0.05	0.04
K	1.97	1.96	1.98	1.96	1.99	1.98	1.94	1.94	1.94	1.92	1.94	1.91	1.89	1.90	1.84	1.88	1.89	1.84	1.89	1.84
X site	2.01	1.97	2.01	1.98	2.05	2.00	1.98	1.97	1.98	1.95	1.97	1.94	1.95	1.93	1.88	1.92	1.92	1.88	1.94	1.88
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.98	2.98	3.08	3.04	2.86	3.10	2.90	2.95	2.98	2.93	2.98	2.91	2.67	2.77	2.65	2.74	2.86	2.46	2.87	2.80
F	0.96	0.93	0.87	0.89	1.04	0.83	1.05	1.01	0.95	1.01	0.97	1.01	1.29	1.19	1.30	1.20	1.11	1.51	1.10	1.17
Cl	0.06	0.08	0.05	0.07	0.10	0.07	0.06	0.04	0.06	0.06	0.05	0.08	0.05	0.04	0.05	0.06	0.03	0.03	0.03	0.03
X(F)	0.24	0.23	0.22	0.22	0.26	0.21	0.26	0.25	0.24	0.25	0.24	0.25	0.32	0.30	0.32	0.30	0.28	0.38	0.28	0.29
X(Cl)	0.02	0.02	0.01	0.02	0.03	0.02	0.01	0.01	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
X(F/Cl)	15.69	11.17	16.99	13.44	10.40	11.21	18.82	23.25	15.26	17.46	19.60	13.33	26.77	32.19	26.76	20.30	34.95	48.33	36.24	37.29
F/FH	0.89	0.89	0.89	0.89	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.91	0.93	0.92	0.93	0.95	0.92	0.92	0.94	0.92
X(Phl)	0.07	0.07	0.07	0.07	0.07	0.06	0.07	0.07	0.07	0.06	0.06	0.06	0.05	0.05	0.05	0.03	0.05	0.05	0.04	0.05
X(Ann)	0.30	0.27	0.28	0.25	0.31	0.27	0.28	0.29	0.26	0.29	0.32	0.29	0.25	0.20	0.23	0.20	0.20	0.19	0.21	0.25
X(Srp)	0.63	0.66	0.65	0.69	0.63	0.66	0.65	0.65	0.68	0.65	0.62	0.65	0.70	0.75	0.72	0.77	0.75	0.77	0.76	0.70

	2A1-9	2A1-10	2A1-11	2A1-12	2B1-1	2B1-2	2B1-3	2B1-4	2B1-5	2B1-6	2B1-7	2B1-8	2B1-9	2B1-10	2B1-11	2B1-12	2B1-13	2B1-14	2B1-15	2B1-16
SiO2	46.65	47.96	45.90	37.78	37.83	36.46	47.63	37.48	38.16	38.12	37.55	37.10	36.17	37.38	38.19	38.72	36.94	37.92	37.34	37.21
TiO2	0.16	0.02	0.23	1.97	2.00	2.01	0.23	0.66	2.01	2.10	1.89	1.69	2.15	2.44	2.35	1.54	2.27	2.22	2.13	2.56
Al2O3	26.90	26.56	26.69	18.23	19.23	18.11	27.21	20.13	18.48	18.15	18.39	18.32	17.26	18.06	17.73	18.89	17.92	18.24	17.61	18.05
FeO	9.46	7.94	9.90	26.22	23.39	25.28	6.61	23.73	25.00	24.91	26.04	25.45	23.74	26.02	25.01	24.57	25.14	23.70	24.00	25.60
MnO	0.19	0.17	0.19	0.54	0.54	0.47	0.19	0.54	0.51	0.54	0.52	0.52	0.47	0.50	0.51	0.41	0.54	0.51	0.44	0.50
MgO	0.42	0.52	0.47	1.24	1.22	1.26	0.63	1.05	1.32	1.25	1.35	1.39	1.18	1.20	1.26	1.20	1.16	1.35	1.35	1.25
CaO	-	0.01	-	0.01	0.03	0.03	-	-	0.01	-	-	0.03	-	0.01	0.02	0.02	-	-	-	0.03
Na2O	0.14	0.11	0.11	0.23	0.12	0.19	0.20	0.17	0.13	0.12	0.09	0.03	0.13	0.15	0.16	0.16	0.11	0.16	0.09	0.10
K2O	10.73	10.60	10.67	9.32	9.43	8.93	9.95	8.82	9.48	9.55	9.54	9.16	8.69	9.35	9.16	9.30	9.30	9.15	8.85	9.47
H2O#	3.26	3.18	3.27	2.51	2.62	2.72	3.27	2.57	2.57	2.54	2.65	2.65	2.72	2.55	2.54	2.48	2.62	2.61	2.44	2.51
F	2.29	2.56	2.21	2.69	2.62	2.26	2.42	2.72	2.67	2.65	2.46	2.45	2.23	2.61	2.64	2.93	2.47	2.61	2.85	2.69
Cl	0.01	0.00	0.03	0.28	0.13	0.16	0.00	0.11	0.17	0.31	0.15	0.13	0.14	0.22	0.31	0.14	0.16	0.13	0.18	0.23
O=F+Cl	-1.00	-1.00	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00
Total	99.25	98.55	98.74	99.82	98.02	96.90	97.32	96.82	99.35	99.06	99.55	97.86	93.91	99.35	98.70	99.09	97.55	97.46	96.05	99.02
Li2O*	3.84	4.21	3.62	1.29	1.31	0.91	4.12	1.20	1.40	1.39	1.23	1.10	0.83	1.17	1.41	1.56	1.05	1.33	1.16	1.13

Structural formula on the basis of 24 (O,OH,F,Cl)

IVSi	6.24	6.36	6.21	5.78	5.81	5.76	6.34	5.82	5.82	5.84	5.76	5.79	5.87	5.76	5.86	5.87	5.79	5.86	5.89	5.75
IVAl	1.76	1.64	1.79	2.22	2.19	2.24	1.66	2.18	2.18	2.16	2.24	2.21	2.13	2.24	2.14	2.13	2.21	2.14	2.11	2.25
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	2.48	2.52	2.46	1.07	1.29	1.14	2.62	1.51	1.14	1.11	1.09	1.16	1.18	1.04	1.07	1.25	1.10	1.18	1.16	1.04
VITi	0.02	0.00	0.02	0.23	0.23	0.24	0.02	0.08	0.23	0.24	0.22	0.20	0.26	0.28	0.27	0.18	0.27	0.26	0.25	0.30
Fe2+	1.06	0.88	1.12	3.36	3.00	3.34	0.74	3.08	3.19	3.19	3.34	3.32	3.22	3.35	3.21	3.12	3.30	3.06	3.16	3.31
Mn2+	0.02	0.02	0.02	0.07	0.07	0.06	0.02	0.07	0.07	0.07	0.07	0.07	0.07	0.07	0.07	0.05	0.07	0.07	0.06	0.07
Mg	0.08	0.10	0.10	0.28	0.28	0.30	0.12	0.24	0.30	0.29	0.31	0.32	0.28	0.28	0.29	0.27	0.27	0.31	0.32	0.29
Li	2.06	2.25	1.97	0.79	0.81	0.58	2.21	0.75	0.86	0.85	0.76	0.69	0.54	0.73	0.87	0.95	0.66	0.83	0.74	0.70
X site	5.72	5.77	5.69	5.80	5.68	5.66	5.73	5.74	5.78	5.76	5.79	5.75	5.56	5.74	5.77	5.81	5.67	5.71	5.69	5.70
Ca	-	0.00	-	0.00	0.00	0.01	-	-	0.00	-	-	0.01	-	0.00	0.00	0.00	-	-	-	0.00
Na	0.04	0.03	0.03	0.07	0.04	0.06	0.05	0.05	0.04	0.04	0.03	0.01	0.04	0.04	0.05	0.05	0.03	0.05	0.03	0.03
K	1.83	1.79	1.84	1.82	1.85	1.80	1.69	1.75	1.84	1.87	1.82	1.80	1.84	1.79	1.80	1.86	1.80	1.78	1.78	1.87
X site	1.87	1.82	1.87	1.89	1.89	1.86	1.74	1.80	1.88	1.90	1.89	1.84	1.84	1.88	1.84	1.85	1.89	1.85	1.81	1.90
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.03	2.92	3.05	2.63	2.69	2.83	2.98	2.64	2.67	2.64	2.77	2.76	2.81	2.67	2.64	2.56	2.73	2.69	2.53	2.62
F	0.97	1.08	0.94	1.30	1.27	1.13	1.02	1.34	1.29	1.28	1.19	1.21	1.15	1.27	1.28	1.41	1.22	1.28	1.42	1.32
Cl	0.00	0.00	0.01	0.07	0.03	0.04	0.00	0.03	0.04	0.08	0.04	0.03	0.04	0.06	0.08	0.04	0.04	0.03	0.05	0.06
X(F)	0.24	0.27	0.24	0.33	0.32	0.28	0.25	0.33	0.32	0.32	0.30	0.30	0.29	0.32	0.32	0.35	0.31	0.32	0.36	0.33
X(Cl)	0.00	0.00	0.00	0.02	0.01	0.01	0.00	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.02
X(F/Cl)	473.78	2516.3	124.75	18.25	38.83	26.84	902.82	47.85	29.89	16.06	29.82	35.43	29.35	22.16	15.92	39.96	29.73	37.79	29.40	21.48
F/FM	0.93	0.90	0.92	0.92	0.92	0.92	0.86	0.93	0.91	0.92	0.92	0.91	0.92	0.92	0.92	0.92	0.92	0.91	0.91	0.92
X(Phl)	0.01	0.02	0.02	0.05	0.05	0.05	0.02	0.04	0.05	0.05	0.05	0.06	0.05	0.05	0.05	0.05	0.05	0.05	0.06	0.05
X(Ann)	0.12	0.16	0.12	0.28	0.23	0.25	0.13	0.18	0.27	0.29	0.26	0.25	0.29	0.28	0.31	0.27	0.27	0.28	0.30	0.27
X(Srp)	0.86	0.82	0.87	0.67	0.72	0.70	0.85	0.78	0.68	0.66	0.69	0.69	0.66	0.68	0.64	0.69	0.68	0.67	0.65	0.68

Appendix IV.4: Results of EMPA (spot) analyses of mica in altered granites (albitized and greisenized) plutonic rocks. Albitized biotite granite (1C, 15, 17, 22, 22A, 22C1, 22C2, 22C3, 23, 24, 74C, 76A, 76B), host rock of greisens (35A1, 35A2, 36A1, & 36A2), and post-joint greisens (35C, 36B1, 36C, 84).

	1C-1	1C-2	1C-3	1C-4	1C-5	1C-6	1C-7	1C-8	1C-9	1C-10	1C-11	15-1	15-2	15-3	15-4	15-5	15-6	15-7	15-8	15-9
SiO ₂	37.43	37.14	36.92	36.60	36.97	37.49	37.54	37.62	37.53	36.34	36.91	38.27	38.19	38.91	39.24	38.03	38.74	38.56	39.57	38.91
TiO ₂	1.57	1.41	1.90	1.86	1.93	1.71	1.01	1.80	2.03	1.21	1.14	0.91	1.01	0.95	0.75	0.83	0.87	0.63	0.39	0.63
Al ₂ O ₃	18.84	19.58	18.74	18.72	18.37	18.44	19.10	18.38	17.81	19.16	19.22	18.79	18.97	18.99	20.20	18.94	19.88	18.84	20.21	19.79
FeO	26.07	25.46	26.15	26.87	26.92	25.89	24.55	26.39	26.16	27.10	25.81	25.37	26.71	25.49	23.83	25.94	24.17	23.99	22.70	24.70
MnO	0.55	0.55	0.52	0.53	0.59	0.62	0.45	0.50	0.47	0.58	0.52	0.56	0.52	0.42	0.55	0.58	0.56	0.59	0.54	0.57
MgO	0.64	0.58	0.59	0.62	0.62	0.65	0.63	0.60	0.70	0.63	0.63	0.15	0.15	0.23	0.15	0.16	0.17	0.21	0.19	0.17
CaO	-	-	-	0.03	-	-	0.01	0.05	0.03	0.01	0.03	0.02	-	-	-	-	0.01	-	0.01	0.04
Na ₂ O	0.11	0.15	0.12	0.16	0.21	0.18	0.12	0.17	0.23	0.16	0.18	0.25	0.16	0.25	0.23	0.22	0.23	0.24	0.19	0.20
K ₂ O	9.91	10.02	9.83	9.76	9.73	9.82	9.71	9.62	9.53	9.63	9.46	9.51	9.91	9.74	9.55	9.59	9.65	9.37	9.45	9.89
H ₂ O#	2.16	2.12	2.20	2.27	2.19	2.29	2.86	2.19	2.12	2.23	2.40	1.91	1.97	1.75	1.84	1.72	1.88	1.66	1.78	1.99
F	3.39	3.50	3.30	3.13	3.26	3.12	2.00	3.32	3.46	3.14	2.86	3.94	3.81	4.32	4.23	4.32	4.10	4.45	4.35	3.85
Cl	0.13	0.12	0.12	0.11	0.19	0.11	0.10	0.15	0.11	0.20	0.16	0.08	0.11	0.14	0.10	0.11	0.09	0.13	0.07	0.11
O=F+Cl	-1.00	-2.00	-1.00	-1.00	-1.00	-1.00	-0.90	-1.00	-1.00	-1.00	-1.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00
Total	99.35	99.13	98.96	99.30	99.57	98.98	97.23	96.37	98.67	99.03	98.06	98.09	99.89	99.34	98.85	98.60	98.59	96.77	97.61	99.20
Li ₂ O*	1.19	1.11	1.04	0.95	1.06	1.21	1.22	1.24	1.22	0.88	1.04	1.43	1.41	1.61	1.71	1.36	1.57	1.52	1.80	1.62
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.79	5.75	5.75	5.71	5.75	5.81	5.86	5.81	5.84	5.70	5.77	5.93	5.86	5.94	5.94	5.89	5.91	6.00	6.01	5.92
IVAl	2.21	2.25	2.25	2.29	2.25	2.19	2.14	2.19	2.16	2.30	2.23	2.07	2.14	2.06	2.06	2.11	2.09	2.00	1.99	2.08
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	1.22	1.32	1.19	1.15	1.11	1.18	1.37	1.16	1.10	1.23	1.31	1.36	1.29	1.35	1.54	1.35	1.49	1.46	1.63	1.47
VITi	0.18	0.16	0.22	0.22	0.23	0.20	0.12	0.21	0.24	0.14	0.13	0.11	0.12	0.11	0.09	0.10	0.10	0.07	0.04	0.07
Fe ²⁺	3.37	3.29	3.41	3.51	3.50	3.36	3.20	3.41	3.40	3.55	3.37	3.29	3.43	3.25	3.02	3.36	3.09	3.12	2.89	3.14
Mn ²⁺	0.07	0.07	0.07	0.07	0.08	0.08	0.06	0.07	0.06	0.08	0.07	0.07	0.07	0.05	0.07	0.08	0.07	0.08	0.07	0.07
Mg	0.15	0.13	0.14	0.14	0.14	0.15	0.15	0.14	0.16	0.15	0.15	0.03	0.03	0.05	0.03	0.04	0.04	0.05	0.04	0.04
Li	0.74	0.69	0.65	0.60	0.66	0.75	0.77	0.77	0.76	0.55	0.66	0.89	0.87	0.99	1.04	0.85	0.96	0.95	1.10	0.99
Y site	5.73	5.67	5.67	5.68	5.72	5.72	5.67	5.75	5.72	5.71	5.69	5.75	5.80	5.81	5.79	5.77	5.75	5.73	5.78	5.78
Ca	-	-	-	0.00	-	-	0.00	0.01	0.00	0.00	0.00	0.00	-	-	-	-	0.00	-	0.00	0.01
Na	0.03	0.05	0.04	0.05	0.06	0.05	0.04	0.05	0.07	0.05	0.05	0.07	0.05	0.07	0.07	0.07	0.07	0.07	0.06	0.06
K	1.95	1.98	1.95	1.94	1.93	1.94	1.93	1.90	1.89	1.93	1.89	1.88	1.94	1.90	1.84	1.90	1.88	1.86	1.83	1.92
A site	1.99	2.02	1.99	1.99	1.99	2.00	1.97	1.96	1.96	1.98	1.95	1.96	1.99	1.97	1.91	1.96	1.95	1.93	1.89	1.98
O	20.08	20.07	20.06	20.06	20.08	20.07	20.01	20.08	20.07	20.05	20.04	20.07	20.10	20.10	20.10	20.08	20.09	20.06	20.08	20.10
OH	2.23	2.18	2.28	2.36	2.27	2.37	2.98	2.26	2.20	2.33	2.51	1.98	2.02	1.78	1.85	1.78	1.91	1.72	1.81	2.02
F	1.66	1.71	1.62	1.55	1.60	1.53	0.99	1.62	1.70	1.56	1.41	1.93	1.85	2.09	2.02	2.11	1.98	2.19	2.09	1.85
Cl	0.03	0.03	0.03	0.03	0.05	0.03	0.03	0.04	0.03	0.05	0.04	0.02	0.03	0.04	0.02	0.03	0.02	0.03	0.02	0.03
X(F)	0.42	0.44	0.41	0.39	0.41	0.39	0.25	0.41	0.43	0.40	0.36	0.49	0.47	0.54	0.52	0.54	0.51	0.56	0.53	0.47
X(Cl)	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01
X(F/Cl)	48.62	52.69	53.02	54.15	52.87	54.98	37.32	40.21	57.07	29.63	34.41	87.62	65.30	58.04	82.17	76.00	85.92	65.88	109.77	65.23
F/FM	0.96	0.96	0.96	0.96	0.96	0.96	0.96	0.96	0.95	0.96	0.96	0.99	0.99	0.98	0.99	0.99	0.99	0.98	0.99	0.99
X(Phl)	0.03	0.02	0.02	0.03	0.03	0.03	0.03	0.02	0.03	0.03	0.03	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
X(Ann)	0.24	0.20	0.24	0.23	0.25	0.26	0.23	0.27	0.30	0.20	0.21	0.27	0.26	0.28	0.23	0.26	0.23	0.28	0.23	0.24
X(Srp)	0.73	0.78	0.74	0.75	0.72	0.71	0.74	0.70	0.67	0.77	0.76	0.72	0.73	0.71	0.77	0.74	0.76	0.72	0.76	0.76

	17-1	17-2	17-3	17-4	17-5	22-1	22-2	22-3	22-4	22-5	22-6	22-7	22-8	22A-1	22A-2	22A-3	22A-4	22A-5	22A-6	22A-7
SiO2	47.80	51.55	51.35	53.04	50.35	37.63	36.92	36.77	36.44	37.56	37.61	36.98	38.88	24.16	37.37	22.91	36.85	37.27	37.98	37.31
TiO2	0.32	0.10	0.15	0.08	0.09	1.28	1.05	1.89	1.97	1.15	1.15	1.58	1.33	0.05	1.43	0.09	1.23	1.14	1.07	1.18
Al2O3	26.48	26.30	28.35	23.58	30.33	19.32	20.03	18.06	18.20	19.87	19.60	18.34	18.59	18.94	19.79	19.31	19.51	19.19	20.08	19.45
FeO	6.94	5.01	4.49	5.75	4.23	25.24	24.89	27.46	27.38	25.15	25.15	24.97	25.69	44.08	24.16	44.67	24.34	25.35	24.85	24.98
MnO	0.08	0.01	0.06	-	0.01	0.68	0.53	0.67	0.71	0.64	0.65	0.42	0.47	0.18	0.39	0.56	0.62	0.58	0.42	0.60
MgO	0.40	0.52	0.55	0.77	0.25	0.38	0.32	0.32	0.32	0.33	0.35	0.44	0.38	0.15	0.35	0.30	0.31	0.33	0.33	0.31
CaO	0.05	0.06	0.10	0.04	0.04	-	0.01	-	0.01	-	-	-	-	0.01	-	0.01	-	-	-	-
Na2O	0.03	0.05	0.03	0.03	0.03	0.20	0.09	0.15	0.15	0.21	0.09	0.16	0.17	-	0.09	0.00	0.11	0.21	0.17	0.20
K2O	10.24	10.28	9.90	10.53	10.04	9.77	9.90	9.90	10.03	9.81	9.99	9.45	9.82	0.03	9.57	0.05	9.30	9.51	9.66	9.43
H2O#	3.39	3.00	3.57	2.33	3.86	2.01	2.12	2.35	2.14	2.07	2.09	2.38	2.38	3.52	2.22	3.49	2.14	2.16	2.31	2.27
F	1.97	2.96	1.87	4.30	1.29	3.76	3.50	2.86	3.31	3.64	3.57	2.90	3.03	0.10	3.32	0.10	3.44	3.39	3.18	3.19
Cl	0.03	0.01	0.01	0.00	0.01	0.08	0.09	0.25	0.21	0.10	0.12	0.12	0.08	0.01	0.10	0.01	0.10	0.10	0.10	0.10
O=F+Cl	-0.80	-1.00	-0.80	-2.00	-0.50	-2.00	-1.00	-1.00	-1.00	-2.00	-2.00	-1.00	-1.00	-0.04	-1.00	-0.04	-1.00	-1.00	-1.00	-1.00
Total	96.91	98.60	99.64	98.64	99.99	98.75	97.96	99.43	98.97	98.83	96.50	99.53	91.18	97.36	91.47	96.46	97.77	98.80	97.66	
Li2O*	4.17	5.24	5.19	5.67	4.90	1.25	1.04	1.00	0.91	1.23	1.24	1.06	1.61	-	1.17	-	1.02	1.14	1.35	1.16
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	6.42	6.63	6.51	6.82	6.37	5.82	5.76	5.75	5.72	5.79	5.81	5.85	5.92	4.61	5.82	4.43	5.81	5.82	5.82	5.82
IVAl	1.58	1.37	1.49	1.18	1.63	2.18	2.24	2.25	2.28	2.21	2.19	2.15	2.08	3.39	2.18	3.57	2.19	2.18	2.18	2.18
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	2.60	2.62	2.74	2.40	2.90	1.34	1.45	1.09	1.09	1.40	1.38	1.28	1.26	0.87	1.45	0.82	1.44	1.36	1.45	1.40
VITi	0.03	0.01	0.01	0.01	0.01	0.15	0.12	0.22	0.23	0.13	0.13	0.19	0.15	0.01	0.17	0.01	0.15	0.13	0.12	0.14
Fe2+	0.78	0.54	0.48	0.62	0.45	3.26	3.25	3.59	3.59	3.24	3.25	3.31	3.27	7.03	3.14	7.22	3.21	3.31	3.19	3.26
Mn2+	0.01	0.00	0.01	-	0.00	0.09	0.07	0.09	0.09	0.08	0.09	0.06	0.06	0.03	0.05	0.09	0.08	0.08	0.05	0.08
Mg	0.08	0.10	0.10	0.15	0.05	0.09	0.08	0.07	0.07	0.08	0.08	0.10	0.09	0.04	0.08	0.09	0.07	0.08	0.08	0.07
Li	2.25	2.71	2.64	2.93	2.49	0.78	0.66	0.63	0.57	0.76	0.77	0.68	0.98	-	0.73	-	0.65	0.72	0.83	0.73
Y site	5.75	5.98	5.98	6.10	5.90	5.70	5.62	5.70	5.65	5.69	5.70	5.61	5.82	5.97	5.62	5.92	5.60	5.68	5.72	5.67
Ca	0.01	0.01	0.01	0.00	0.01	-	0.00	-	0.00	-	-	-	-	0.00	-	0.00	-	-	-	-
Na	0.01	0.01	0.01	0.01	0.01	0.06	0.03	0.04	0.05	0.06	0.03	0.05	0.05	-	0.03	0.00	0.03	0.06	0.05	0.06
K	1.75	1.69	1.60	1.73	1.62	1.93	1.97	1.98	2.01	1.93	1.97	1.91	1.91	0.01	1.90	0.01	1.87	1.90	1.89	1.88
X site	1.77	1.71	1.62	1.74	1.63	1.99	2.00	2.02	2.06	1.99	2.00	1.96	1.96	0.01	1.93	0.02	1.90	1.96	1.94	1.94
O	20.12	20.22	20.23	20.25	20.22	20.07	20.04	20.07	20.07	20.07	20.07	20.00	20.10	19.46	20.03	19.45	20.01	20.04	20.07	20.04
OH	3.03	2.57	3.01	2.00	3.26	2.07	2.20	2.45	2.24	2.12	2.16	2.51	2.42	4.48	2.31	4.49	2.25	2.25	2.36	2.36
F	0.84	1.20	0.75	1.75	0.51	1.84	1.73	1.42	1.64	1.77	1.74	1.45	1.46	0.06	1.63	0.06	1.71	1.67	1.54	1.57
Cl	0.01	0.00	0.00	0.00	0.00	0.02	0.03	0.07	0.06	0.03	0.03	0.03	0.02	0.00	0.03	0.00	0.03	0.03	0.02	0.03
X(F)	0.22	0.32	0.20	0.47	0.14	0.47	0.44	0.36	0.42	0.45	0.44	0.36	0.37	0.01	0.41	0.01	0.43	0.42	0.39	0.40
X(Cl)	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.01
X(F/Cl)	108.18	501.64	268.72	4016.8	159.99	87.59	68.81	21.46	29.39	65.31	56.39	46.33	72.44	22.86	64.50	23.09	62.25	61.40	61.76	57.26
F/FM	0.91	0.84	0.82	0.81	0.91	0.97	0.98	0.98	0.98	0.98	0.98	0.97	0.97	0.99	0.98	0.99	0.98	0.98	0.98	0.98
X(Phl)	0.01	0.02	0.02	0.02	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
X(Ann)	0.16	0.23	0.16	0.37	0.09	0.23	0.18	0.27	0.25	0.20	0.21	0.26	0.30	-1E-01	0.20	-1E-01	0.20	0.22	0.20	0.21
X(Srp)	0.83	0.75	0.82	0.61	0.90	0.76	0.81	0.72	0.73	0.79	0.77	0.72	0.69	1.09	0.79	1.12	0.79	0.76	0.79	0.77

	22A-8	22A-9	22A-10	22C1-1	22C1-2	22C1-3	22C1-4	22C1-5	22C1-6	22C1-7	22C1-8	22C2-1	22C2-2	22C2-3	22C2-4	22C2-5	22C2-6	22C2-7	22C3-1	22C3-2
SiO2	38.29	37.36	38.01	46.55	39.20	38.00	38.92	37.70	38.68	38.85	37.59	40.24	22.84	40.63	38.28	39.13	38.22	45.78	37.28	37.44
TiO2	1.15	1.50	1.39	0.16	1.06	0.82	1.11	1.21	1.17	1.09	0.73	0.96	0.07	0.86	0.99	1.23	1.53	0.15	1.26	1.13
Al2O3	20.31	18.96	19.15	27.44	19.78	20.12	19.78	19.24	19.86	20.20	19.26	20.24	19.13	20.24	19.25	19.16	18.97	26.78	18.82	19.24
FeO	23.19	25.23	24.86	6.86	22.58	24.18	22.05	22.54	24.65	23.77	25.29	21.32	45.79	21.21	25.23	23.04	25.99	10.08	26.15	26.27
MnO	0.54	0.53	0.66	0.07	0.43	0.38	0.33	0.32	0.41	0.37	0.46	0.32	0.37	0.36	0.34	0.37	0.37	0.09	0.54	0.57
MgO	0.31	0.32	0.34	0.15	0.45	0.42	0.40	0.45	0.44	0.40	0.43	0.50	0.22	0.47	0.45	0.47	0.45	0.10	0.48	0.44
CaO	-	0.01	0.02	-	0.03	0.02	-	-	0.01	-	-	0.02	0.03	0.02	0.02	-	-	-	0.01	-
Na2O	0.21	0.23	0.20	0.15	0.08	0.23	0.17	0.12	0.18	0.11	0.25	0.16	0.04	0.17	0.13	0.10	0.12	0.09	0.31	0.18
K2O	9.52	9.61	9.61	10.55	9.75	9.66	9.46	9.28	9.60	9.41	9.08	9.54	0.02	9.82	9.38	9.52	9.42	10.57	9.23	9.34
H2O#	2.09	2.34	2.16	3.19	2.16	2.19	2.07	2.42	2.23	2.38	2.14	1.87	3.46	1.79	2.27	2.38	2.55	3.19	1.95	1.95
F	3.68	3.03	3.45	2.34	3.52	3.43	3.72	2.93	3.41	3.12	3.46	4.24	0.12	4.41	3.24	3.07	2.69	2.23	3.83	3.84
Cl	0.08	0.09	0.10	0.01	0.16	0.07	0.06	0.09	0.06	0.08	0.08	0.09	0.01	0.10	0.09	0.13	0.07	0.01	0.09	0.10
O=F+Cl	-2.00	-1.00	-1.00	-1.00	-2.00	-1.00	-2.00	-1.00	-1.00	-1.00	-1.00	-2.00	-0.05	-2.00	-1.00	-1.00	-1.00	-0.09	-2.00	-2.00
Total	97.80	97.93	98.48	96.47	97.68	98.06	96.49	95.05	99.26	98.44	97.30	97.70	92.05	98.21	98.27	97.27	99.24	98.13	98.33	98.85
Li2O*	1.44	1.17	1.36	3.81	1.70	1.35	1.62	1.27	1.55	1.60	1.24	2.00	-	2.11	1.43	1.68	1.42	3.59	1.15	1.19
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.87	5.83	5.86	6.32	5.97	5.85	5.98	5.94	5.87	5.90	5.87	6.04	4.41	6.06	5.89	5.99	5.85	6.24	5.81	5.80
IVAl	2.13	2.17	2.14	1.68	2.03	2.15	2.02	2.06	2.13	2.10	2.13	1.96	3.59	1.94	2.11	2.01	2.15	1.76	2.19	2.20
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIA1	1.54	1.31	1.34	2.71	1.52	1.50	1.57	1.51	1.42	1.51	1.41	1.62	0.76	1.62	1.38	1.45	1.27	2.55	1.27	1.31
VITi	0.13	0.18	0.16	0.02	0.12	0.10	0.13	0.14	0.13	0.12	0.09	0.11	0.01	0.10	0.11	0.14	0.18	0.02	0.15	0.13
Fe2+	2.97	3.29	3.21	0.78	2.88	3.11	2.84	2.97	3.13	3.02	3.30	2.68	7.39	2.65	3.25	2.95	3.33	1.15	3.41	3.40
Mn2+	0.07	0.07	0.09	0.01	0.05	0.05	0.04	0.04	0.05	0.05	0.06	0.04	0.06	0.05	0.04	0.05	0.05	0.01	0.07	0.07
Mg	0.07	0.07	0.08	0.03	0.10	0.10	0.09	0.11	0.10	0.09	0.10	0.11	0.06	0.11	0.10	0.11	0.10	0.02	0.11	0.10
Li	0.89	0.73	0.84	2.08	1.04	0.84	1.00	0.80	0.95	0.98	0.78	1.21	-	1.27	0.89	1.03	0.87	1.97	0.72	0.74
Y site	5.67	5.65	5.71	5.62	5.72	5.69	5.67	5.57	5.78	5.77	5.74	5.77	5.96	5.78	5.78	5.73	5.80	5.71	5.73	5.77
Ca	-	0.00	0.00	-	0.01	0.00	-	-	0.00	-	-	0.00	0.01	0.00	0.00	-	-	-	0.00	-
Na	0.06	0.07	0.06	0.04	0.02	0.07	0.05	0.04	0.05	0.03	0.07	0.05	0.01	0.05	0.04	0.03	0.04	0.02	0.09	0.05
K	1.86	1.91	1.89	1.83	1.89	1.90	1.86	1.87	1.82	1.81	1.83	0.00	1.87	1.84	1.86	1.84	1.84	1.84	1.84	1.85
X site	1.93	1.98	1.95	1.87	1.92	1.97	1.91	1.90	1.91	1.85	1.88	1.88	0.02	1.92	1.88	1.89	1.88	1.86	1.93	1.90
O	20.06	20.04	20.07	20.10	20.07	20.06	20.05	19.98	20.09	20.08	20.04	20.09	19.47	20.11	20.07	20.05	20.08	20.14	20.06	20.07
OH	2.13	2.44	2.22	2.89	2.20	2.25	2.13	2.54	2.25	2.41	2.23	1.87	4.45	1.78	2.33	2.43	2.60	2.90	2.03	2.02
F	1.79	1.49	1.68	1.01	1.69	1.67	1.81	1.46	1.64	1.50	1.71	2.02	0.07	2.08	1.58	1.49	1.30	0.96	1.89	1.88
Cl	0.02	0.03	0.03	0.00	0.04	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.00	0.03	0.02	0.03	0.02	0.00	0.02	0.03
X(F)	0.45	0.38	0.43	0.26	0.43	0.42	0.46	0.36	0.42	0.38	0.43	0.52	0.02	0.54	0.40	0.38	0.33	0.25	0.48	0.48
X(Cl)	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.01
X(F/Cl)	88.09	59.48	62.52	397.48	41.79	92.68	106.91	63.64	99.40	76.66	77.86	84.27	32.26	83.20	70.39	45.08	72.70	593.69	75.97	68.92
F/FM	0.98	0.98	0.98	0.96	0.97	0.97	0.97	0.97	0.97	0.97	0.97	0.96	0.99	0.96	0.97	0.96	0.97	0.98	0.97	0.97
X(Phl)	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.01	0.02	0.02	0.02	0.02	0.00	0.02	0.02
X(Ann)	0.20	0.24	0.24	0.11	0.24	0.20	0.24	0.23	0.23	0.21	0.23	0.25	-1E-01	0.25	0.25	0.27	0.26	0.11	0.24	0.22
X(Srp)	0.79	0.75	0.74	0.89	0.74	0.79	0.75	0.75	0.76	0.77	0.75	0.74	1.12	0.73	0.74	0.71	0.72	0.88	0.74	0.76

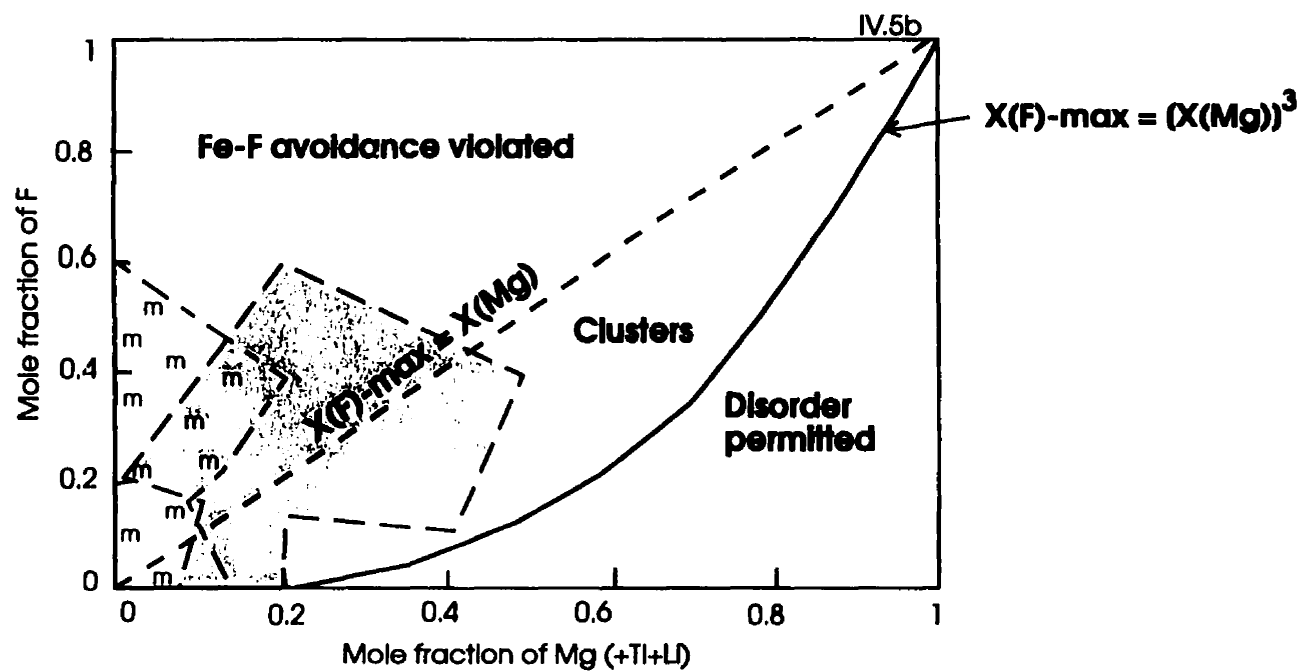
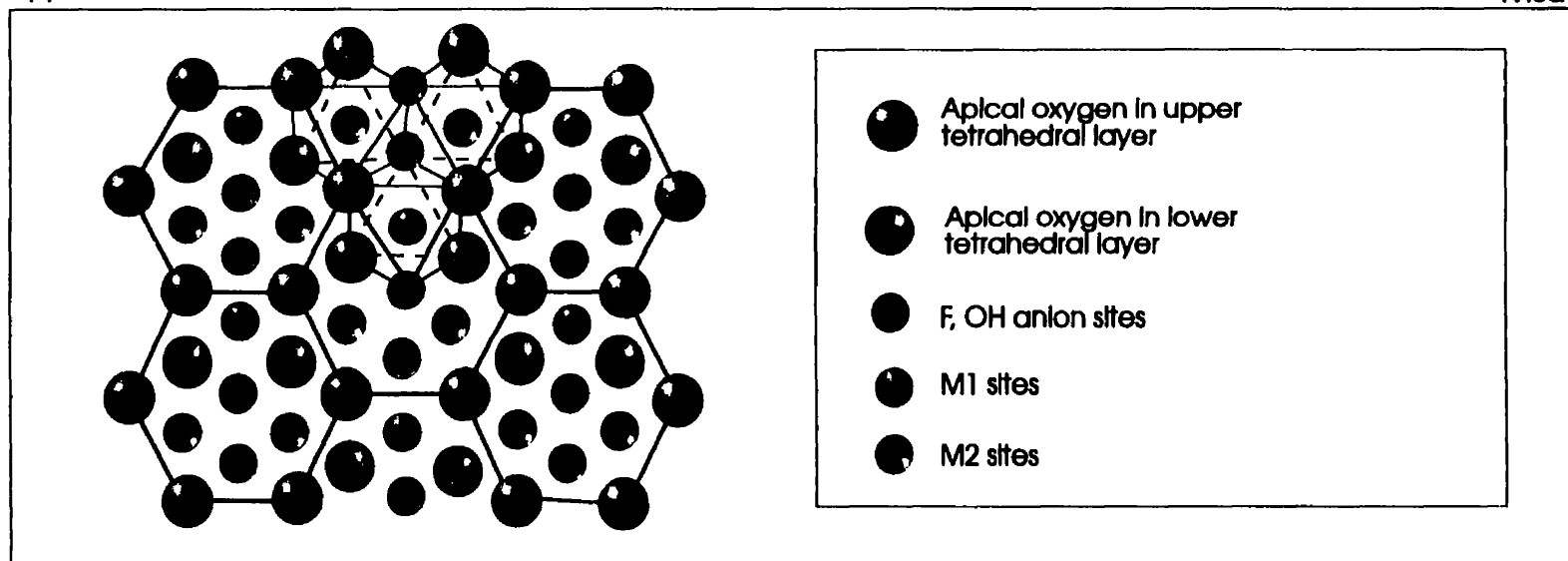
	22C3-3	22C3-4	22C3-5	22C3-6	23-1	23-2	23-3	23-4	23-5	23-6	23-7	23-8	24-1	24-2	24-3	24-4	24-5	74C.1	74C.2	74C.3
SiO2	37.85	37.61	36.74	36.69	37.58	38.81	38.80	37.79	38.20	38.01	38.43	38.40	38.84	37.75	39.28	39.28	38.49	37.01	38.22	37.88
TiO2	1.02	1.59	2.24	2.05	0.85	1.09	1.24	1.16	1.11	1.29	1.36	1.69	2.79	2.39	3.02	2.45	2.76	1.76	1.11	1.03
Al2O3	19.36	17.95	17.49	18.11	19.53	20.33	20.02	19.61	20.13	20.66	20.30	19.68	15.01	13.28	14.45	14.81	14.70	18.62	19.90	19.35
FeO	25.04	27.77	28.12	28.34	24.48	22.04	22.55	24.09	24.78	22.77	23.37	24.33	23.88	26.24	23.22	23.17	23.35	25.20	23.66	24.92
MnO	0.44	0.55	0.58	0.61	0.49	0.50	0.51	0.47	0.47	0.47	0.51	0.58	0.48	0.46	0.31	0.43	0.51	0.81	0.72	0.66
MgO	0.48	0.44	0.39	0.45	0.30	0.23	0.23	0.26	0.26	0.25	0.24	0.25	3.36	3.48	3.63	3.54	3.27	3.29	0.35	0.35
CaO	-	0.03	-	-	-	0.02	0.04	0.00	-	-	-	0.01	0.04	0.02	0.01	0.02	0.03	-	0.02	-
Na2O	0.28	0.14	0.26	0.22	0.13	0.21	0.21	0.17	0.28	0.25	0.17	0.15	0.05	-	0.04	0.06	0.07	0.14	0.23	0.24
K2O	9.36	9.67	9.38	9.42	9.38	9.39	9.39	9.57	9.87	9.92	9.77	9.81	9.73	9.50	9.86	10.12	9.88	9.69	9.53	9.62
H2O#	1.87	2.01	2.15	2.01	2.23	1.87	2.16	1.93	1.53	1.51	1.68	1.95	2.32	2.42	1.89	2.48	2.61	2.69	2.07	1.92
F	4.07	3.69	3.24	3.62	3.27	4.14	3.55	3.93	4.83	4.85	4.54	3.95	3.07	2.72	4.01	2.77	2.46	3.45	3.68	3.95
Cl	0.06	0.11	0.29	0.18	0.12	0.08	0.08	0.09	0.07	0.10	0.05	0.09	0.17	0.17	0.12	0.15	0.16	0.11	0.08	0.05
O=F+Cl	-2.00	-2.00	-1.00	-2.00	-1.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-2.00	-1.00	-1.00	-2.00	-1.00	-1.00	-1.00	-2.00	-2.00
Total	98.12	99.96	99.46	100.15	96.95	96.96	97.26	97.39	99.47	98.03	98.52	99.21	98.42	97.26	98.13	98.08	97.22	1238.2	98.01	98.30
Li2O*	1.31	1.24	0.99	0.98	1.23	1.59	1.58	1.29	1.41	1.36	1.48	1.47	1.59	1.28	1.72	1.72	1.49	1.07	1.42	1.32
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.86	5.82	5.76	5.71	5.87	5.94	5.94	5.87	5.83	5.83	5.86	5.85	5.97	5.99	6.03	6.03	5.99	0.67	5.87	5.86
IVAl	2.14	2.18	2.24	2.29	2.13	2.06	2.06	2.13	2.17	2.17	2.14	2.15	2.03	2.01	1.97	1.97	2.01	0.40	2.13	2.14
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.02	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	1.09	8.00	8.00
VAl	1.39	1.09	0.99	1.04	1.47	1.62	1.55	1.46	1.44	1.57	1.51	1.38	0.69	0.47	0.65	0.72	0.69	-	1.48	1.39
VITi	0.12	0.18	0.26	0.24	0.10	0.13	0.14	0.14	0.13	0.15	0.16	0.19	0.32	0.29	0.35	0.28	0.32	-	0.13	0.12
Fe2+	3.24	3.59	3.69	3.69	3.20	2.82	2.89	3.13	3.16	2.92	2.98	3.10	3.07	3.48	2.98	2.98	3.04	0.38	3.04	3.22
Mn2+	0.06	0.07	0.08	0.08	0.06	0.06	0.07	0.06	0.06	0.06	0.07	0.07	0.06	0.06	0.04	0.06	0.07	12.39	0.09	0.09
Mg	0.11	0.10	0.09	0.10	0.07	0.05	0.05	0.06	0.06	0.06	0.06	0.06	0.77	0.82	0.83	0.81	0.76	8.84	0.08	0.08
Li	0.82	0.77	0.62	0.61	0.77	0.98	0.97	0.81	0.87	0.84	0.91	0.90	0.99	0.82	1.06	1.06	0.94	0.08	0.88	0.82
Y site	5.73	5.81	5.73	5.76	5.68	5.66	5.67	5.65	5.72	5.60	5.68	5.71	5.91	5.94	5.91	5.91	5.82	21.69	5.70	5.72
Ca	-	0.00	-	-	-	0.00	0.01	0.00	-	-	-	0.00	0.01	0.00	0.00	0.00	0.00	-	0.00	-
Na	0.09	0.04	0.08	0.07	0.04	0.06	0.06	0.05	0.08	0.08	0.05	0.04	0.01	-	0.01	0.02	0.02	0.00	0.07	0.07
K	1.85	1.91	1.88	1.87	1.87	1.84	1.83	1.90	1.92	1.94	1.90	1.91	1.91	1.92	1.93	1.98	1.96	0.22	1.87	1.90
X site	1.93	1.95	1.96	1.94	1.91	1.90	1.90	1.95	2.00	2.02	1.95	1.95	1.93	1.93	1.95	2.00	1.99	0.23	1.94	1.97
O	20.07	20.09	20.07	20.08	20.03	20.06	20.06	20.05	20.09	20.07	20.08	20.09	20.08	20.03	20.09	20.07	20.03	23.48	20.06	20.07
OH	1.93	2.07	2.24	2.08	2.33	1.91	2.20	2.00	1.56	1.55	1.71	1.98	2.38	2.56	1.93	2.54	2.71	0.32	2.12	1.98
F	1.99	1.81	1.61	1.78	1.61	2.01	1.72	1.93	2.33	2.36	2.19	1.90	1.49	1.36	1.95	1.35	1.21	0.20	1.79	1.93
Cl	0.02	0.03	0.08	0.05	0.03	0.02	0.02	0.02	0.02	0.03	0.01	0.02	0.05	0.05	0.03	0.04	0.04	0.00	0.02	0.01
X(F)	0.51	0.46	0.41	0.46	0.41	0.51	0.44	0.49	0.60	0.60	0.56	0.49	0.38	0.34	0.50	0.34	0.31	0.38	0.46	0.49
X(Cl)	0.00	0.01	0.02	0.01	0.01	0.00	0.01	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00
X(F/Cl)	124.54	65.03	20.73	37.55	53.01	103.08	86.13	82.42	121.80	88.80	166.12	81.08	32.94	29.16	60.85	34.91	28.03	59.66	83.86	134.19
F/FM	0.97	0.97	0.98	0.97	0.98	0.98	0.98	0.98	0.98	0.98	0.98	0.98	0.80	0.81	0.78	0.79	0.80	0.04	0.97	0.98
X(Phl)	0.02	0.02	0.02	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.13	0.14	0.14	0.14	0.13	0.41	0.01	0.01
X(Ann)	0.23	0.30	0.30	0.26	0.22	0.21	0.22	0.22	0.20	0.18	0.20	0.23	0.47	0.57	0.52	0.49	0.48	0.15	0.21	0.23
X(Srp)	0.75	0.69	0.68	0.72	0.77	0.78	0.77	0.77	0.79	0.81	0.79	0.76	0.40	0.29	0.34	0.37	0.39	0.44	0.77	0.75

	74C.4	76A.1	76A.2	76A.3	76A.4	76A.5	76A.6	76A.7	76A.8	76B.1	76B.2	76B.3	76B.4	76B.5	76B.6	35A1-1	35A1-2	35A1-3	35A1-4	35A2-1
SiO2	37.45	36.28	35.81	35.60	36.21	35.44	36.59	35.65	35.77	37.93	37.59	36.57	35.16	38.15	35.97	37.18	37.09	38.19	38.64	37.06
TiO2	1.14	2.51	2.43	2.39	1.95	22.68	2.45	2.60	2.19	1.35	1.57	1.98	1.74	1.91	1.46	1.17	1.36	1.28	1.08	1.52
Al2O3	19.01	18.93	18.26	18.31	18.85	18.12	18.55	17.76	17.97	19.88	19.93	18.96	18.43	19.63	18.99	18.15	17.77	18.67	19.14	16.95
FeO	25.27	27.30	28.19	27.49	27.59	28.46	26.99	29.14	28.38	26.38	25.24	27.60	27.66	26.43	27.17	26.37	27.77	25.54	24.40	28.49
MnO	0.57	0.54	0.44	0.49	0.47	0.57	0.59	0.56	0.54	0.52	0.71	0.57	0.57	0.60	0.76	0.52	0.45	0.54	0.43	0.49
MgO	0.29	0.79	0.79	0.79	0.85	0.74	0.86	0.85	0.86	0.27	0.32	0.35	0.31	0.31	0.25	0.41	0.38	0.47	0.46	0.36
CaO	-	0.01	-	-	-	-	-	-	-	-	0.04	0.01	-	0.00	-	-	-	0.04	0.01	0.01
Na2O	0.19	0.09	0.16	0.09	0.09	0.08	0.11	0.14	0.16	0.13	0.11	0.12	0.15	0.12	0.12	0.24	0.20	0.22	0.23	0.16
K2O	9.38	10.01	9.79	9.69	9.90	10.11	10.01	9.74	9.82	9.93	9.73	9.78	9.74	10.06	10.13	9.52	9.76	9.45	9.44	9.68
H2O#	2.02	2.73	2.73	2.64	2.58	2.83	2.56	2.64	2.70	2.65	2.60	2.73	2.43	2.58	2.43	2.41	2.26	2.28	2.33	2.39
F	3.69	2.15	2.09	2.29	2.46	2.35	2.54	2.22	2.14	2.45	2.56	2.20	2.68	2.58	2.74	2.82	2.99	3.20	3.15	2.66
Cl	0.08	0.19	0.20	0.16	0.14	0.13	0.11	0.23	0.15	0.10	0.08	0.08	0.11	0.15	0.12	0.13	0.30	0.11	0.08	0.37
O=F+Cl	-2.00	-0.90	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-0.90	-1.00	-1.00	-0.90	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00
Total	97.52	100.56	99.96	98.94	100.01	120.48	100.26	100.53	99.74	100.54	99.40	100.00	97.83	101.41	98.97	97.71	99.00	98.62	98.05	98.95
Li2O*	1.20	0.86	0.73	0.66	0.84	0.62	0.95	0.68	0.71	1.33	1.24	0.94	0.54	1.40	0.77	1.12	1.09	1.41	1.54	1.09
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.86	5.61	5.61	5.62	5.64	4.64	5.66	5.59	5.63	5.77	5.76	5.68	5.65	5.75	5.68	5.85	5.82	5.89	5.93	5.85
IVAl	2.14	2.39	2.39	2.38	2.36	2.79	2.34	2.41	2.37	2.23	2.24	2.32	2.35	2.25	2.32	2.15	2.18	2.11	2.07	2.15
IVTi	-	-	-	-	-	0.57	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	1.36	1.06	0.99	1.03	1.10	-	1.05	0.88	0.97	1.33	1.36	1.15	1.13	1.24	1.21	1.22	1.11	1.28	1.39	1.00
VITi	0.13	0.29	0.29	0.28	0.23	1.66	0.28	0.31	0.26	0.15	0.18	0.23	0.21	0.22	0.17	0.14	0.16	0.15	0.12	0.18
Fe2+	3.30	3.53	3.70	3.63	3.59	3.11	3.49	3.82	3.74	3.35	3.23	3.58	3.71	3.33	3.58	3.47	3.64	3.29	3.13	3.76
Mn2+	0.08	0.07	0.06	0.07	0.06	0.06	0.08	0.07	0.07	0.07	0.09	0.08	0.08	0.08	0.10	0.07	0.06	0.07	0.06	0.07
Mg	0.07	0.18	0.18	0.19	0.20	0.14	0.20	0.20	0.20	0.06	0.07	0.08	0.07	0.07	0.06	0.10	0.09	0.11	0.11	0.09
Li	0.75	0.54	0.46	0.42	0.53	0.33	0.59	0.43	0.45	0.81	0.76	0.59	0.35	0.85	0.49	0.71	0.69	0.87	0.95	0.69
Y site	5.69	5.68	5.67	5.62	5.70	5.31	5.69	5.71	5.69	5.78	5.70	5.70	5.56	5.79	5.62	5.70	5.75	5.77	5.76	5.77
Ca	-	0.00	-	-	-	-	-	-	-	-	0.01	0.00	-	0.00	-	-	-	0.01	0.00	0.00
Na	0.06	0.03	0.05	0.03	0.03	0.02	0.03	0.04	0.05	0.04	0.03	0.04	0.05	0.04	0.04	0.07	0.06	0.07	0.07	0.05
K	1.87	1.98	1.96	1.95	1.97	1.69	1.98	1.95	1.97	1.93	1.90	1.94	1.99	1.94	2.04	1.91	1.95	1.86	1.85	1.95
X site	1.93	2.00	2.00	1.98	2.00	1.71	2.01	1.99	2.02	1.96	1.94	1.97	2.04	1.97	2.00	1.99	2.01	1.93	1.92	2.00
O	20.04	20.08	20.06	20.03	20.07	20.53	20.08	20.08	20.06	20.11	20.08	20.07	20.00	20.13	20.04	20.03	20.06	20.07	20.07	20.06
OH	2.11	2.81	2.85	2.78	2.68	2.47	2.64	2.76	2.84	2.69	2.66	2.82	2.61	2.60	2.56	2.53	2.37	2.34	2.38	2.51
F	1.82	1.05	1.04	1.15	1.21	0.97	1.25	1.10	1.07	1.18	1.24	1.08	1.36	1.23	1.37	1.40	1.49	1.56	1.53	1.32
Cl	0.02	0.05	0.05	0.04	0.04	0.03	0.03	0.06	0.04	0.03	0.02	0.02	0.03	0.04	0.03	0.03	0.08	0.03	0.02	0.10
X(F)	0.46	0.27	0.26	0.29	0.31	0.28	0.32	0.28	0.27	0.30	0.32	0.28	0.34	0.32	0.35	0.35	0.38	0.40	0.39	0.34
X(Cl)	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.03
X(F/Cl)	82.87	21.71	19.69	27.43	33.96	33.25	42.39	17.89	27.02	43.89	58.30	51.41	47.15	31.30	41.59	41.45	18.75	56.34	76.44	13.25
F/FM	0.98	0.95	0.95	0.95	0.95	0.96	0.95	0.95	0.95	0.98	0.98	0.98	0.98	0.98	0.97	0.98	0.97	0.97	0.98	0.98
X(Phl)	0.01	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.02	0.01
X(Ann)	0.24	0.21	0.23	0.22	0.21	0.23	0.23	0.25	0.24	0.21	0.20	0.22	0.21	0.23	0.20	0.27	0.29	0.27	0.26	0.34
X(Srp)	0.75	0.76	0.74	0.75	0.76	0.75	0.73	0.72	0.72	0.78	0.79	0.77	0.78	0.76	0.79	0.71	0.69	0.71	0.72	0.64

	35A2-2	35A2-3	35A2-4	35A2-5	36A1-1	36A1-2	36A1-3	36A1-4	36A1-5	36A1-6	36A1-7	36A2-1	36A2-2	36A2-3	36A2-4	36A2-5	36A2-6	35C-1	35C-2	35C-3
SiO2	38.24	37.90	37.88	38.32	36.55	36.22	35.86	35.33	34.44	36.15	23.34	23.69	23.67	36.25	36.37	28.05	36.72	37.42	36.47	36.47
TiO2	1.31	1.57	1.82	1.05	2.20	1.86	2.24	2.37	1.74	1.67	0.07	0.02	0.17	2.07	1.95	0.01	1.98	-	-	0.01
Al2O3	18.53	18.60	18.18	18.82	16.41	17.24	16.87	17.26	17.28	17.42	18.84	19.33	18.84	16.98	17.44	19.03	17.26	19.17	19.68	18.69
FeO	25.27	25.48	26.48	25.27	29.12	29.58	29.14	29.68	31.10	28.86	45.71	45.50	46.12	29.28	28.93	41.62	28.68	27.48	28.29	27.95
MnO	0.50	0.44	0.52	0.51	0.45	0.49	0.49	0.55	0.53	0.56	0.33	0.25	0.40	0.43	0.47	0.29	0.55	0.48	0.46	0.39
MgO	0.41	0.40	0.42	0.41	1.31	0.69	0.73	0.64	0.67	0.69	0.09	0.33	0.08	0.69	0.72	0.24	0.67	0.04	0.03	0.03
CaO	-	-	-	-	0.01	0.03	-	0.01	0.01	-	0.03	-	0.01	-	-	0.03	0.01	-	0.00	-
Na2O	0.20	0.20	0.23	0.19	0.21	0.22	0.29	0.17	0.19	0.17	0.03	0.03	0.03	0.25	0.23	1.71	0.22	0.16	0.08	0.08
K2O	9.73	9.32	9.25	9.66	10.05	9.66	9.54	9.68	8.77	9.99	-	0.01	0.01	9.57	9.80	0.02	9.68	9.69	9.71	9.45
H2O#	2.27	2.35	2.49	2.39	2.39	2.31	2.55	2.69	2.61	2.56	3.49	3.42	3.46	2.47	2.53	3.55	2.49	2.33	2.42	2.60
F	3.21	3.03	2.75	3.00	2.56	2.67	2.19	2.02	2.15	2.28	0.08	0.23	0.13	2.31	2.28	0.20	2.38	2.98	2.79	2.37
Cl	0.07	0.10	0.09	0.06	0.52	0.61	0.55	0.31	0.26	0.37	-	0.02	0.00	0.65	0.56	0.01	0.53	0.14	0.09	0.13
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.90	-1.00	-1.00	-0.03	-0.10	-0.06	-1.00	-1.00	-0.09	-1.00	-1.00	-1.00	-1.00
Total	98.39	98.11	98.93	98.41	100.58	100.32	99.40	99.79	98.78	99.68	91.97	92.73	92.87	99.83	100.18	94.68	100.06	98.60	98.84	97.13
Li2O*	1.42	1.32	1.32	1.45	0.94	0.84	0.74	0.59	0.33	0.82	-	-	-	0.85	0.89	-	0.99	1.19	0.91	0.91
Structural formula on the basis of 24 (O,OH,F,Cl)																				
IVSi	5.91	5.87	5.85	5.91	5.73	5.70	5.70	5.62	5.57	5.71	4.49	4.49	4.50	5.72	5.71	4.95	5.75	5.85	5.74	5.83
IVAl	2.09	2.13	2.15	2.09	2.27	2.30	2.30	2.38	2.43	2.29	3.51	3.51	3.50	2.28	2.29	3.05	2.25	2.15	2.26	2.17
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	1.28	1.27	1.16	1.33	0.76	0.90	0.86	0.85	0.87	0.96	0.76	0.81	0.72	0.88	0.93	0.91	0.93	1.38	1.39	1.35
VITi	0.15	0.18	0.21	0.12	0.26	0.22	0.27	0.28	0.21	0.20	0.01	0.00	0.02	0.25	0.23	0.00	0.23	-	-	0.00
Fe2+	3.27	3.30	3.42	3.26	3.82	3.89	3.87	3.95	4.21	3.81	7.35	7.21	7.34	3.87	3.80	6.14	3.75	3.59	3.72	3.74
Mn2+	0.07	0.06	0.07	0.07	0.06	0.07	0.07	0.07	0.07	0.08	0.05	0.04	0.06	0.06	0.06	0.04	0.07	0.06	0.06	0.05
Mg	0.10	0.09	0.10	0.09	0.31	0.16	0.17	0.15	0.16	0.16	0.03	0.09	0.02	0.16	0.17	0.06	0.16	0.01	0.01	0.01
Li	0.88	0.83	0.82	0.90	0.59	0.53	0.47	0.38	0.22	0.52	-	-	-	0.54	0.56	-	0.62	0.75	0.58	0.59
Y site	5.74	5.73	5.78	5.76	5.80	5.78	5.71	5.68	5.74	5.73	5.99	6.06	6.06	5.75	5.75	6.09	5.77	5.80	5.76	5.73
Ca	-	-	-	-	0.00	0.01	-	0.00	0.00	-	0.01	-	0.00	-	-	0.01	0.00	-	0.00	-
Na	0.06	0.06	0.07	0.06	0.06	0.07	0.09	0.05	0.06	0.05	0.01	0.01	0.01	0.08	0.07	0.58	0.07	0.05	0.03	0.02
K	1.92	1.84	1.82	1.90	2.01	1.94	1.93	1.96	1.81	2.01	-	0.00	0.00	1.93	1.96	0.00	1.93	1.93	1.95	1.93
X site	1.98	1.90	1.89	1.96	2.07	2.01	2.02	2.02	1.87	2.07	0.02	0.01	0.02	2.01	2.03	0.59	2.00	1.98	1.98	1.95
O	20.07	20.06	20.07	20.07	20.09	20.08	20.05	20.05	20.01	20.06	19.48	19.53	19.53	20.07	20.08	19.70	20.08	20.06	20.05	20.00
OH	2.34	2.43	2.56	2.46	2.50	2.43	2.70	2.85	2.82	2.70	4.47	4.33	4.39	2.60	2.64	4.18	2.60	2.43	2.54	2.77
F	1.57	1.48	1.34	1.46	1.27	1.33	1.10	1.02	1.10	1.14	0.05	0.14	0.08	1.16	1.13	0.11	1.18	1.47	1.39	1.20
Cl	0.02	0.03	0.02	0.02	0.14	0.16	0.15	0.08	0.07	0.10	-	0.01	0.00	0.17	0.15	0.00	0.14	0.04	0.02	0.03
X(F)	0.40	0.38	0.34	0.37	0.32	0.34	0.28	0.26	0.28	0.29	0.01	0.03	0.02	0.29	0.29	0.03	0.30	0.37	0.35	0.30
X(Cl)	0.00	0.01	0.01	0.00	0.04	0.04	0.04	0.02	0.02	0.03	-	0.00	0.00	0.04	0.04	0.00	0.04	0.01	0.01	0.01
X(F/Cl)	81.05	54.88	56.94	96.39	9.09	8.15	7.52	12.13	15.36	11.42	0.01	25.03	81.49	6.68	7.62	28.57	8.45	39.72	56.49	35.10
F/FM	0.97	0.97	0.97	0.97	0.93	0.96	0.96	0.96	0.96	0.96	1.00	0.99	1.00	0.96	0.96	0.99	0.96	1.00	1.00	1.00
X(Phl)	0.02	0.02	0.02	0.02	0.05	0.03	0.03	0.03	0.03	0.03	0.00	0.02	0.00	0.03	0.03	0.01	0.03	0.00	0.00	0.00
X(Ann)	0.28	0.27	0.29	0.27	0.35	0.30	0.31	0.27	0.24	0.28	-1E-01	-1E-01	-1E-01	0.31	0.29	0.00	0.31	0.23	0.18	0.23
X(Srp)	0.70	0.71	0.69	0.72	0.60	0.68	0.66	0.70	0.73	0.69	1.11	1.10	1.10	0.66	0.68	0.99	0.66	0.77	0.82	0.77

	35C-4	35C-5	35C-6	36B1-1	36B1-2	36B1-3	36B1-4	36B1-5	36C1-1	36C1-2	36C1-3	36C2-1	36C2-2	36C2-3	36C2-4	84.1	84.2	84.3	84.4	84.5	84.6
SiO2	37.49	35.97	35.90	49.36	48.61	49.13	48.40	49.47	47.22	47.97	49.35	48.15	48.07	48.29	48.27	33.02	32.15	33.37	33.19	34.00	34.38
TiO2	0.01	0.01	0.03	0.11	0.06	0.01	0.04	0.03	-	0.17	0.10	0.10	0.10	0.05	0.04	2.08	2.20	0.03	0.14	0.08	0.41
Al2O3	18.96	19.61	19.27	30.14	30.51	31.46	31.00	31.79	29.65	29.07	30.63	28.68	28.66	29.70	29.84	19.73	19.39	21.80	21.58	21.64	20.83
FeO	28.01	28.46	27.67	3.19	2.94	2.78	3.12	2.78	4.93	4.81	3.86	4.84	4.92	4.81	4.71	28.11	31.19	28.85	27.12	27.89	28.73
MnO	0.41	0.38	0.36	0.05	0.04	-	0.03	-	0.30	0.38	0.33	0.50	0.47	0.35	0.38	0.27	0.30	0.33	0.28	0.27	0.31
MgO	0.03	0.05	0.04	0.56	0.45	0.47	0.39	0.48	0.20	0.16	0.09	0.14	0.95	0.07	0.08	-	-	0.01	-	0.02	-
CaO	-	-	-	0.05	0.01	-	0.02	0.04	0.29	0.00	0.02	0.02	0.03	0.03	-	0.01	0.00	0.01	0.01	0.01	0.01
Na2O	0.12	0.17	0.09	0.11	0.06	0.08	0.04	0.07	0.12	0.05	0.09	0.04	0.04	0.05	0.07	0.09	0.09	0.14	0.08	0.07	0.08
K2O	9.60	9.38	9.32	10.06	9.90	10.20	10.18	8.75	9.20	10.34	9.95	10.16	10.40	10.50	9.80	9.60	9.73	9.80	9.83	10.01	9.98
H2O#	2.52	2.49	2.51	3.76	3.75	3.77	3.70	3.87	3.99	3.64	3.90	3.46	3.64	3.61	3.57	2.73	2.78	2.62	2.57	2.50	2.50
F	2.61	2.57	2.53	1.48	1.51	1.50	1.60	1.38	0.91	1.59	1.19	1.95	1.58	1.66	1.77	1.90	1.66	2.16	2.27	2.44	2.45
Cl	0.08	0.16	0.16	0.02	-	0.00	-	0.00	0.05	0.00	0.01	-	0.01	0.03	0.01	0.33	0.39	0.31	0.30	0.31	0.29
O=F,Cl	-1.00	-1.00	-1.00	-0.60	-0.60	-0.60	-0.70	-0.60	-0.40	-0.70	-0.50	-0.80	-0.70	-0.70	-0.70	-0.90	-0.80	-1.00	-1.00	-1.00	-1.00
Total	98.72	98.13	96.78	98.25	97.20	98.77	97.84	98.07	96.46	97.52	99.01	97.24	98.21	98.44	97.80	96.99	99.10	98.46	96.34	98.15	98.86
Li2O*	1.21	0.77	0.75	4.61	4.40	4.55	4.34	4.64	4.00	4.22	4.61	4.27	4.24	4.31	4.30	-	-	0.02	-	0.21	0.31
Structural formula on the basis of 24 (O,OH,F,Cl)																					
IVSi	5.86	5.72	5.76	6.35	6.32	6.29	6.28	6.30	6.27	6.33	6.33	6.36	6.31	6.31	6.32	5.42	5.28	5.39	5.44	5.47	5.50
IVAl	2.14	2.28	2.24	1.65	1.68	1.71	1.72	1.70	1.73	1.67	1.67	1.64	1.69	1.69	1.68	2.58	2.72	2.61	2.56	2.53	2.50
IVTi	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAl	1.35	1.39	1.41	2.93	3.00	3.03	3.02	3.08	2.92	2.85	2.95	2.83	2.75	2.88	2.92	1.24	1.03	1.54	1.61	1.56	1.43
VITi	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	-	0.02	0.01	0.01	0.01	0.01	0.00	0.26	0.27	0.00	0.02	0.01	0.05
Fe2+	3.66	3.78	3.72	0.34	0.32	0.30	0.34	0.30	0.55	0.53	0.41	0.54	0.54	0.53	0.52	3.86	4.28	3.90	3.72	3.75	3.84
Mn2+	0.05	0.05	0.05	0.01	0.00	-	0.00	-	0.03	0.04	0.04	0.06	0.05	0.04	0.04	0.04	0.04	0.05	0.04	0.04	0.04
Mg	0.01	0.01	0.01	0.11	0.09	0.09	0.08	0.09	0.04	0.03	0.02	0.03	0.19	0.01	0.02	-	-	0.00	-	0.01	-
Li	0.76	0.49	0.48	2.39	2.30	2.34	2.26	2.38	2.14	2.24	2.38	2.27	2.24	2.26	2.26	-	-	0.02	-	0.13	0.20
Y	5.83	5.73	5.68	5.78	5.72	5.76	5.70	5.85	5.67	5.70	5.80	5.73	5.78	5.73	5.76	5.34	5.42	5.50	5.37	5.50	5.56
Ca	-	-	-	0.01	0.00	-	0.00	0.01	0.04	0.00	0.00	0.00	0.00	0.00	-	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.04	0.05	0.03	0.03	0.02	0.02	0.01	0.02	0.03	0.01	0.02	0.01	0.01	0.01	0.02	0.03	0.03	0.05	0.02	0.02	0.03
K	1.91	1.90	1.91	1.65	1.64	1.67	1.69	1.42	1.56	1.74	1.63	1.71	1.74	1.75	1.64	2.01	2.04	2.02	2.06	2.05	2.04
X	1.95	1.96	1.94	1.68	1.66	1.69	1.70	1.44	1.63	1.75	1.65	1.72	1.76	1.77	1.65	2.04	2.07	2.07	2.08	2.08	2.06
O	20.06	20.02	19.99	20.16	20.12	20.18	20.14	20.16	20.07	20.13	20.18	20.13	20.15	20.16	20.15	19.94	19.98	19.99	19.93	19.99	20.02
OH	2.63	2.64	2.68	3.23	3.26	3.22	3.20	3.29	3.53	3.20	3.33	3.05	3.19	3.14	3.12	2.99	3.05	2.82	2.81	2.68	2.66
F	1.29	1.29	1.29	0.60	0.62	0.61	0.66	0.55	0.38	0.66	0.48	0.81	0.66	0.69	0.73	0.99	0.86	1.10	1.18	1.24	1.24
Cl	0.02	0.04	0.04	0.00	-	0.00	-	0.00	0.01	0.00	0.00	-	0.00	0.01	0.00	0.09	0.11	0.09	0.08	0.08	0.08
X(F)	0.33	0.33	0.32	0.16	0.16	0.16	0.17	0.14	0.10	0.17	0.13	0.21	0.17	0.18	0.19	0.24	0.21	0.28	0.29	0.31	0.31
X(Cl)	0.01	0.01	0.01	0.00	-	0.00	-	0.00	0.00	0.00	0.00	-	0.00	0.00	0.00	0.02	0.03	0.02	0.02	0.02	0.02
F/Cl	65.02	30.98	29.92	162.46	0.16	1399.6	0.17	856.54	32.04	1487.3	246.95	0.21	210.47	115.01	235.66	10.87	8.02	12.94	14.29	14.74	15.85
F/FH	1.00	1.00	1.00	0.76	0.79	0.77	0.82	0.76	0.93	0.95	0.96	0.95	0.74	0.98	0.97	1.00	1.00	1.00	1.00	1.00	1.00
X(Phl)	0.00	0.00	0.00	0.02	0.02	0.02	0.01	0.02	0.01	0.01	0.00	0.00	0.03	0.00	0.00	-	-	0.00	-	0.00	-
X(Ann)	0.24	0.17	0.19	0.08	0.06	0.04	0.04	0.04	0.06	0.09	0.07	0.10	0.10	0.07	0.07	0.10	0.09	0.03	0.03	0.05	0.09
X(Srp)	0.75	0.82	0.81	0.90	0.93	0.94	0.94	0.94	0.94	0.91	0.93	0.90	0.87	0.92	0.93	0.90	0.91	0.97	0.97	0.95	0.91

Appendix IV.5: a) A schematic illustration of the structure of mica (modified from Mason 1991), and illustrating a cluster of 3 octahedral M-sites (M_3). b) Plot of mole fraction of F (X_F) vs. $X_{Mg(+Ti+Li)}$ for mica from Kwadonkaya complex. The area shaded m is the compositional field in which X_F is plotted against X_{Mg} , and the area shaded grey is the same field when the mole fraction of Ti and Li are added to that of Mg.



Appendix V.1: Unit-cell parameters determined for alkali feldspars in representative samples. Symbols are explained in detail on Table 5.2 (chapter 5).

APPENDIX V-1: UNIT-CELL PARAMETERS OF FELDSPARS IN REPRESENTATIVE SAMPLES OF THE KWANDONKAYA YOUNGER GRANITE COMPLEX

	<u>a</u>	<u>b</u>	<u>c</u>	<u>a</u>	<u>β</u>	<u>γ</u>	<u>v</u>	<u>#</u>	<u>a</u> *	<u>b</u> *	<u>c</u> *	<u>a</u> *	<u>β</u> *	<u>γ</u> *
1B	8.5799	12.9740	7.2192	90.457	116.018	88.426	721.89	37	0.129741	0.077107	0.154142	90.260	63.984	91.529
core	0.0017	0.0030	0.0013	0.023	0.015	0.025	0.20		0.000030	0.000018	0.000026	0.023	0.015	0.025
	8.1402	12.7897	7.1598	94.211	116.599	87.794	664.72	60	0.137390	0.078401	0.156510	86.394	63.484	90.360
	0.0006	0.0010	0.0008	0.011	0.009	0.008	0.09		0.000013	0.000006	0.000020	0.011	0.009	0.008
1B	8.5838	12.9661	7.2187	90.53	115.927	88.064	722.14	41	0.129609	0.077170	0.154035	90.371	64.075	91.904
orange	0.0016	0.0017	0.0013	0.018	0.014	0.021	0.16		0.000026	0.000010	0.000024	0.018	0.014	0.021
core	8.1378	12.7896	7.1557	94.245	116.605	87.735	664.06	58	0.137439	0.078406	0.156610	86.386	63.482	90.409
	0.0009	0.0013	0.0009	0.013	0.010	0.009	0.09		0.000018	0.000008	0.000018	0.012	0.010	0.009
1B	8.5923	12.9696	7.2146	90.532	115.936	87.919	722.51	36	0.129502	0.077156	0.154136	90.420	64.066	92.055
matrix	0.0014	0.0018	0.0011	0.023	0.014	0.021	0.16		0.000023	0.000011	0.000021	0.023	0.014	0.021
	8.1377	12.7840	7.1558	94.277	116.591	87.731	663.82	47	0.137424	0.078443	0.156595	86.352	63.496	90.399
	0.0011	0.0022	0.0015	0.016	0.013	0.012	0.14		0.000023	0.000013	0.000035	0.016	0.013	0.012
22	8.5877	12.9684	7.2176	90.573	115.929	87.845	722.37	46	0.129568	0.077167	0.154062	90.411	64.074	92.118
sub bt	0.0014	0.0014	0.0010	0.014	0.010	0.014	0.13		0.000022	0.000008	0.000020	0.015	0.010	0.015
granite	8.1421	12.7912	7.1592	94.258	116.646	87.743	664.57	48	0.137415	0.078397	0.156593	86.368	63.440	90.391
	0.0011	0.0013	0.0011	0.014	0.014	0.011	0.12		0.000023	0.000008	0.000027	0.014	0.014	0.010
22	8.5766	12.9674	7.2122	90	116.15	90	720.04	22	0.129886	0.077117	0.154459	90	63.855	90
altered	0.0023	0.0036	0.0016		0.020		0.26		0.000040	0.000021	0.000035		0.020	
part	8.1385	12.7890	7.1595	94.237	116.625	87.729	664.33	58	0.137451	0.078409	0.156551	86.398	63.462	90.418
*	0.0011	0.0016	0.0009	0.013	0.011	0.011	0.11		0.000024	0.000010	0.000020	0.013	0.012	0.011
25D	8.5883	12.9618	7.2165	90.565	115.933	87.832	721.91	52	0.129564	0.077207	0.154091	90.426	64.070	92.135
	0.0013	0.0018	0.0011	0.015	0.011	0.014	0.13		0.000023	0.000011	0.000023	0.014	0.011	0.013
	8.1382	12.7889	7.1602	94.236	116.665	87.746	664.14	47	0.137504	0.078409	0.156593	86.391	63.420	90.397
	0.0011	0.0017	0.0010	0.017	0.012	0.012	0.11		0.000022	0.000010	0.000022	0.016	0.012	0.011
26B	8.5932	12.9679	7.2167	90.516	115.983	88.153	722.52	50	0.129520	0.077155	0.154151	90.326	64.020	91.803
main	0.0017	0.0016	0.0010	0.016	0.014	0.015	0.15		0.000026	0.000010	0.000022	0.016	0.014	0.015
	8.1387	12.7907	7.1594	94.238	116.602	87.717	664.56	50	0.137420	0.078398	0.156522	86.402	63.486	90.433
	0.0008	0.0013	0.0007	0.010	0.008	0.010	0.08		0.000014	0.000008	0.000016	0.010	0.008	0.010
30A	8.5810	12.9656	7.2173	90.569	115.868	87.878	722.01	26	0.129599	0.077182	0.153988	90.396	64.135	92.082
f.g.	0.0027	0.0032	0.0016	0.031	0.030	0.033	0.28		0.000055	0.000020	0.000040	0.029	0.030	0.032
bt gr	8.1404	12.7876	7.1595	94.236	116.593	87.715	664.59	52	0.137381	0.078417	0.156508	86.406	63.494	90.437
	0.0012	0.0016	0.0010	0.014	0.013	0.010	0.12		0.000025	0.000010	0.000024	0.014	0.013	0.011
35A	8.5841	12.9633	7.2147	90.404	115.934	88.350	721.65	24	0.129598	0.077175	0.154136	90.354	64.062	91.639
rim,	0.0029	0.0027	0.0015	0.025	0.030	0.044	0.28		0.000045	0.000015	0.000042	0.026	0.030	0.045
pheno	8.1394	12.7915	7.1579	94.217	116.584	87.791	664.65	40	0.137385	0.078391	0.156532	86.388	63.499	90.362
	0.0009	0.0014	0.0010	0.014	0.011	0.013	0.11		0.000020	0.000009	0.000021	0.014	0.011	0.014
35A	8.5824	12.9568	7.2163	90.503	116.032	88.188	720.68	31	0.129735	0.077220	0.154223	90.325	63.971	91.771
core,	0.0013	0.0014	0.0007	0.012	0.013	0.013	0.14		0.000025	0.000009	0.000021	0.014	0.013	0.015
pheno	8.1357	12.7828	7.1599	94.231	116.598	87.723	663.97	50	0.137467	0.078446	0.156505	86.407	63.490	90.430
	0.0009	0.0015	0.0007	0.012	0.009	0.011	0.09		0.000016	0.000009	0.000017	0.011	0.009	0.010
35A	8.5826	12.9658	7.2125	90.442	115.975	88.272	721.18	32	0.129665	0.077163	0.154231	90.350	64.027	91.707
matrix	0.0018	0.0023	0.0016	0.024	0.028	0.024	0.28		0.000039	0.000014	0.000054	0.029	0.028	0.029
	8.1419	12.7923	7.1599	94.243	116.647	87.765	664.69	43	0.137419	0.078388	0.156578	86.373	63.438	90.374
	0.0011	0.0013	0.0007	0.013	0.011	0.010	0.10		0.000019	0.000008	0.000016	0.013	0.011	0.010

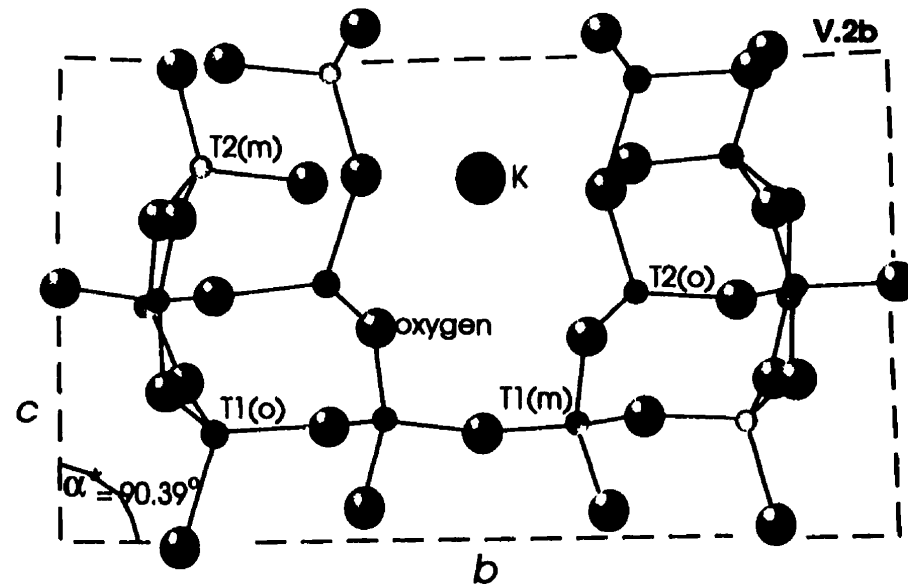
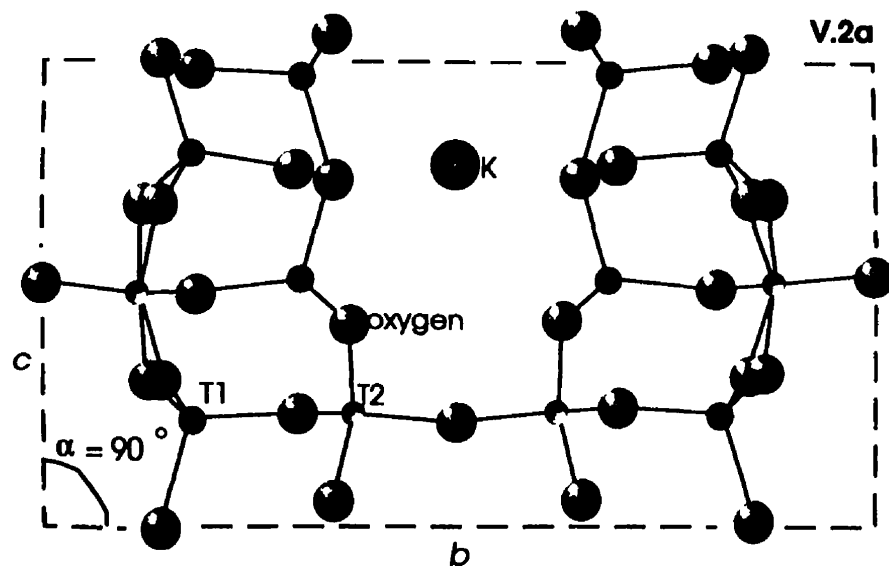
	<u>a</u>	<u>b</u>	<u>c</u>	<u>a</u>	<u>b</u>	<u>Y</u>	<u>V</u>	<u>#</u>	<u>a</u>	<u>b</u>	<u>c</u>	<u>a</u>	<u>b</u>	<u>Y</u>
36A	8.5850	12.9674	7.2156	90.545	115.968	88.215	721.82	31	0.129622	0.077155	0.154153	90.264	64.036	91.720
trans	0.0025	0.0026	0.0015	0.027	0.018	0.034	0.24		0.000044	0.000016	0.000035	0.026	0.018	0.034
	8.1409	12.7941	7.1611	94.252	116.593	87.796	665.11	52	0.137372	0.078378	0.156482	86.347	63.490	90.338
	0.0010	0.0012	0.0009	0.013	0.011	0.009	0.10		0.000019	0.000007	0.000021	0.012	0.011	0.009
36A	8.5833	12.9705	7.2194	90.614	115.924	87.831	722.33	56	0.129629	0.077155	0.154016	90.372	64.080	92.113
pink	0.0018	0.0019	0.0013	0.017	0.015	0.019	0.17		0.000027	0.000011	0.000025	0.016	0.015	0.018
	8.1393	12.7911	7.1605	94.247	116.610	87.733	664.68	56	0.137420	0.078397	0.156511	86.385	63.477	90.410
	0.0010	0.0012	0.0008	0.012	0.010	0.009	0.09		0.000017	0.000007	0.000018	0.012	0.010	0.008
36A	8.5851	12.9589	7.2188	90.578	115.918	87.881	721.82	60	0.129592	0.077222	0.154023	90.387	64.085	92.075
overgrowth	0.0011	0.0014	0.0009	0.012	0.010	0.011	0.11		0.000016	0.000008	0.000019	0.012	0.010	0.011
and	8.1327	12.7898	7.1592	94.260	116.605	87.771	663.96	47	0.137525	0.078406	0.156539	86.351	63.479	90.362
substrate	0.0011	0.0014	0.0009	0.015	0.012	0.010	0.11		0.000021	0.000009	0.000024	0.014	0.012	0.010
37	8.5786	12.9799	7.2135	90	116.059	90	721.56	26	0.129761	0.077043	0.154318	90	63.941	90
drusy	0.0029	0.0042	0.0021		0.027		0.32		0.000057	0.000025	0.000047		0.027	
bt gr	8.1390	12.7897	7.1585	94.241	116.586	87.727	664.54	46	0.137397	0.078405	0.156521	86.393	63.501	90.421
white, *	0.0012	0.0020	0.0009	0.017	0.012	0.013	0.12		0.000022	0.000012	0.000022	0.016	0.012	0.012
37	8.5872	12.9685	7.2175	90.640	115.985	87.806	721.97	37	0.129638	0.077168	0.154136	90.357	64.020	92.129
drusy	0.0015	0.0015	0.0011	0.018	0.013	0.016	0.14		0.000022	0.000009	0.000023	0.018	0.013	0.016
bt gr	8.1378	12.7887	7.1585	94.264	116.614	87.750	664.20	44	0.137450	0.078413	0.156568	86.357	63.471	90.383
pink, *	0.0011	0.0018	0.0011	0.017	0.015	0.013	0.13		0.000027	0.000011	0.000027	0.017	0.015	0.013
38B	8.5871	12.9808	7.2094	90	116.068	90	721.87	30	0.129642	0.077037	0.154415	90	63.932	90
pheno	0.0017	0.0026	0.0015		0.017		0.20		0.000032	0.000016	0.000032		0.017	
	8.1392	12.7842	7.1590	94.216	116.617	87.803	664.16	29	0.137429	0.078435	0.156554	86.384	63.465	90.347
	0.0017	0.0022	0.0009	0.017	0.014	0.019	0.15		0.000029	0.000014	0.000019	0.016	0.014	0.018
38B	8.5874	12.9604	7.2141	90.412	115.997	88.271	721.32	23	0.129618	0.077195	0.154225	90.385	64.004	91.723
matrix	0.0023	0.0038	0.0021	0.038	0.036	0.028	0.31		0.000042	0.000023	0.000064	0.040	0.036	0.029
	8.1406	12.7880	7.1563	94.273	116.602	87.770	664.25	38	0.137388	0.078418	0.156602	86.337	63.483	90.357
	0.0012	0.0016	0.0009	0.014	0.013	0.013	0.12		0.000026	0.000010	0.000023	0.012	0.013	0.011
39	8.5839	12.9701	7.2217	90.628	115.930	87.788	722.53	53	0.129629	0.077159	0.153976	90.378	64.074	92.155
pink	0.0015	0.0012	0.0009	0.011	0.011	0.012	0.13		0.000028	0.000007	0.000017	0.011	0.011	0.012
parthite	8.1450	12.7854	7.1621	94.254	116.639	87.779	664.81	43	0.137358	0.078432	0.156523	86.353	63.445	90.353
	0.0011	0.0020	0.0014	0.022	0.013	0.014	0.15		0.000025	0.000012	0.000031	0.021	0.013	0.013
40A	8.5851	12.9753	7.1978	90	116.145	90	719.75	25	0.129758	0.077070	0.154766	90	63.855	90
pheno	0.0033	0.0039	0.0016		0.021		0.27		0.000056	0.000023	0.000039		0.021	
grey core	8.1372	12.8058	7.1524	94.086	116.576	88.137	664.86	27	0.137412	0.078289	0.156647	86.362	63.484	90.040
	0.0017	0.0033	0.0015	0.032	0.020	0.023	0.21		0.000037	0.000021	0.000040	0.032	0.020	0.025
40A	8.5890	12.9760	7.2102	90	116.020	90	722.13	17	0.129560	0.077065	0.154336	90	63.980	90
pheno	0.0033	0.0029	0.0016		0.033		0.36		0.000071	0.000017	0.000049		0.033	
white rim	8.1379	12.7879	7.1558	94.309	116.566	87.798	664.17	39	0.137389	0.078422	0.156573	86.282	63.517	90.310
*	0.0012	0.0018	0.0011	0.018	0.012	0.013	0.13		0.000022	0.000010	0.000023	0.018	0.012	0.013
40A	8.5867	12.9813	7.2051	90	116.027	90	721.68	30	0.129602	0.077034	0.154454	90	63.973	90
matrix	0.0011	0.0014	0.0008		0.009		0.11		0.000018	0.000008	0.000016		0.009	
	8.1376	12.7916	7.1548	94.166	116.572	87.888	664.33	37	0.137400	0.078384	0.156582	86.398	63.505	90.280
	0.0009	0.0011	0.0011	0.016	0.011	0.012	0.11		0.000017	0.000007	0.000027	0.015	0.011	0.010
40D	8.5977	12.9768	7.2031	90	116.006	90	722.28	22	0.129414	0.077061	0.154470	90	63.994	90
matrix	0.0032	0.0034	0.0022		0.034		0.30		0.000052	0.000020	0.000043		0.034	
	8.1361	12.7918	7.1572	94.230	116.588	87.822	664.29	36	0.137447	0.078390	0.156559	86.359	63.493	90.321
	0.0013	0.0020	0.0012	0.017	0.012	0.014	0.12		0.0022	0.0012	0.0026	0.017	0.012	0.013

	<u>a</u>	<u>b</u>	<u>c</u>	<u>α</u>	<u>β</u>	<u>γ</u>	<u>ν</u>	<u>#</u>	<u>a</u> *	<u>b</u> *	<u>c</u> *	<u>α</u> *	<u>β</u> *	<u>γ</u> *
41	8.5868	12.9782	7.1994	90	115.998	90	721.13	22	0.129569	0.077052	0.154537	90	64.002	90
pheno	0.0015	0.0030	0.0012		0.020		0.20		0.000027	0.000018	0.000035		0.020	
	8.1352	12.7868	7.1561	94.219	116.538	87.699	664.14	28	0.137404	0.078421	0.156503	86.432	63.551	90.467
	0.0022	0.0034	0.0025	0.046	0.035	0.029	0.32		0.000048	0.000021	0.000071	0.049	0.035	0.035
41	8.5969	12.9695	7.2067	90	115.896	90	722.85	15	0.129305	0.077104	0.154248	90	64.104	90
matrix	0.0033	0.0054	0.0023		0.029		0.36		0.000058	0.000032	0.000050		0.029	
	8.1374	12.7910	7.1582	94.277	116.597	87.707	664.34	30	0.137437	0.078400	0.156549	86.364	63.492	90.426
	0.0018	0.0027	0.0016	0.026	0.018	0.021	0.18		0.000036	0.000016	0.000031	0.025	0.018	0.020
49A	8.6017	12.9744	7.2058	90	116.047	90	722.50	25	0.129399	0.077075	0.154466	90	63.953	90
pheno	0.0024	0.0036	0.0021		0.023		0.27		0.000042	0.000022	0.000050		0.023	
	8.1411	12.7934	7.1562	94.095	116.604	87.939	664.71	33	0.137380	0.078366	0.156585	86.452	63.469	90.256
	0.0019	0.0030	0.0017	0.023	0.023	0.029	0.21		0.000032	0.000019	0.000049	0.023	0.023	0.029
49A	8.5937	12.9755	7.2085	90	116.032	90	722.25	21	0.129503	0.077068	0.154389	90	63.968	90
matrix	0.0021	0.0040	0.0016		0.029		0.26		0.000035	0.000024	0.000039		0.029	
	8.1376	12.7763	7.1546	94.213	116.565	87.924	663.52	26	0.137392	0.078483	0.156590	86.326	63.510	90.217
	0.0020	0.0041	0.0015	0.022	0.017	0.022	0.20		0.000040	0.000025	0.000031	0.019	0.017	0.020
49B	8.1420	12.7962	7.1575	94.256	116.656	87.888	664.62	22	0.137427	0.078365	0.156656	86.298	63.421	90.229
inclusion	0.0022	0.0024	0.0017	0.030	0.030	0.022	0.19		0.000030	0.000016	0.000037	0.033	0.030	0.025
49B	8.6023	12.9733	7.2046	90	116.132	90	721.85	20	0.129484	0.077082	0.154603	90	63.868	90
matrix	0.0029	0.0042	0.0022		0.033		0.33		0.000047	0.000025	0.000052		0.033	
	8.1344	12.8051	7.1543	94.113	116.596	88.096	664.63	26	0.137483	0.078296	0.156634	86.353	63.467	90.072
	0.0020	0.0032	0.0019	0.034	0.028	0.022	0.24		0.000046	0.000020	0.000048	0.034	0.028	0.023
56K	8.5774	12.9828	7.2109	90	116.004	90	721.70	31	0.129718	0.077025	0.154300		63.996	
pheno	0.0011	0.0016	0.0009		0.010		0.12		0.000016	0.000009	0.000017		0.010	
	8.1421	12.7951	7.1590	94.198	116.614	87.830	664.99	29	0.137377	0.078366	0.156549	86.391	63.466	90.327
	0.0017	0.0029	0.0013	0.022	0.015	0.020	0.17		0.000034	0.000017	0.000030	0.020	0.015	0.017
56K	8.5696	12.9696	7.2107	90.527	115.928	88.269	720.42	23	0.129806	0.077139	0.154206	90.255	64.076	91.668
matrix	0.0044	0.0056	0.0026	0.050	0.034	0.045	0.38		0.000068	0.000034	0.000055	0.043	0.034	0.045
	8.1346	12.7895	7.1553	94.213	116.572	87.802	663.98	41	0.137454	0.078402	0.156573	86.388	63.510	90.353
	0.0011	0.0018	0.0009	0.015	0.015	0.014	0.12		0.000023	0.000011	0.000024	0.015	0.015	0.014
58B	8.5782	12.9781	7.2105	90	116.061	90	721.12	26	0.129768	0.077053	0.154384	90	63.939	90
green Kfs	0.0017	0.0018	0.0012		0.017		0.17		0.000028	0.000011	0.000024		0.017	
*	8.1406	12.7904	7.1582	94.188	116.600	87.865	664.65	47	0.137384	0.078394	0.156549	86.384	63.478	90.292
	0.0011	0.0016	0.0009	0.015	0.011	0.014	0.11		0.000021	0.000009	0.000019	0.014	0.010	0.012
58K	8.5790	12.9661	7.2174	90.490	116.037	88.381	721.06	18	0.129778	0.077156	0.154206	90.246	63.966	91.562
green Kfs	0.0037	0.0058	0.0025	0.063	0.042	0.057	0.47		0.000065	0.000035	0.000060	0.060	0.042	0.055
	8.1403	12.7913	7.1601	94.269	116.595	87.782	664.80	38	0.137384	0.078397	0.156508	86.336	63.489	90.346
	0.0010	0.0017	0.0009	0.012	0.011	0.010	0.10		0.000023	0.000010	0.000020	0.012	0.011	0.009
61A-1	8.5863	12.9796	7.2096	90	116.015	90	722.07	35	0.129595	0.077044	0.154343	90	63.985	90
pheno	0.0013	0.0016	0.0010		0.012		0.13		0.000021	0.000010	0.000019		0.012	
	8.1384	12.8012	7.1549	94.216	116.545	87.852	665.02	34	0.137354	0.078331	0.156550	86.360	63.535	90.298
	0.0013	0.0024	0.0012	0.022	0.016	0.019	0.16		0.000028	0.000014	0.000033	0.021	0.016	0.018
61A-1	8.5908	12.9584	7.2094	90	116.070	90	720.92	17	0.129588	0.077170	0.154419	90	63.930	90
matrix	0.0058	0.0066	0.0042		0.067		0.67		0.000133	0.000039	0.000085		0.067	
	8.1389	12.7933	7.1512	94.184	116.56	87.905	664.23	34	0.137368	0.078375	0.156654	86.369	63.513	90.252
	0.0008	0.0012	0.0013	0.013	0.011	0.010	0.12		0.000019	0.000007	0.000029	0.013	0.011	0.011
61A-1	8.5955	12.9842	7.2067	90	116.012	90	722.83	19	0.129454	0.077017	0.154400	90	63.988	90
matrix	0.0034	0.0050	0.0027		0.043		0.42		0.000058	0.000030	0.000072		0.043	
	8.1404	12.7961	7.1532	94.246	116.574	87.906	664.56	29	0.137357	0.078365	0.156638	86.299	63.501	90.219
	0.0015	0.0041	0.0015	0.027	0.017	0.018	0.21		0.000032	0.000026	0.000032	0.028	0.017	0.019

	<u>a</u>	<u>b</u>	<u>c</u>	<u>α</u>	<u>β</u>	<u>γ</u>	<u>ν</u>	<u>θ</u>	<u>a</u> *	<u>b</u> *	<u>c</u> *	<u>α</u> *	<u>β</u> *	<u>γ</u> *
64	8.5799	12.9888	7.2028	90	115.989	90	721.53	34	0.129664	0.076990	0.154454	90	64.011	90
yellow	0.0013	0.0016	0.0013		0.013		0.15		0.000023	0.000010	0.000027		0.013	
rim of	8.1383	12.7944	7.1578	94.193	116.629	87.934	664.46	42	0.137457	0.078370	0.156603	86.344	63.445	90.210
pheno	0.0010	0.0018	0.0008	0.013	0.011	0.013	0.11		0.000019	0.000011	0.000020	0.013	0.011	0.012
64	8.5920	12.9831	7.1994	90	116.029	90	721.64	35	0.129524	0.077023	0.154580	90	63.971	90
whitish	0.0014	0.0020	0.0009		0.013		0.15		0.000028	0.000012	0.000019		0.013	
core of	8.1419	12.7922	7.1582	94.233	116.616	87.820	664.72	26	0.137381	0.078387	0.156574	86.357	63.466	90.320
pheno	0.0018	0.0026	0.0011	0.025	0.021	0.020	0.18		0.000030	0.000016	0.000034	0.025	0.022	0.021
64B	8.5950	12.9851	7.2037	90	116.036	90	722.40	27	0.129486	0.077011	0.154496	90	63.964	90
matrix	0.0023	0.0037	0.0021		0.028		0.28		0.000037	0.000022	0.000046		0.028	
	8.1361	12.7863	7.1538	94.148	116.551	87.949	663.98	25	0.137401	0.078415	0.156578	86.388	63.522	90.223
	0.0020	0.0057	0.0019	0.029	0.028	0.031	0.27		0.000043	0.000036	0.000041	0.029	0.028	0.031
66	8.5819	12.9609	7.2160	90.440	116.016	88.173	720.92	40	0.129728	0.077197	0.154210	90.402	63.984	91.818
red perth	0.0017	0.0020	0.0010	0.021	0.015	0.023	0.17		0.000026	0.000012	0.000020	0.018	0.015	0.021
	8.1378	12.7806	7.1528	94.233	116.572	87.837	663.53	38	0.137398	0.078459	0.156634	86.347	63.508	90.303
	0.0011	0.0015	0.0010	0.014	0.013	0.013	0.12		0.000022	0.000009	0.000021	0.014	0.013	0.013
66	8.5796	12.9701	7.2139	90.543	116.001	88.148	721.11	30	0.129744	0.077142	0.154234	90.299	64.003	91.795
cream core	0.0022	0.0031	0.0016	0.028	0.018	0.028	0.22		0.000029	0.000018	0.000038	0.025	0.018	0.024
	8.1362	12.7913	7.1550	94.235	116.620	87.893	663.88	32	0.137482	0.078393	0.156659	86.318	63.456	90.236
	0.0010	0.0020	0.0008	0.014	0.012	0.016	0.12		0.000021	0.000012	0.000020	0.015	0.013	0.016
68B	8.5854	12.9857	7.2033	90	116.023	90	721.66	27	0.129617	0.077008	0.154488	90	63.977	90
pheno	0.0021	0.0028	0.0019		0.023		0.23		0.000037	0.000017	0.000040		0.023	
	8.1401	12.7892	7.1563	94.140	116.552	87.955	664.69	39	0.137334	0.078396	0.156522	86.392	63.521	90.219
	0.0011	0.0015	0.0011	0.016	0.012	0.013	0.12		0.000023	0.000009	0.000028	0.015	0.012	0.011
68B	8.5711	12.9790	7.2095	90	116.057	90	720.49	22	0.129872	0.077048	0.154399	90	63.943	90
matrix	0.0022	0.0040	0.0015		0.023		0.27		0.000038	0.000024	0.000037		0.023	
	8.1365	12.7927	7.1555	94.210	116.605	87.877	664.13	34	0.137460	0.078382	0.156620	86.353	63.472	90.267
	0.0009	0.0017	0.0008	0.013	0.011	0.012	0.11		0.000020	0.000011	0.000020	0.013	0.011	0.011
82K	8.5887	12.9810	7.2029	90	116.025	90	721.62	30	0.129570	0.077036	0.154500	90	63.975	90
microgr.	0.0019	0.0023	0.0013		0.013		0.17		0.000034	0.000014	0.000026		0.013	
pheno	8.1384	12.7988	7.1564	94.241	116.623	87.857	664.55	39	0.137450	0.078348	0.156631	86.329	63.455	90.274
	0.0010	0.0014	0.0010	0.015	0.012	0.015	0.11		0.000019	0.000008	0.000022	0.014	0.012	0.014
82K	8.5842	12.9712	7.2141	90.556	115.941	88.312	722.02	20	0.129597	0.077128	0.154149	90.203	64.064	91.607
microgr.	0.0036	0.0041	0.0023	0.060	0.042	0.061	0.39		0.000060	0.000025	0.000067	0.057	0.042	0.058
matrix	8.1412	12.7860	7.1543	94.230	116.613	87.795	663.99	47	0.137390	0.078426	0.156653	86.373	63.470	90.350
θ	0.0011	0.0013	0.0013	0.016	0.015	0.011	0.14		0.000023	0.000008	0.000033	0.016	0.015	0.011
83R	8.5879	12.9582	7.2109	90.519	115.846	88.180	721.81	22	0.129448	0.077211	0.154095	90.305	64.157	91.771
c.-g.	0.0044	0.0059	0.0031	0.049	0.057	0.065	0.53		0.000106	0.000036	0.000081	0.047	0.057	0.065
hyp. gr.	8.1360	12.7866	7.1576	94.263	116.591	87.742	664.00	45	0.137452	0.078426	0.156554	86.362	63.495	90.393
θ	0.0009	0.0013	0.0009	0.013	0.011	0.011	0.10		0.000018	0.000008	0.000020	0.012	0.011	0.011
83R	8.5747	12.9742	7.2108	90	116.015	90	720.92	25	0.129772	0.077076	0.154316	90	63.985	90
f.-g.	0.0025	0.0024	0.0015		0.016		0.20		0.000039	0.000014	0.000025		0.016	
leucogr.	8.5987	12.9663	7.2221	90.157	116.126	88.468	722.65	17	0.129584	0.077154	0.154230	90.576	63.868	91.629
dyke	0.0059	0.0051	0.0029	0.053	0.045	0.039	0.43		0.000070	0.000030	0.000060	0.050	0.045	0.034
	8.1406	12.7908	7.1592	94.257	116.626	87.739	664.54	48	0.137417	0.078400	0.156564	86.371	63.460	90.398
	0.0010	0.0015	0.0009	0.015	0.012	0.011	0.12		0.000021	0.000009	0.000025	0.016	0.012	0.013

* intermediate microcline also is present, but in too small a volume to give a sufficient number of peaks for a cell-parameter refinement. θ A trace of orthoclase also is present.

Appendix V.2: Schematic illustration of structure, site occupancy, and monoclinic - triclinic inversion of alkali feldspar. **a)** Monoclinic (disordered) structure of sanidine/orthoclase, **b)** Triclinic (ordered) structure of low microcline, and **c)** Topology of the tetrahedral nodes of the feldspar network, projected down the *c*-axis onto the *ab*-plane. The diagram shows slightly distorted double-crankshaft (magnified in inset) produced by the joining of alternate 4-membered rings along the *a*-axis. Adjacent crankshafts are connected across mirror planes which relate nodes T_2O and T_2m . The structure also illustrates the loss of symmetry associated with Al-Si order and monoclinic to triclinic inversion. Both Figures were modified from Ribbe (1983).



Monoclinic structure

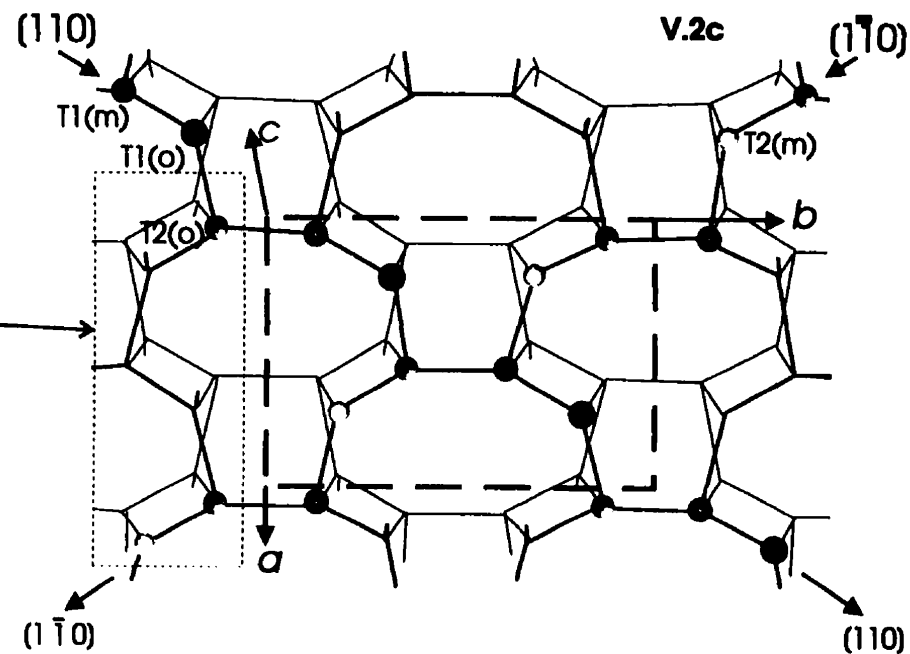
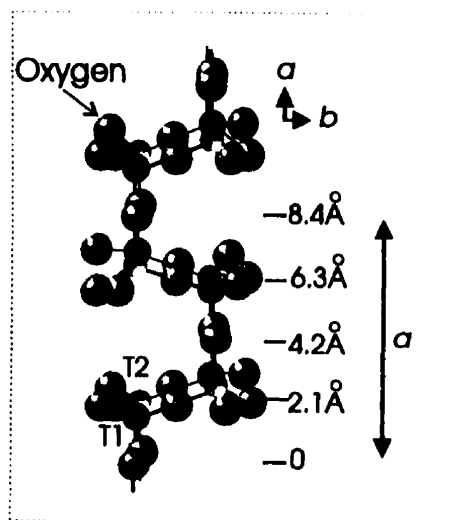
- T1
- T2

Triclinic structure

- T2(m)
- T1(m)
- T2(o)
- T1(o)

Other symbols

- Alkali site (e.g., K)
- Oxygen



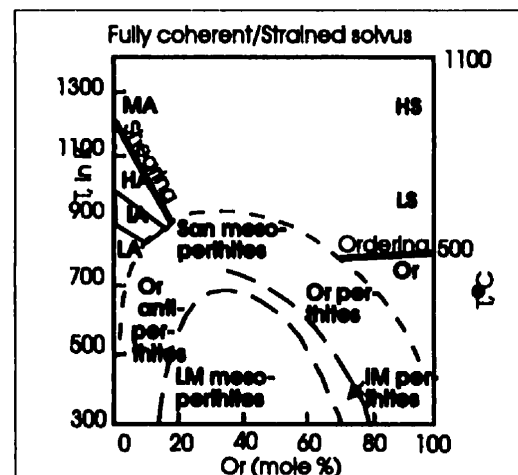
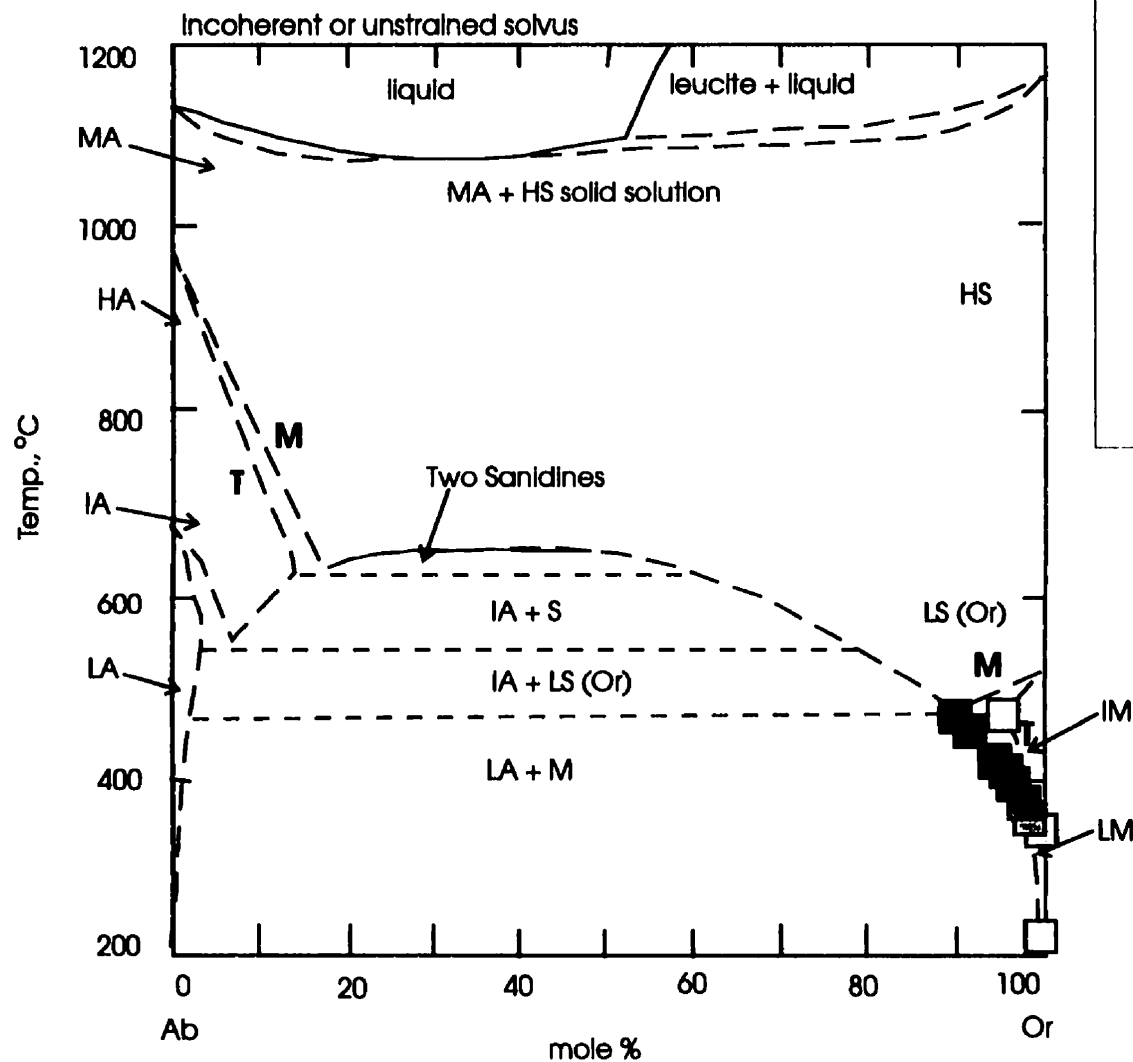
Appendix V.3: T-X phase diagram for alkali feldspars. Shaded squares indicate composition (N_{Or}) of orthoclase, and defines a metastable extension of the unstrained alkali feldspar solvus below the indicated temperature of monoclinic (orthoclase/low sanidine) to triclinic (microcline) inversion. Open squares indicate range of microcline composition (N_{Or}). From the composition of orthoclase a range of temperature of closure of 300 to 450°C is inferred, whereas rocks bearing microcline $\pm$ orthoclase had a final equilibration temperature range of 200 to 450°C.

Insets include T-X diagram for fully coherent/strained solvus, and summary of N_{Or} values for orthoclase-bearing rock units in Kwandonkaya complex.

Abbreviations: MA = monalbite, HS = high sanidine, HA = high albite, IA = intermediate albite, LA = low albite, S = sanidine, LS = low sanidine, Or = orthoclase, M = microcline, IM = intermediate microcline, LM = low microcline. **M** = monoclinic, **T** = triclinic.

The T-X phase diagrams were modified from Smith (1974).

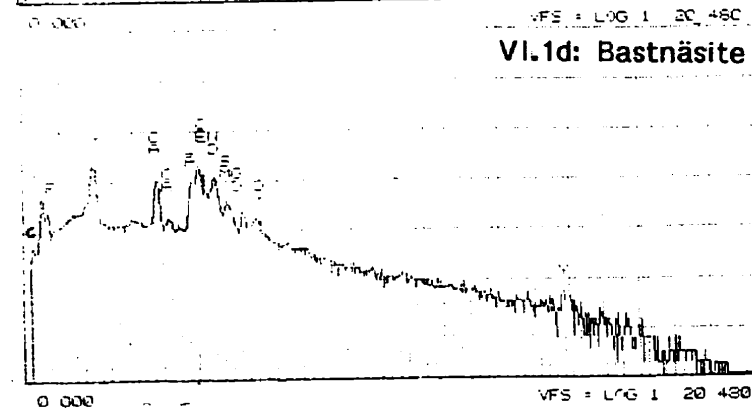
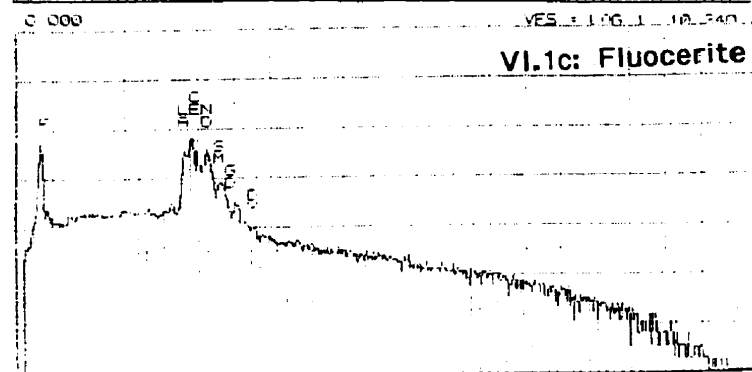
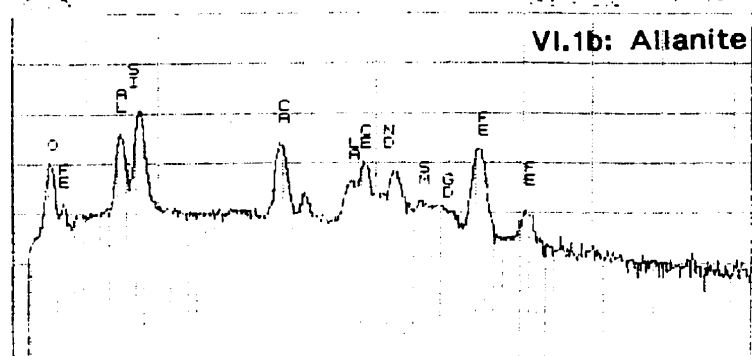
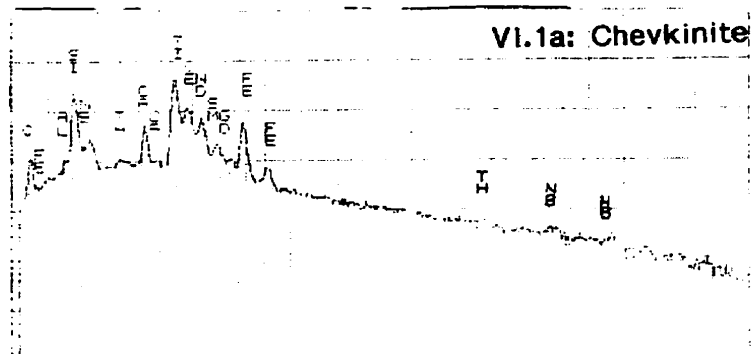
Appendix V.3



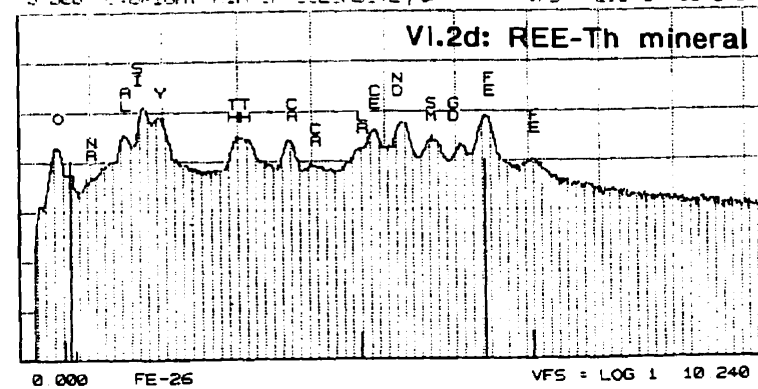
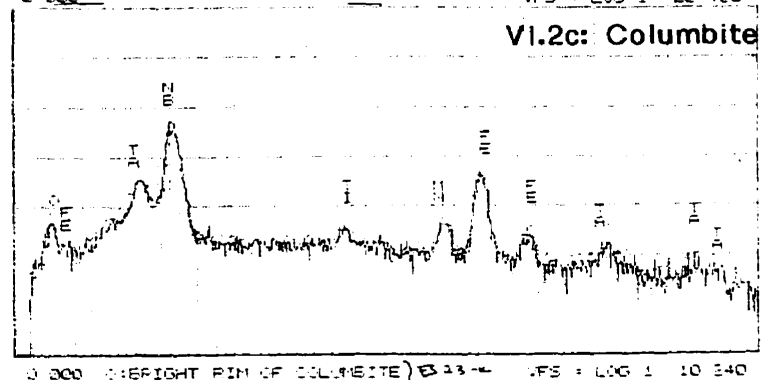
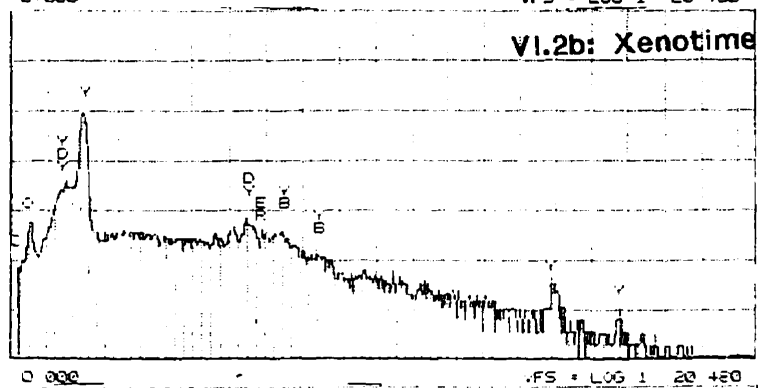
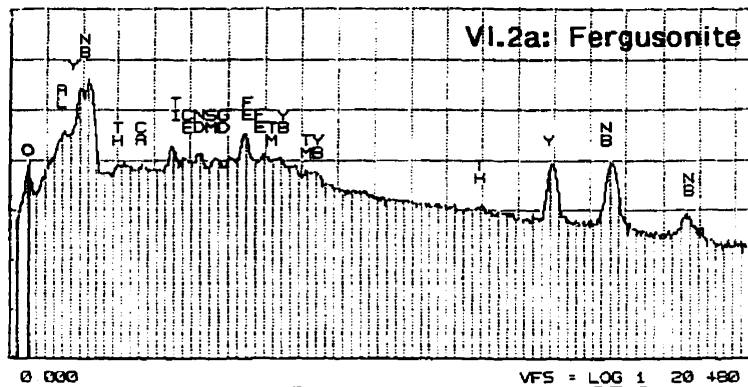
Summary of N(Or)-values for orthoclase-bearing rock units in Kwadonkaya complex

N(Or)	Sample#(Description in Table 5.2)
0.91	40A(Gcr)
0.92	22(Hsg)
0.95	41(Ph), 58(Or), 61A(Mtx), 83(Lgd)
0.96	37(Wpr), 82(Ph), 64(Yrp)
0.97	38(Ph), 56(Ph), 68B(Ph), 40A(Mtx), 64(Wph), 49B(Tandy)
0.98	40D(Mtx), 61A(Ph), 40A(Wrm), 49A(Mtx)
0.99	64(Mtx), 49A(Ph)
1.00	61A(Mtx), 41(Mtx)

Appendix VI.1: SEM - EDS spectra of: **a)** chevkinite, **b)** allanite, **c)** fluocerite, and **d)** bastnäsite



Appendix VI.2: SEM - EDS spectra of: a) fergusonite, b) xenotime, c) columbite, and d) unidentified REE-Th mineral.



Appendix VI.3: Photomicrographs of accessory minerals (chevkinite: a-c, allanite: d-e, and fluocerite: f).

a) Prismatic section of chevkinite (Ckt) along the boundary of orthoclase (Or), with opaque metamict rim, in hastingsite granite porphyry (40A).

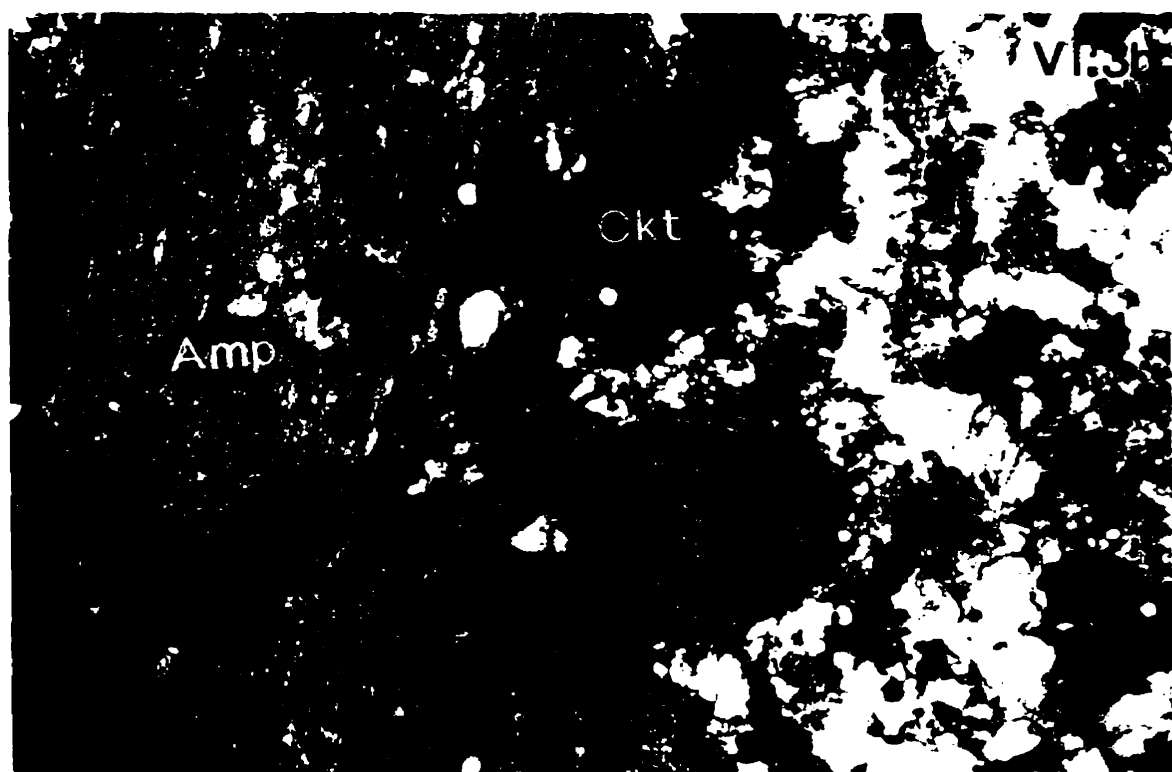
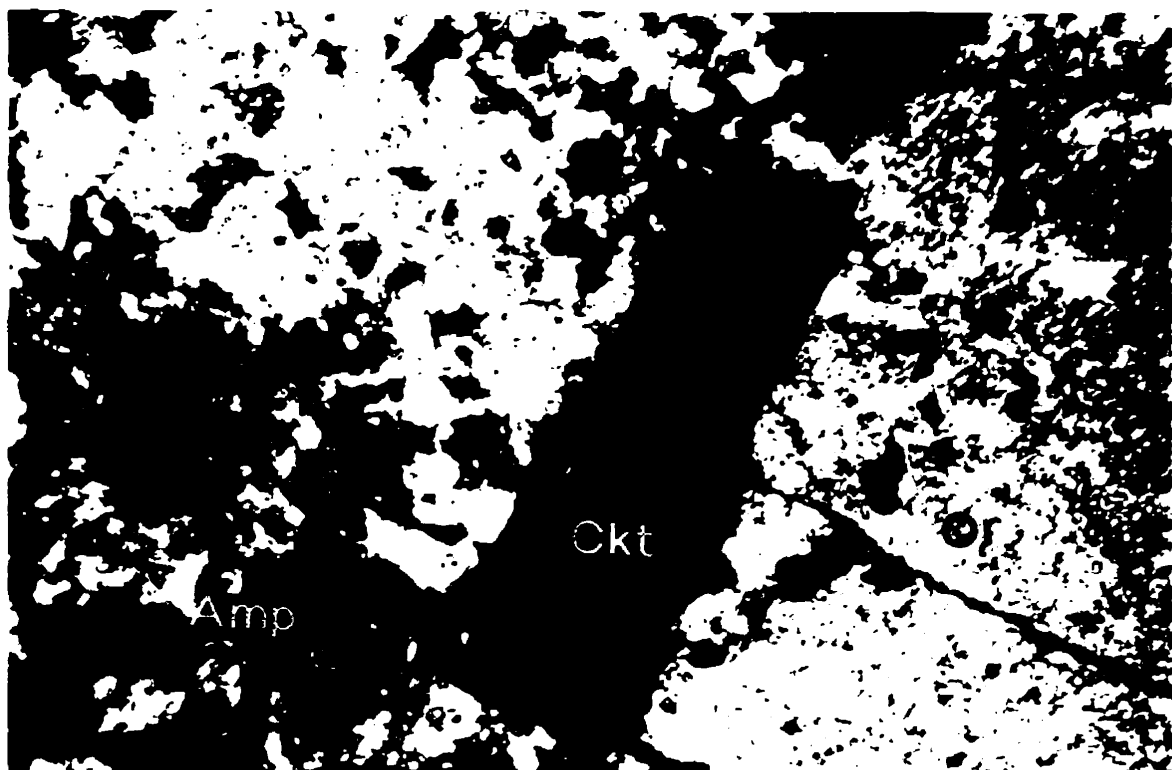
b) Equant cross-sections of chevkinite (Ckt) along the margin of hastingsitic hornblende (Amp) in sample 40A (note upper grain is sector-zoned).

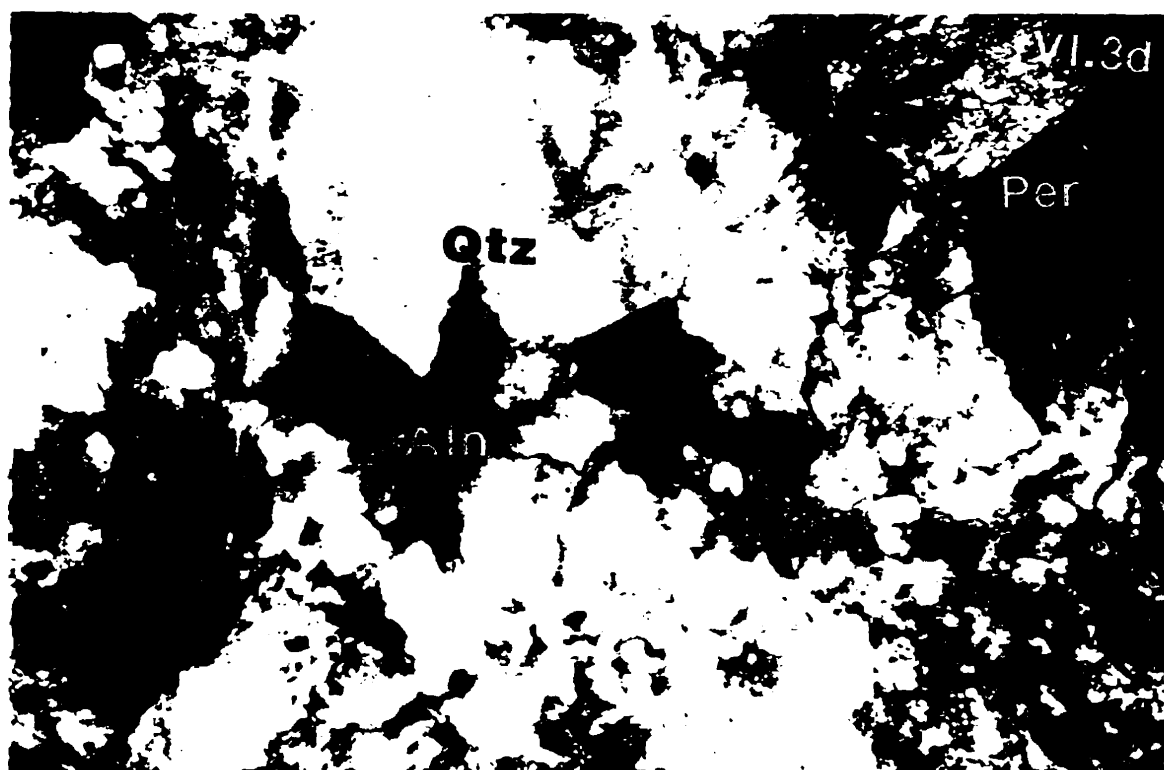
c) Drumstick-shaped chevkinite (Ckt) displaying flame-like structure with a bluish core and reddish brown to yellowish rim in microcrystalline groundmass of sample 40A.

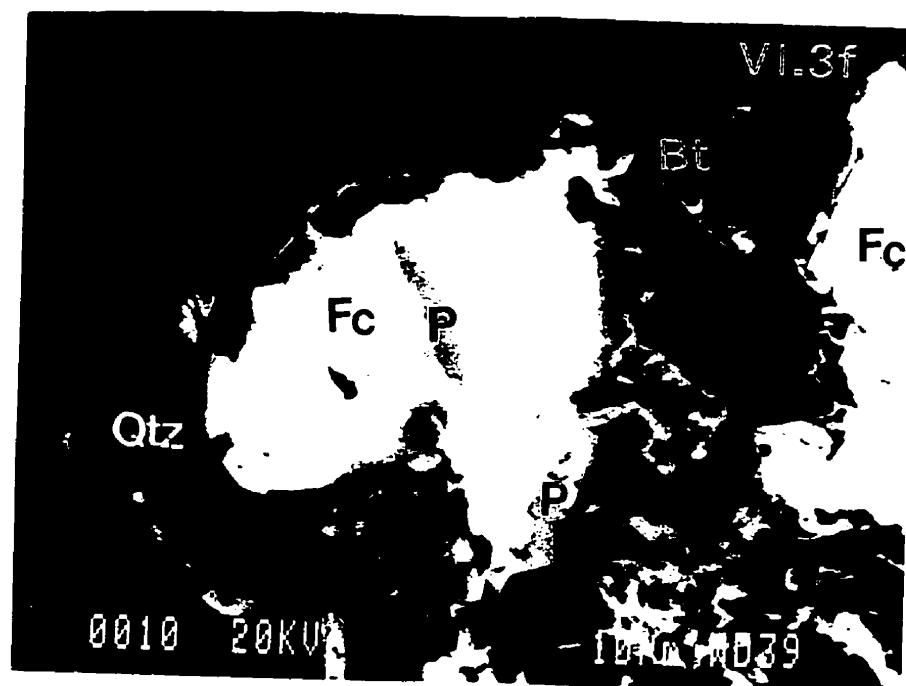
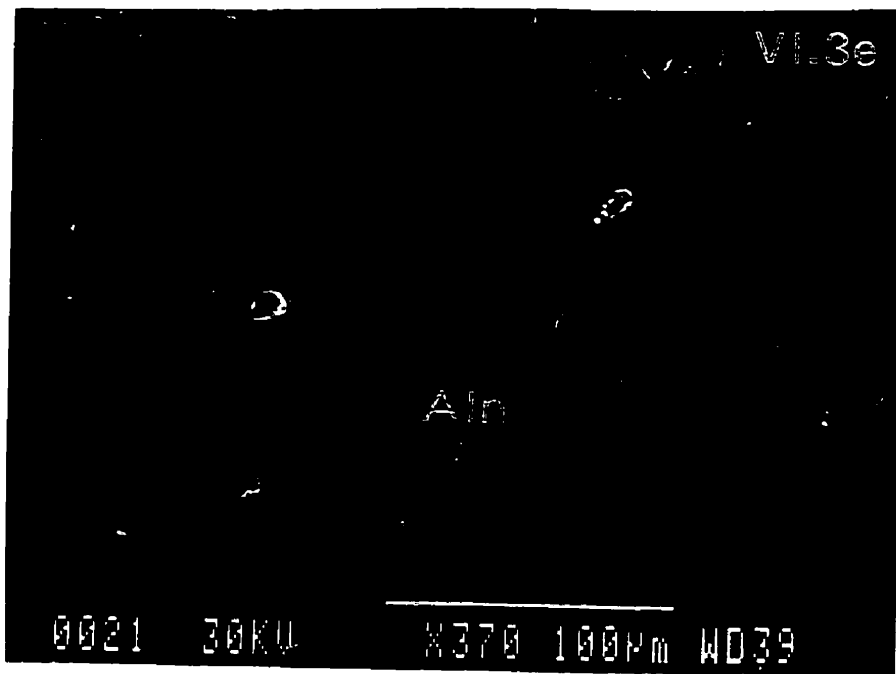
d) Zoned grains of allanite (Aln), in a microcrystalline groundmass of sample 40A, characterized by pyramidal terminations. [The width of all photomicrographs is 3.1mm; Mag = magnetite, Per = perthite].

e) SEM - BSE image of one of the allanite (Aln) in Appendix VI.3d; note that pyramidal terminations have distinct oscillatory zoning.

f) SEM - BSE image of fluocerite (Fc) inclusion in biotite in coarse-grained biotite granite (sample 26A). The fluocerite grains are replaced along the margin by parsite.



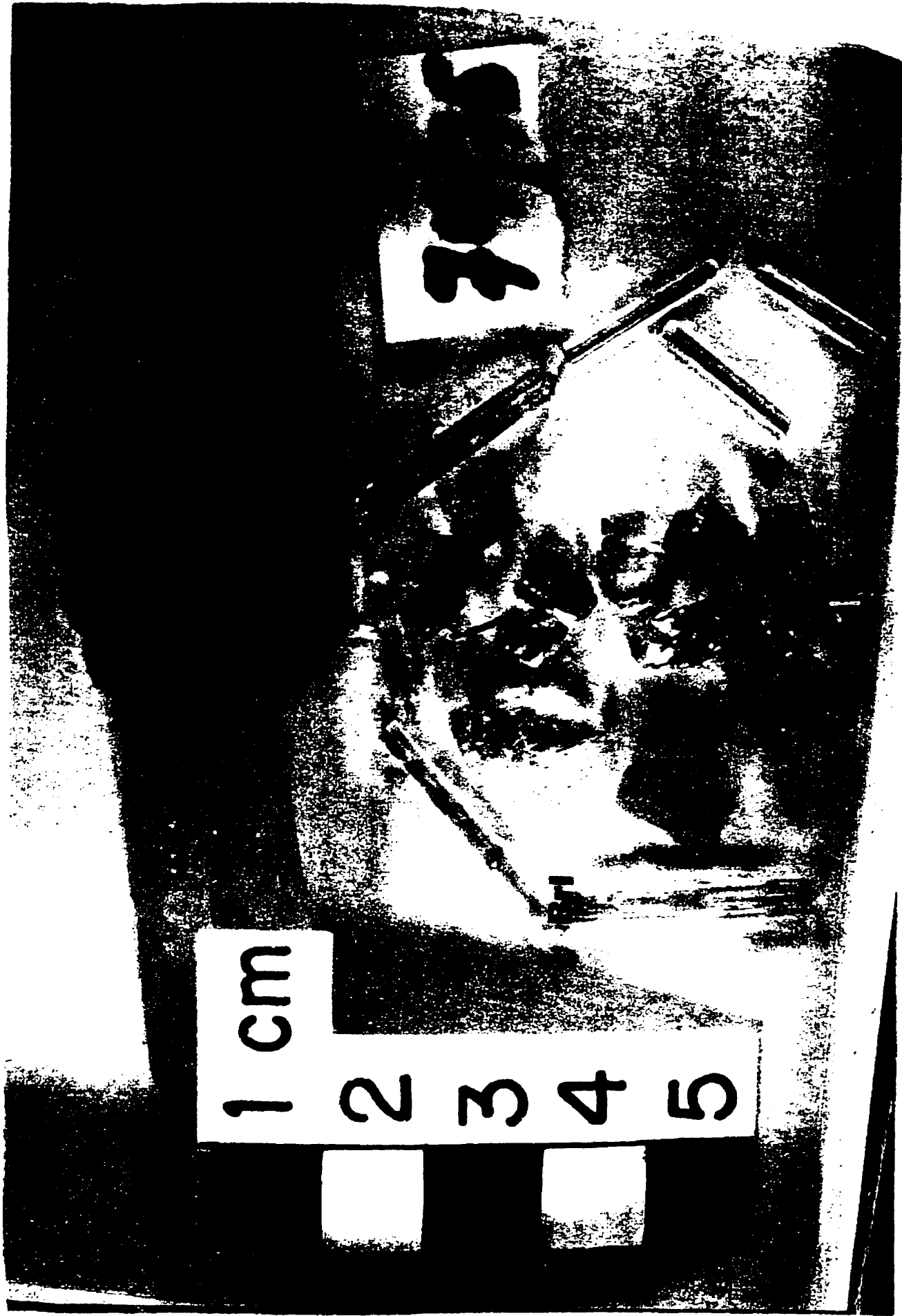




Appendix VI.4: Photograph of samples of beryl (Brl), quartz (Qtz), and topaz (Toz). The mineral samples were collected from alluvial, eluvial, and primary occurrences in Kwandonkaya complex (beryl or aquamarine from pegmatites in Panshanu center, fenestral quartz from Jubga village, and topaz from Jubga area).

705

1 cm
1 2 3 4 5



Appendix VI.5: SEM - BSE image of columbite (Col) - cassiterite (Cst) intergrowth. These minerals occur in albitized biotite granite (sample 23). This is an enlargement of Fig. 6.4d.



Col

Col

Cst

Appendix VII.1: Thin section photograph (a, b, e, and f), photomicrograph (c), and SEM-BSE image (f) of enclaves.

a) Thin section photograph of trachyandesite enclave. This has a sharp outline, and is zoned with a predominant, dark inner part and a greyish outer margin. The enclave also has xenocrysts of quartz (Qtz) with similar characteristics as those in the rhyolite porphyry host. The rhyolite porphyry host is comprised of phenocrysts of quartz (Qtz) and orthoclase (Or) in a microcrystalline groundmass of quartz and orthoclase. Accessory chevkinite (Ckt) also occurs.

b) Thin section photograph of another example of trachyandesite (Trand.) enclave in rhyolite porphyry. In this case, the enclave is characterized by pillowed margins, partially enclosed quartz phenocrysts, and arborescent inclusions (trachytic) in the dark enclave. In a) and b): Mag. X10. Blue circles indicate location of probe analyses points.

c) Photomicrograph of trachyandesite enclave (Trand.) in rhyolite porphyry (RPor) host. In this example, the enclave encloses a rhyolitic/felsic enclave (f-encl.).

d) SEM - BSE image of the margin of trachyandesite enclave (Trand.) in Appendix VII.1a. Although zoning is documented (from EDS area scans) with decreases in CaO, TiO₂, and MgO outwards and increases in alkalis in the same direction; the contrast between the margin and inner parts is not very obvious from the image.

e) Thin section photograph of microsyenodiorite (Microsyd.) enclave in hedenbergite granite porphyry. The enclave composition is shown in Table 7, whereas the granite porphyry is composed of phenocrysts of quartz (Qtz), orthoclase (Or), locally turbid perthite (Per), and hedenbergite largely replaced by amphibole (Amp) in a microcrystalline groundmass of quartz, feldspar, hedenbergite, and Fe-Ti oxides.

f) Thin section photograph of andesitic enclave (And.) within a host of biotite granite. In the andesitic enclave, prismatic phenocrysts of plagioclase (Plag) are set in a fine-grained matrix of biotite ($\pm$ amphibole) and quartz. Clots of microgranular quartz and plagioclase are common (may be fragments of basement).



Per

Qtz

Trand.

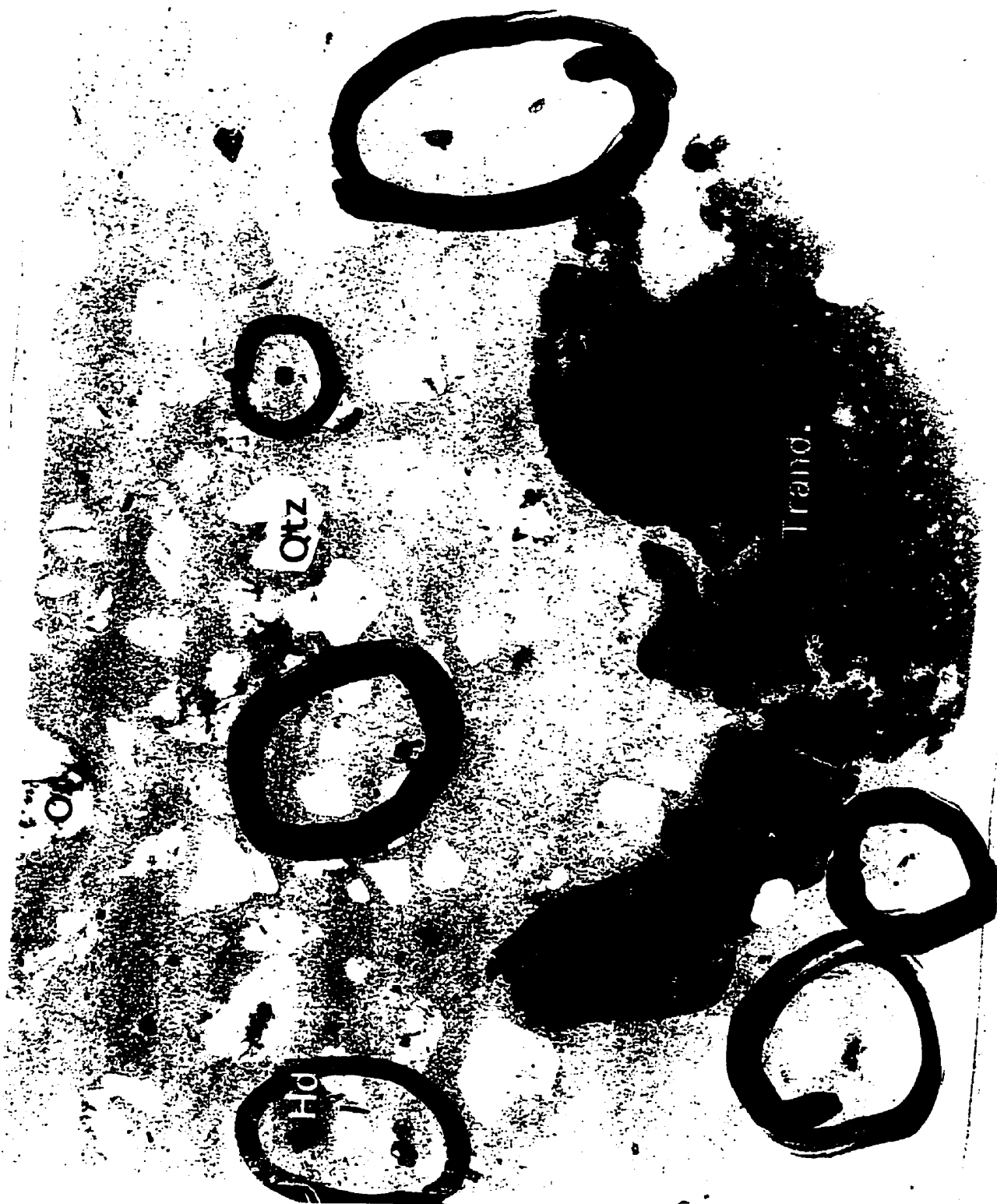
Or

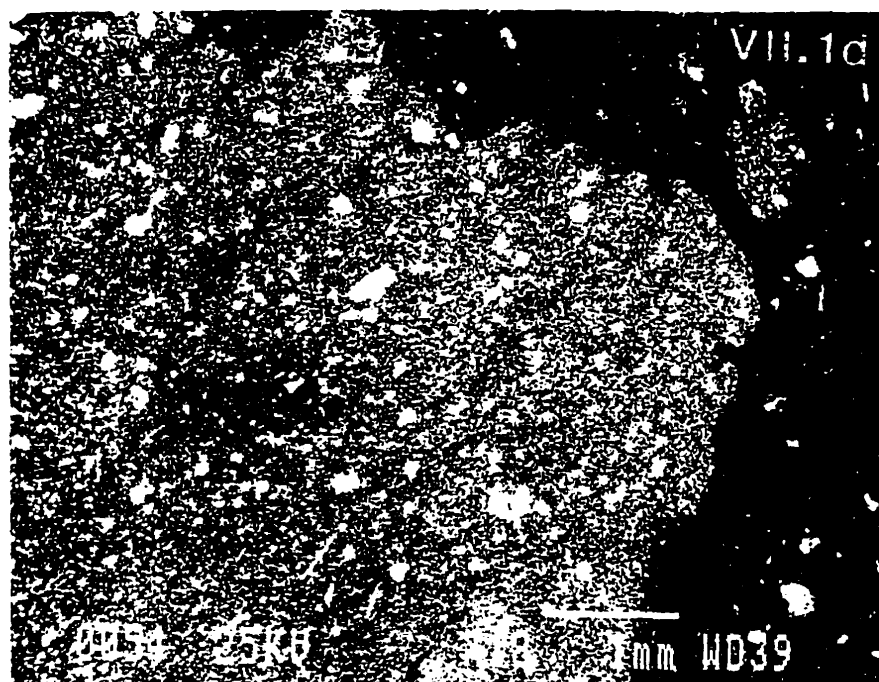
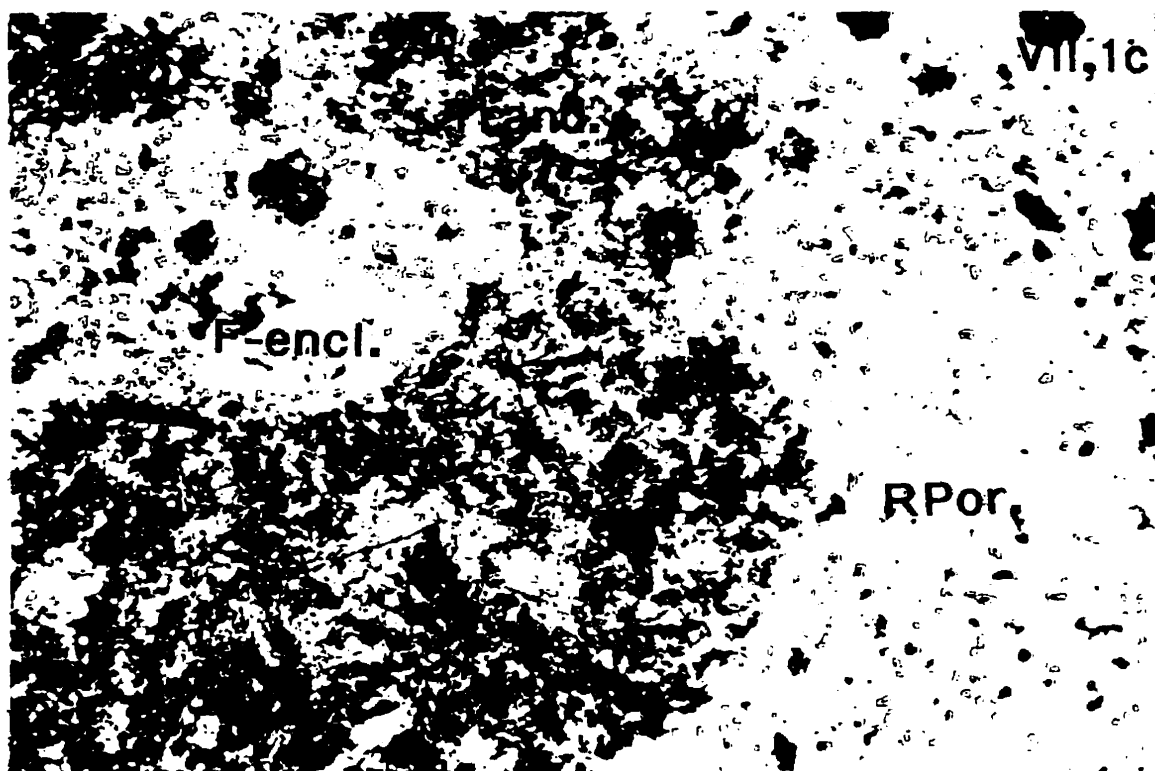
Or

Qtz

Qtz

VII-1b





Ville

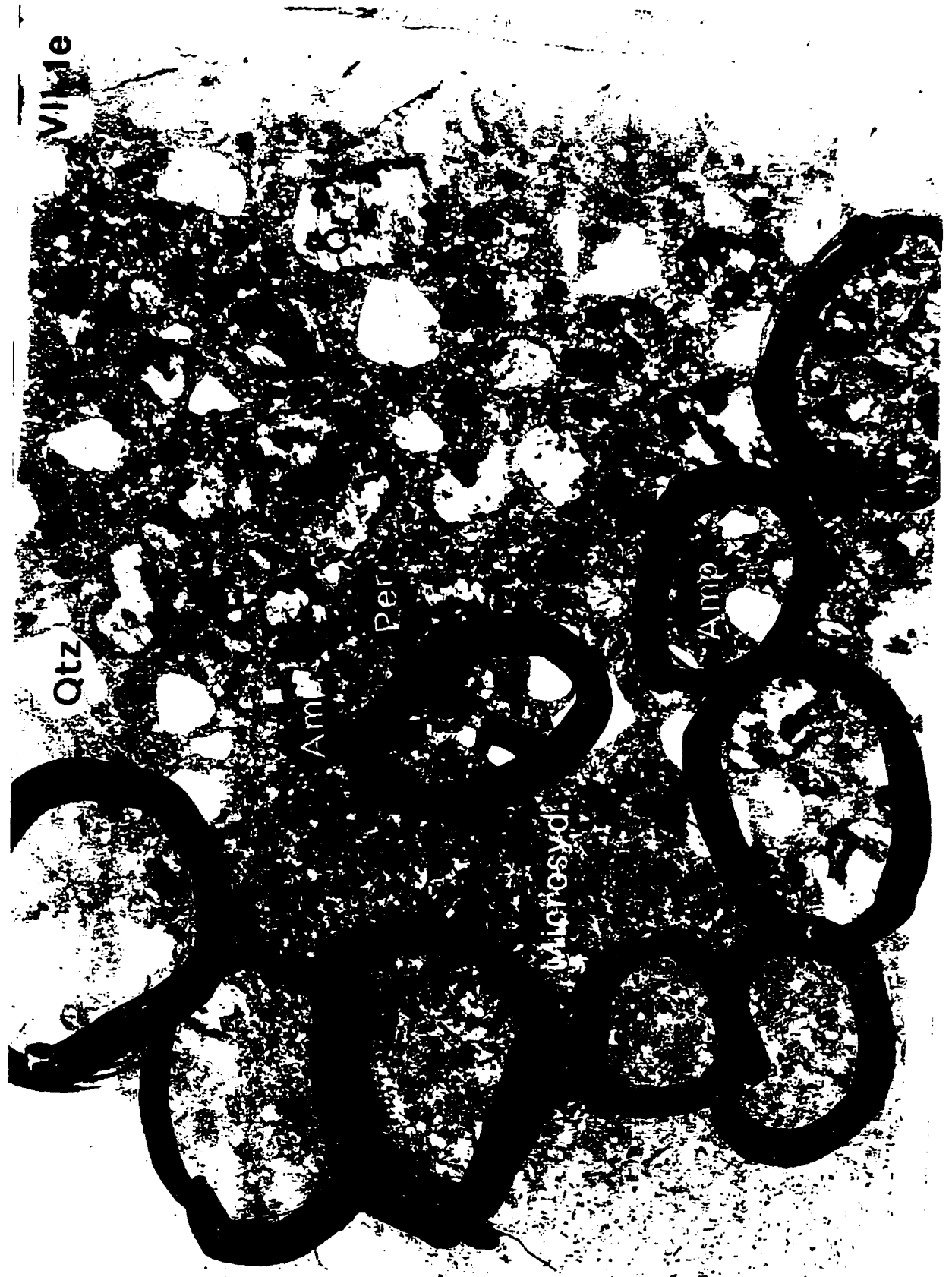
Qtz

Amph

Per

Amph

Microsyd.

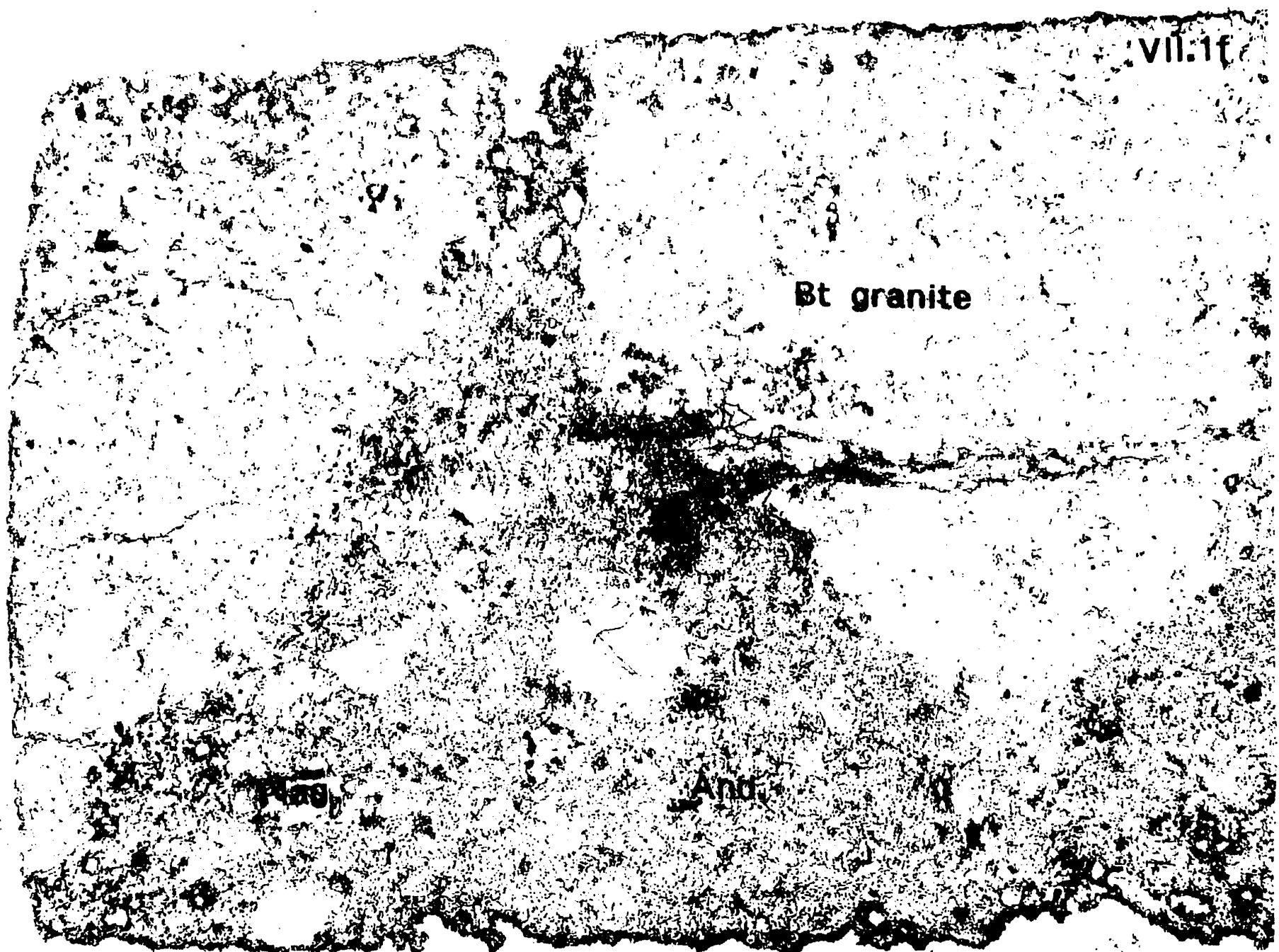


VII.1f

Bt granite

Plag.

And.



Appendix VII.2: Results of EMPA (spot) analyses of plagioclase in enclaves within plutonic rocks.

	73B.1C	3B.2Mr	3B.3Mt	73B.4C	3B.5Mr	73B.6	73B.7	73B.8	73B1.1	73B1.2
SiO2	61.81	62.91	61.53	52.83	53.00	63.75	53.20	65.62	53.63	53.03
Al2O3	23.71	22.62	23.52	29.08	29.36	22.98	29.36	21.54	29.25	29.13
Fe2O3	0.09	0.17	0.09	0.51	0.22	0.13	0.13	0.06	0.08	0.46
CaO	5.07	4.11	5.20	12.02	11.92	4.33	12.07	2.81	12.00	11.83
Na2O	8.63	9.01	6.52	4.32	4.34	8.88	4.36	9.88	4.58	4.54
K2O	0.26	0.26	0.23	0.32	0.22	0.20	0.15	0.12	0.12	0.25
BaO	0.00	-	-	0.01	0.04	0.08	-	0.01	-	0.05
SrO	-	-	-	-	0.04	-	-	-	-	-
Total	99.57	99.08	97.09	99.10	99.12	100.35	99.27	100.05	99.67	99.30

Structural formula

Si4+	2.75	2.81	2.78	2.42	2.42	2.81	2.42	2.88	2.43	2.42
Al3+	1.24	1.19	1.25	1.57	1.58	1.19	1.58	1.12	1.56	1.57
Fe3+	0.00	0.01	0.00	0.02	0.01	0.00	0.00	0.00	0.00	0.02
Ca2+	0.24	0.20	0.25	0.59	0.58	0.20	0.59	0.13	0.58	0.58
Na+	0.75	0.78	0.57	0.38	0.38	0.76	0.39	0.84	0.40	0.40
K+	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01
Ba2+	0.00	-	-	0.00	0.00	0.00	-	0.00	-	0.00
Sr2+	-	-	-	-	0.00	-	-	-	-	-

	73B1.3	3B1.4M	73B1.5	73B1.6	73B1.7	73B1.8	73B1.9	3B1.10	73B2.1	73B2.2
SiO2	53.74	53.14	54.89	54.40	54.02	53.49	53.21	65.59	54.17	53.42
Al2O3	28.80	29.36	28.50	28.35	28.47	29.55	29.37	21.57	28.48	29.52
Fe2O3	0.58	0.18	0.10	0.28	0.23	0.18	0.14	0.14	0.17	0.33
CaO	11.65	12.12	11.02	10.77	10.97	12.02	12.10	2.76	11.20	11.83
Na2O	4.59	4.44	5.00	5.02	4.98	4.57	4.56	9.67	5.01	4.44
K2O	0.31	0.15	0.14	0.23	0.19	0.13	0.15	0.38	0.20	0.22
BaO	-	-	-	-	-	0.07	0.01	0.02	-	0.07
SrO	-	-	-	-	-	-	-	-	-	-
Total	99.66	99.39	99.65	99.04	98.86	100.00	99.53	100.14	99.22	99.84

Structural formula

Si4+	2.44	2.42	2.48	2.48	2.46	2.42	2.42	2.88	2.46	2.42
Al3+	1.54	1.57	1.52	1.52	1.53	1.58	1.57	1.12	1.53	1.58
Fe3+	0.02	0.01	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.01
Ca2+	0.57	0.59	0.53	0.53	0.54	0.58	0.59	0.13	0.55	0.57
Na+	0.40	0.39	0.44	0.44	0.44	0.40	0.40	0.82	0.44	0.39
K+	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01
Ba2+	-	-	-	-	-	0.00	0.00	0.00	-	0.00
Sr2+	-	-	-	-	-	-	-	-	-	-

	73B2.3	73B2.4	73B2.5	73B2.6	73C.1	73C.2	73C.3	73C.4C	73C.5M	73C.6M
SiO2	63.21	63.16	63.94	54.34	63.30	60.89	61.62	25.51	62.45	62.31
Al2O3	23.26	23.13	22.88	26.50	22.66	24.17	23.52	25.51	23.18	22.59
Fe2O3	0.07	0.11	0.12	0.19	0.07	0.17	0.14	0.07	0.09	0.09
CaO	4.63	4.64	4.29	11.27	3.97	5.54	4.95	7.56	4.99	3.94
Na2O	8.64	8.64	8.96	4.86	9.16	8.07	8.49	7.18	8.97	8.97
K2O	0.12	0.27	0.17	0.19	0.19	0.20	0.28	0.12	0.15	0.25
BaO	-	0.05	-	0.04	-	0.05	0.04	0.04	0.03	0.03
SrO	-	-	-	-	-	-	-	-	-	-
Total	99.92	100.00	100.35	97.39	99.36	99.09	99.05	65.98	99.86	98.18

Structural formula

Si4+	2.79	2.79	2.81	2.52	2.81	2.73	2.76	1.83	2.77	2.80
Al3+	1.21	1.21	1.19	1.45	1.19	1.28	1.24	2.16	1.21	1.20
Fe3+	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
Ca2+	0.22	0.22	0.20	0.56	0.19	0.27	0.24	0.58	0.24	0.19
Na+	0.74	0.74	0.76	0.44	0.79	0.70	0.74	1.00	0.77	0.78
K+	0.01	0.02	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01
Ba2+	-	0.00	-	0.00	-	0.00	0.00	0.00	0.00	0.00
Sr2+	-	-	-	-	-	-	-	-	-	-

	73C.7M	73C.8C	73C.9M	73C.10	3C.11M	3C.12I	3C.13C	3C.14C	73C.15	3C.16M
SiO2	61.02	64.28	63.19	63.47	63.78	61.03	60.75	59.35	60.85	59.26
Al2O3	23.78	21.96	22.57	22.32	22.30	24.02	24.31	25.15	24.01	245.90
Fe2O3	0.10	0.17	0.07	0.12	0.13	0.12	0.14	0.13	0.11	0.09
CaO	5.22	3.19	4.02	3.65	3.38	5.86	5.81	6.91	5.52	6.71
Na2O	8.33	9.61	9.11	9.44	9.57	8.18	8.02	7.44	8.20	7.55
K2O	0.15	0.18	0.13	0.13	0.18	0.20	0.27	0.22	0.19	0.27
BaO	-	0.05	0.03	0.03	-	0.04	0.11	0.02	0.00	0.11
SrO	-	-	-	-	-	-	-	-	-	-
Total	98.60	99.43	99.14	99.16	99.34	99.45	99.41	99.22	98.88	319.89

Structural formula

Si4+	2.74	2.85	2.81	2.83	2.83	2.73	2.72	2.67	2.73	0.83
Al3+	1.26	1.15	1.18	1.17	1.17	1.27	1.28	1.33	1.27	4.08
Fe3+	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca2+	0.25	0.15	0.19	0.17	0.16	0.28	0.28	0.33	0.27	0.10
Na+	0.73	0.83	0.79	0.81	0.82	0.71	0.70	0.65	0.71	0.21
K+	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.00
Ba2+	-	0.00	0.00	0.00	-	0.00	0.00	0.00	0.00	0.00
Sr2+	-	-	-	-	-	-	-	-	-	-

	3C.17M	73B4.1	73B4.1
SiO2	60.92	53.50	51.74
Al2O3	24.14	28.03	27.81
Fe2O3	0.11	-	-
CaO	5.56	15.77	17.76
Na2O	8.17	2.70	2.70
K2O	0.14	-	-
BaO	0.03	-	-
SrO	-	-	-
Total	99.07	100.00	100.01

Structural formula

Si4+	2.73	2.43	2.37
Al3+	1.27	1.50	1.50
Fe3+	0.00	-	-
Ca2+	0.27	0.77	0.87
Na+	0.71	0.24	0.24
K+	0.01	-	-
Ba2+	0.00	-	-
Sr2+	-	-	-

Appendix VII.3: Results of EMPA (spot) analyses of pyroxenes in enclaves and their host rocks. The trachyandesite enclave in rhyolite porphyry (49B), and microsyenodiorite enclave within hedenbergite fayalite granite porphyry (67) all within the Buri ring dyke.

	49B.1	49B.2c	49B.3r	49B.4r	49B.5	49B.6c	49B.7r	67.1	67.2	67.3	67.4	67.5c	67.6r
SiO ₂	49.00	48.08	48.94	49.22	49.75	47.28	47.65	48.32	48.63	49.96	50.00	50.65	50.22
TiO ₂	0.23	0.25	0.01	0.13	0.16	0.30	0.24	0.17	0.15	0.19	0.16	0.45	0.22
Al ₂ O ₃	0.35	1.38	0.74	0.31	0.22	0.43	0.30	0.29	0.29	0.66	0.67	1.10	0.62
Cr ₂ O ₃	0.04	-	0.00	0.01	0.02	-	-	-	-	-	-	-	-
Fe ₂ O ₃ ⁺	-	-	-	1.26	0.38	2.47	1.95	1.22	0.90	-	1.20	1.21	0.27
FeO	28.78	28.13	25.64	26.72	27.14	26.97	24.89	38.24	38.54	21.30	20.47	16.26	21.45
MnO	0.80	0.73	0.68	0.78	0.76	0.67	0.74	0.94	0.85	0.49	0.50	0.48	0.47
NiO	-	-	-	0.06	-	-	-	-	-	-	-	-	-
MgO	1.82	2.21	3.55	2.18	2.20	1.83	2.20	9.40	9.46	7.08	7.91	11.05	7.74
CaO	18.23	16.34	17.77	19.44	19.61	18.44	19.71	1.51	1.54	18.16	18.46	18.14	18.06
Na ₂ O	0.55	0.46	0.34	0.57	0.56	0.48	0.51	0.01	0.01	0.27	0.26	0.27	0.30
K ₂ O	0.00	0.02	0.16	-	0.00	0.01	-	-	-	-	-	0.01	-
Total	99.79	97.60	97.85	100.63	100.86	98.87	98.20	100.10	100.37	98.11	99.63	99.62	99.35
FeO(t)	28.78	28.13	25.66	27.85	27.48	29.19	26.64	39.34	39.35	21.30	21.55	17.35	21.69
Structural formula on the basis of 6 oxygens													
IVSi	2.01	2.00	2.01	2.00	2.01	1.97	1.99	1.98	1.98	2.00	1.98	1.96	1.99
IVAl	-	0.00	-	0.00	-	0.02	0.01	0.01	0.01	-	0.02	0.04	0.01
T site	2.01	2.00	2.01	2.00	2.01	2.00	2.00	1.99	1.99	2.00	2.00	2.00	2.00
VIAl	0.02	0.07	0.04	0.01	0.01	-	0.00	-	-	0.03	0.01	0.01	0.02
Ti	0.01	0.01	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.01	0.00	0.01	0.01
Cr	0.00	-	0.00	0.00	0.00	-	-	-	-	-	-	-	-
Fe ₃ ⁺	-	-	-	0.04	0.01	0.08	0.06	0.04	0.03	-	0.04	0.04	0.01
Fe ₂ ⁺	0.99	0.98	0.88	0.90	0.92	0.94	0.86	1.30	1.31	0.71	0.68	0.52	0.71
Mn ₂ ⁺	0.03	0.03	0.02	0.03	0.03	0.02	0.03	0.03	0.03	0.02	0.02	0.02	0.02
Ni	-	-	-	-	0.00	-	-	-	-	-	-	-	-
Mg	0.11	0.14	0.22	0.13	0.13	0.11	0.14	0.57	0.57	0.42	0.47	0.64	0.46
Ca	0.80	0.73	0.78	0.84	0.85	0.82	0.88	0.07	0.07	0.78	0.78	0.75	0.77
Na	0.04	0.04	0.03	0.05	0.04	0.04	0.04	0.00	0.00	0.02	0.02	0.02	0.02
K	0.00	0.00	0.01	-	0.00	0.00	-	-	-	-	-	0.00	-
N1,M2	1.99	1.98	1.98	2.01	1.99	2.02	2.01	2.02	2.01	1.99	2.01	2.01	2.00
F/FM	0.90	0.88	0.80	0.88	0.88	0.90	0.87	0.70	0.70	0.63	0.60	0.47	0.61
Fs	51.96	53.07	46.84	48.13	48.31	50.08	46.03	67.17	67.17	37.24	35.16	27.47	36.74
Wo	42.17	39.50	41.58	44.86	44.71	43.87	46.72	3.40	3.44	40.69	40.62	39.26	39.63
En	5.87	7.43	11.57	7.01	6.98	6.05	7.25	29.43	29.39	22.07	24.22	33.27	23.63

	73B1.7	73C.1c	73C.2r	73C.3r	73C.4	73C.5	73C.6	73C.7	73C.8	73C.9
SiO2	46.94	49.02	46.85	45.05	49.24	43.25	43.30	46.28	46.29	44.20
TiO2	0.71	0.94	1.09	1.24	0.78	0.53	0.31	1.00	0.99	0.87
Al2O3	4.60	3.88	5.41	6.56	3.72	7.69	8.10	5.51	5.83	7.07
FeO	22.21	20.44	20.34	20.76	20.41	21.31	21.48	20.58	20.85	21.04
MnO	0.64	0.42	0.39	0.44	0.48	0.39	0.36	0.46	0.44	0.43
MgO	8.42	10.72	9.73	8.95	10.90	8.03	7.92	9.05	8.88	8.40
ZnO	-	-	-	-	-	-	-	-	-	-
CaO	10.64	10.20	10.68	10.87	10.05	10.81	11.20	10.27	10.78	10.25
Na2O	1.15	1.30	1.50	1.60	1.27	1.55	1.25	1.41	1.24	1.42
K2O	0.78	0.39	0.63	0.93	0.41	1.22	1.18	0.72	0.82	1.04
H2O#	1.17	1.30	1.27	1.34	1.27	1.18	1.30	1.34	1.29	1.25
F	1.53	1.40	1.36	1.14	1.45	1.35	1.11	1.12	1.25	1.26
Cl	0.12	0.09	0.18	0.23	0.10	0.26	0.29	0.21	0.20	0.22
O=F+Cl	-0.70	-0.60	-0.60	-0.50	-0.60	-0.60	-0.50	-0.50	-0.60	-0.60
Total	98.25	99.48	98.80	98.58	99.45	96.94	97.25	97.44	98.29	96.86

Structural formula on the basis of 24(O,OH,F,Cl)

IVSi	7.31	7.41	7.18	6.98	7.44	6.87	6.85	7.21	7.16	6.98
IVAl	0.69	0.59	0.82	1.02	0.56	1.13	1.15	0.79	0.84	1.02
IVTi	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	0.15	0.10	0.16	0.18	0.10	0.30	0.36	0.22	0.22	0.29
Ti	0.08	0.11	0.13	0.14	0.09	0.06	0.04	0.12	0.12	0.10
Mg	1.95	2.42	2.22	2.07	2.45	1.90	1.87	2.10	2.05	1.98
Fe2+	2.81	2.38	2.49	2.61	2.35	2.73	2.74	2.57	2.61	2.63
Zn	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	-	-	-	-	-	-
Ca	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	-	-	-	-	-	-	-	-	-	-
Fe2+	0.08	0.21	0.11	0.08	0.23	0.10	0.10	0.11	0.08	0.15
Zn	-	-	-	-	-	-	-	-	-	-
Mn	0.08	0.05	0.05	0.06	0.06	0.05	0.05	0.06	0.06	0.06
Ca	1.77	1.65	1.75	1.80	1.63	1.84	1.85	1.71	1.79	1.73
Na	0.06	0.09	0.08	0.06	0.09	0.01	-	0.11	0.07	0.06
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	-	-	-	0.05	-	-	-
Na	0.29	0.29	0.36	0.42	0.29	0.47	0.38	0.31	0.30	0.37
K	0.16	0.08	0.12	0.18	0.08	0.25	0.24	0.14	0.16	0.21
A site	0.44	0.37	0.49	0.61	0.36	0.71	0.67	0.46	0.46	0.59
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	1.21	1.31	1.29	1.38	1.28	1.25	1.37	1.39	1.34	1.31
F	0.76	0.67	0.66	0.56	0.69	0.68	0.55	0.55	0.61	0.63
Cl	0.03	0.02	0.05	0.06	0.02	0.07	0.08	0.06	0.05	0.06
F/FM	0.60	0.52	0.54	0.57	0.51	0.60	0.60	0.56	0.57	0.58

Appendix VII.4: Results of EMPA (spot) analyses of amphibole in enclaves (+host rock). Trachyandesite enclave in rhyolite porphyry (49A, & 49B), microsyenodiorite enclave in hedenbergite fayalite granite porphyry (67), trachybasalt (73B), andesite (73B1), and dacitic enclave (73C).

	49A.1	49A.2	49B.1	49B.2	49B.3	49B.4	49B.5	67.1	73B.1	73B.2	73B.3	73B.4	73B.5	73B.6	73B1.1	73B1.2	73B1.3	73B1.4	73B1.5	73B1.6
SiO ₂	45.54	46.36	43.68	45.24	43.63	46.09	43.33	47.73	43.99	43.31	42.61	43.58	43.88	43.98	47.02	47.65	43.89	43.96	48.13	46.03
TiO ₂	1.26	1.14	1.27	1.18	0.89	1.27	1.44	0.11	1.36	1.44	1.54	1.63	1.46	1.51	0.64	0.57	0.92	0.89	0.54	0.84
Al ₂ O ₃	4.59	4.32	4.63	4.41	4.07	4.32	4.71	1.77	6.96	7.32	7.95	6.90	7.10	6.88	5.10	4.02	6.83	6.89	3.98	5.12
FeO	24.73	25.78	29.70	24.37	24.74	26.08	30.29	34.51	24.31	23.38	23.78	23.51	23.32	24.79	23.36	22.51	23.92	24.01	22.70	22.53
MnO	0.37	0.43	0.57	0.56	0.55	0.47	0.57	0.70	0.47	0.50	0.49	0.44	0.47	0.46	0.62	0.55	0.56	0.56	0.64	0.64
MgO	7.81	7.28	4.23	7.01	6.39	6.72	2.66	1.02	6.99	6.57	6.37	6.68	6.62	6.76	8.08	8.58	7.05	7.13	8.93	8.08
ZnO	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
CaO	9.52	9.42	9.48	9.36	11.08	9.24	9.68	10.95	10.56	10.48	10.65	10.06	10.32	10.37	10.81	10.44	10.72	10.78	10.67	10.34
Na ₂ O	2.47	2.30	2.30	2.46	2.35	2.73	2.21	0.40	1.48	1.57	1.59	1.53	1.58	1.52	1.08	0.98	1.28	1.37	1.03	1.39
K ₂ O	0.93	0.91	1.09	1.01	0.90	0.99	0.98	0.24	0.96	1.08	1.12	0.95	0.99	0.93	0.83	0.67	1.13	1.17	0.69	0.81
H ₂ O	0.89	1.25	1.18	1.05	1.09	0.95	1.37	1.77	1.66	1.53	1.60	1.52	1.58	1.58	1.32	1.29	1.29	1.16	1.23	1.20
F	2.02	1.30	1.24	1.69	1.53	1.97	0.86	0.12	0.37	0.56	0.41	0.60	0.50	0.55	1.25	1.28	1.17	1.46	1.48	1.43
Cl	0.26	0.23	0.34	0.13	0.12	0.14	0.18	0.08	0.31	0.35	0.40	0.32	0.31	0.30	0.12	0.12	0.21	0.20	0.10	0.11
O=F+Cl	-0.90	-0.60	-0.60	-0.70	-0.70	-0.90	-0.40	-0.07	-0.20	-0.30	-0.30	-0.30	-0.30	-0.30	-0.60	-0.60	-0.50	-0.70	-0.60	-0.60
Total	99.47	100.11	99.11	97.73	96.65	100.12	97.88	99.34	99.18	97.78	98.24	97.40	97.86	99.32	99.69	98.09	98.43	98.95	99.46	97.91
Structural formula on the basis of 24(O,OH,F,Cl)																				
IVSi	7.13	7.22	7.07	7.20	7.11	7.20	7.11	7.73	6.89	6.87	6.76	6.93	6.94	6.89	7.24	7.41	6.93	6.91	7.39	7.21
IVAl	0.85	0.78	0.88	0.80	0.78	0.80	0.89	0.27	1.11	1.13	1.24	1.07	1.06	1.11	0.76	0.59	1.07	1.09	0.61	0.79
IVTi	0.02	-	0.05	-	0.11	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	-	0.01	-	0.03	-	0.00	0.03	0.07	0.18	0.24	0.24	0.22	0.26	0.16	0.17	0.15	0.20	0.19	0.11	0.16
Ti	0.13	0.13	0.10	0.14	-	0.15	0.18	0.01	0.16	0.17	0.18	0.20	0.17	0.18	0.07	0.07	0.11	0.11	0.06	0.10
Mg	1.82	1.69	1.02	1.66	1.55	1.57	0.65	0.25	1.63	1.55	1.51	1.58	1.56	1.58	1.86	1.99	1.66	1.67	2.04	1.89
Fe ²⁺	3.05	3.17	3.88	3.16	3.37	3.28	4.14	4.67	3.03	3.03	3.07	3.00	3.01	3.08	2.90	2.80	3.03	3.03	2.79	2.85
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	-	-	-	-	0.08	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Ca	-	-	-	-	0.00	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
M1,2,3	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00	5.00
Mg	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Fe ²⁺	0.19	0.19	0.14	0.08	-	0.13	0.01	0.00	0.15	0.07	0.08	0.12	0.08	0.17	0.11	0.13	0.13	0.12	0.13	0.10
Zn	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	0.05	0.06	0.08	0.08	-	0.06	0.08	0.10	0.06	0.07	0.07	0.06	0.06	0.06	0.08	0.07	0.07	0.08	0.08	0.09
Ca	1.60	1.57	1.64	1.60	1.93	1.55	1.70	1.90	1.77	1.78	1.81	1.71	1.75	1.74	1.78	1.74	1.80	1.76	1.76	1.74
Na	0.17	0.18	0.14	0.24	0.07	0.26	0.20	0.00	0.01	0.08	0.04	0.10	0.11	0.03	0.02	0.06	-	-	0.03	0.08
M4 site	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
Ca	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.02	0.02	-	-
Na	0.58	0.51	0.59	0.52	0.68	0.56	0.50	0.13	0.44	0.40	0.45	0.37	0.37	0.43	0.30	0.24	0.39	0.42	0.27	0.34
K	0.19	0.18	0.22	0.20	0.19	0.20	0.21	0.05	0.19	0.22	0.23	0.19	0.20	0.19	0.16	0.13	0.23	0.24	0.14	0.16
A site	0.77	0.69	0.81	0.72	0.86	0.76	0.71	0.18	0.63	0.62	0.67	0.56	0.57	0.62	0.46	0.37	0.63	0.67	0.41	0.51
O	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00
OH	0.93	1.30	1.27	1.11	1.18	0.99	1.50	1.92	1.74	1.62	1.69	1.61	1.67	1.65	1.36	1.34	1.36	1.22	1.26	1.26
F	1.00	0.64	0.64	0.85	0.79	0.97	0.45	0.06	0.18	0.28	0.20	0.30	0.25	0.27	0.61	0.63	0.58	0.73	0.72	0.71
Cl	0.07	0.06	0.09	0.04	0.03	0.04	0.05	0.02	0.08	0.10	0.11	0.09	0.08	0.08	0.03	0.03	0.06	0.05	0.03	0.03
F/FM	0.64	0.67	0.80	0.66	0.68	0.69	0.86	0.95	0.66	0.67	0.68	0.66	0.66	0.67	0.62	0.60	0.66	0.65	0.59	0.61

Appendix VII.5: Results of EMPA (spot) analyses of mica in enclaves and their host rocks. Sample 73B (trachybasalt), 73B1 and 73B2 (andesite), 73C (dacite) are all enclaves in biotite granite of the Panshanu center.

	73B.1	73B.2	73B.3	73B.4	73B.5	73B.6	73B1.1	73B1.2	73B1.3	73B1.4	73B1.5	73B1.6	73B1.7	73B1.8	73B2.1	73B2.2	73B2.3	73B2.4	73B2.5	73B2.6
SiO ₂	38.17	38.35	38.87	38.08	38.29	38.59	36.98	38.30	37.36	35.83	37.30	38.40	36.62	36.52	36.64	37.27	37.88	37.96	38.51	37.50
TiO ₂	1.93	1.80	1.91	1.89	1.91	1.58	3.24	3.85	3.78	3.31	3.03	2.72	3.26	2.95	4.18	3.42	3.46	3.57	3.39	3.46
Al ₂ O ₃	12.57	12.44	12.46	12.37	12.25	12.69	12.85	12.75	12.59	13.35	12.41	12.60	12.69	13.26	13.32	13.04	12.65	12.89	12.98	13.20
FeO	23.30	23.45	23.59	23.19	23.09	23.52	22.15	21.69	22.76	22.19	23.10	21.71	27.66	27.32	26.67	25.64	22.91	12.89	23.31	22.34
MnO	0.28	0.27	0.24	0.25	0.31	0.31	0.26	0.26	0.21	0.24	0.19	0.14	0.26	0.37	0.29	0.36	0.20	0.15	0.32	0.13
MgO	10.80	10.67	10.60	10.62	10.58	10.27	10.18	10.14	10.24	10.01	9.97	11.22	6.38	6.89	6.80	8.32	10.67	10.30	9.81	9.52
CaO	0.03	0.02	0.07	0.22	-	0.03	-	-	0.03	-	0.02	0.03	-	0.01	0.01	-	0.03	0.01	0.01	0.02
Na ₂ O	0.05	0.09	0.11	0.12	0.08	0.11	0.07	0.14	0.09	0.08	0.09	0.14	0.05	0.05	0.09	0.06	0.09	0.06	0.06	0.09
K ₂ O	9.95	9.57	10.02	9.79	9.85	10.06	10.07	9.97	10.13	10.06	9.90	10.19	9.82	9.69	9.89	9.88	9.86	9.97	10.11	10.25
H ₂ O ⁺	2.38	2.40	2.54	2.36	2.45	2.49	2.87	3.14	2.96	3.00	2.73	2.75	2.89	3.03	3.32	3.07	3.10	3.23	3.08	3.20
F	2.96	2.91	2.63	2.99	2.81	2.70	1.88	1.41	1.72	1.59	2.16	2.23	1.55	1.35	0.77	1.27	1.43	1.38	1.49	1.22
Cl	0.28	0.29	0.30	0.24	0.27	0.32	0.32	0.32	0.31	0.27	0.32	0.29	0.52	0.42	0.42	0.54	0.42	0.47	0.45	0.42
O=F+Cl	-1.00	-1.00	-1.00	-1.00	-1.00	-1.00	-0.90	-0.70	-0.80	-0.70	-1.00	-1.00	-0.80	-0.70	-0.40	-0.70	-0.30	-0.30	-0.40	-0.30
Total	101.37	100.96	102.17	100.82	100.65	101.46	99.99	101.28	101.38	99.20	100.23	101.44	100.93	101.20	101.98	102.22	101.95	92.09	102.71	100.67
Structural formula on the basis of 24(O,OH,F,Cl)																				
IVSi	5.84	5.89	5.90	5.86	5.90	5.90	5.73	5.82	5.72	5.61	5.78	5.83	5.77	5.71	5.67	5.72	5.75	6.11	5.80	5.76
IVAl	2.16	2.11	2.10	2.14	2.10	2.10	2.27	2.18	2.27	2.39	2.22	2.17	2.23	2.29	2.33	2.28	2.25	1.89	2.20	2.24
IVTi	-	-	-	-	-	-	-	-	0.01	-	-	-	-	-	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
VAl	0.11	0.13	0.13	0.11	0.12	0.19	0.08	0.10	-	0.08	0.05	0.09	0.12	0.16	0.10	0.08	0.01	0.55	0.11	0.15
VITi	0.22	0.21	0.22	0.22	0.22	0.18	0.38	0.44	0.43	0.39	0.35	0.31	0.39	0.35	0.49	0.40	0.39	0.43	0.38	0.40
Fe ²⁺	2.98	3.01	2.99	2.99	2.97	3.01	2.87	2.75	2.91	2.91	3.00	2.76	3.64	3.57	3.45	3.29	2.91	1.73	2.94	2.87
Mn ²⁺	0.04	0.03	0.03	0.03	0.04	0.04	0.03	0.03	0.03	0.03	0.02	0.02	0.03	0.05	0.04	0.05	0.03	0.02	0.04	0.02
Mg	2.47	2.44	2.40	2.44	2.43	2.34	2.35	2.29	2.34	2.34	2.30	2.54	1.50	1.61	1.57	1.90	2.41	2.47	2.20	2.18
Y site	5.82	5.83	5.77	5.78	5.79	5.77	5.71	5.62	5.70	5.75	5.73	5.72	5.68	5.74	5.64	5.72	5.75	5.21	5.68	5.62
Ca	0.00	0.00	0.01	0.04	-	0.01	-	-	0.00	-	0.00	0.01	-	0.00	0.00	-	0.00	0.00	0.00	0.00
Na	0.01	0.03	0.03	0.03	0.02	0.03	0.02	0.04	0.03	0.02	0.03	0.04	0.01	0.02	0.03	0.02	0.03	0.02	0.02	0.03
K	1.94	1.87	1.94	1.92	1.94	1.96	1.99	1.93	1.98	2.01	1.96	1.98	1.97	1.93	1.95	1.93	1.91	2.05	1.94	2.01
X site	1.96	1.91	1.98	1.99	1.96	2.00	2.01	1.97	2.01	2.03	1.99	2.02	1.99	1.95	1.98	1.95	1.94	2.07	1.96	2.04
O	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
OH	2.49	2.51	2.66	2.48	2.56	2.61	2.99	3.24	3.09	3.14	2.86	2.85	3.09	3.22	3.51	3.24	3.22	3.21	3.20	3.32
F	1.43	1.41	1.26	1.46	1.37	1.31	0.92	0.68	0.83	0.79	1.06	1.07	0.77	0.67	0.38	0.62	0.69	0.70	0.71	0.59
Cl	0.07	0.07	0.08	0.06	0.07	0.08	0.08	0.08	0.08	0.07	0.08	0.08	0.14	0.11	0.11	0.14	0.09	0.09	0.09	0.09
X(F)	0.36	0.35	0.32	0.36	0.34	0.33	0.23	0.17	0.21	0.20	0.26	0.27	0.19	0.17	0.09	0.15	0.17	0.17	0.18	0.15
X(Cl)	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.03	0.03	0.04	0.02	0.02	0.02	0.02
X(F/Cl)	19.88	18.99	16.59	23.38	19.64	15.94	11.13	8.32	10.47	10.84	12.52	14.20	5.55	6.03	3.38	4.39	7.59	7.37	7.94	6.92
F/FH	0.55	0.55	0.56	0.55	0.55	0.56	0.55	0.55	0.55	0.55	0.57	0.52	0.71	0.69	0.69	0.63	0.55	0.41	0.57	0.57
Phl	42.37	41.90	41.56	42.17	41.98	40.60	41.16	40.83	40.97	40.67	40.20	44.44	26.36	28.01	27.80	33.27	41.96	40.44	38.82	38.81
Srp	38.62	46.84	44.98	37.08	42.36	59.40	27.85	34.71	-	27.85	19.14	29.44	52.49	69.05	42.14	33.68	4.07	32.35	40.43	56.05
Ann	19.01	11.27	13.46	20.76	15.67	0.00	30.99	24.46	59.03	31.48	40.66	26.12	21.15	2.94	30.07	33.04	53.97	27.21	20.75	5.14

	73B2.7	73B2.8	73C.1	73C.2	73C.3	73C.4
SiO2	38.24	38.21	36.38	36.92	37.05	37.37
TiO2	3.68	3.69	3.76	4.14	3.66	3.71
Al2O3	12.96	12.87	12.84	12.65	12.91	13.22
FeO	22.23	21.78	25.46	26.28	26.47	26.71
MnO	0.15	0.16	0.20	0.20	0.19	0.13
MgO	10.32	10.11	7.30	7.35	7.57	7.59
CaO	0.02	0.02	0.04	0.02	0.03	0.01
Na2O	0.06	0.04	0.02	0.10	0.09	0.06
K2O	10.15	10.03	9.79	9.85	9.77	9.63
H2Ow	3.17	3.26	3.46	3.42	3.39	3.50
F	1.28	1.13	0.50	0.55	0.63	0.44
Cl	0.35	0.35	0.35	0.33	0.39	0.36
O=F+Cl	-0.70	-0.70	-0.70	-0.60	-0.60	-0.60
Total	102.03	101.10	99.85	101.61	101.83	102.51

Structural formula on the basis of 24(O,OH,F,Cl)						
IVSi	5.78	5.81	5.73	5.72	5.73	5.72
IVAl	2.22	2.19	2.27	2.28	2.27	2.28
IVTi	-	-	-	-	-	-
T site	8.00	8.00	8.00	8.00	8.00	8.00
VIAI	0.09	0.12	0.11	0.03	0.08	0.11
VITi	0.42	0.42	0.44	0.48	0.43	0.43
Fe2+	2.81	2.77	3.35	3.41	3.42	3.42
Mn2+	0.02	0.02	0.03	0.03	0.02	0.02
Mg	2.33	2.29	1.71	1.70	1.74	1.73
O site	5.66	5.63	5.64	5.65	5.69	5.71
Ca	0.00	0.00	0.01	0.00	0.00	0.00
Na	0.02	0.01	0.01	0.03	0.03	0.02
K	1.96	1.95	1.97	1.95	1.93	1.88
A site	1.98	1.96	1.98	1.98	1.96	1.90
O	20.00	20.00	20.00	20.00	20.00	20.00
OH	3.29	3.36	3.64	3.61	3.58	3.68
F	0.61	0.54	0.25	0.27	0.31	0.21
Cl	0.10	0.09	0.11	0.12	0.12	0.11
X(F)	0.15	0.14	0.06	0.07	0.08	0.05
X(Cl)	0.02	0.02	0.03	0.03	0.03	0.03
F/Cl	6.20	5.86	2.24	2.17	2.62	1.99
F/FH	0.55	0.55	0.66	0.67	0.66	0.66
Phl	41.10	40.74	30.33	30.07	30.65	30.37
Srp	30.42	43.12	45.39	14.25	32.12	45.21
Ann	28.48	16.14	24.28	55.67	37.23	24.42

Appendix VII.6: Calculated proportions of mixing in enclaves assuming two end-member mixing. Bs = basalt, Tr = trachyte, Rh = rhyolite, DM and Dc are dacitic enclaves.

Sample	49Ma (Tr:Bs)	87Ma (Tr:Bs)	67Ma (Tr:Bs)	73a (Tr:Bs)	73B2a (Rh:Bs)	73B2a (DM:Bs)	73B2a (Dc:Bs)
SiO2	0.50:0.50	0.66:0.34	0.55:0.45	0.13:0.87	0.45:0.55	0.68:0.32	0.71:0.29
TiO2	0.52:0.48	0.60:0.40	0.69:0.31	0.37:0.63	0.54:0.46	0.74:0.26	0.81:0.19
Al2O3	2.00:-1.00	2.45:-1.45	-0.49:1.49	-1.50:2.50	-0.07:1.07	-0.16:1.16	-0.17:1.17
Fe2O3	-3.68:4.68	-3.03:4.03	-2.62:3.62	-5.92:6.92	-2.01:3.01	-2.82:3.82	1.44:-0.44
FeO	1.78:-0.78	1.78:-0.78	1.78:-0.78	1.54:-0.54	1.00:0.00	1.63:-0.63	1.00:0.00
MnO	0.25:0.75	1.50:-0.50	-0.50:1.50	1.08:-0.08	0.63:0.38	0.83:0.17	0.87:0.13
MgO	0.67:0.33	0.73:0.27	0.82:0.18	0.20:0.80	0.41:0.59	0.51:0.49	0.56:0.44
CaO	0.77:0.23	0.72:0.28	0.81:0.19	0.29:0.71	0.34:0.66	0.49:0.51	0.52:0.48
Na2O	0.99:0.01	1.26:-0.26	1.45:-0.45	0.43:0.57	1.25:-0.25	1.95:-0.95	0.73:0.27
K2O	0.56:0.44	0.48:0.52	0.65:0.35	0.48:0.52	0.10:0.90	0.11:0.89	0.35:0.65
P2O5	-1.44:2.44	-1.15:2.13	-0.04:1.04	0.05:0.95	0.34:0.66	0.80:0.20	1.23:-0.23
Cr	1.00:0.00	1.00:0.00	1.00:0.00	-0.34:1.34	-1.75:2.75	0.75:0.25	0.55:0.45
Ni	1.00:0.00	1.00:0.00	1.00:0.00	0.35:0.65	-0.55:1.55	0.87:0.13	0.69:0.31
V	0.79:0.21	0.85:0.15	0.72:0.28	0.12:0.88	0.45:0.55	0.55:0.45	0.67:0.33
Pb	1.68:-0.68	1.16:-0.16	5.16:-4.16	0.23:0.77	0.08:0.93	0.17:0.83	0.20:0.80
S	1.00:0.00	1.00:0.00	1.00:0.00	1.00:0.00	1.01:-0.01	1.01:-0.01	1.00:0.00
Rb	-0.86:1.86	-0.20:1.20	-0.14:1.14	-2.59:3.59	-0.23:1.23	0.95:0.05	-0.32:1.32
Ba	2.99:-1.99	2.18:-1.18	4.40:-3.40	0.02:0.98	-0.06:1.06	0.02:0.98	0.07:0.93
Sr	0.69:0.31	0.68:0.32	0.79:0.21	0.80:0.20	0.42:0.58	0.69:0.31	0.54:0.46
Ta	n.d.:n.d.	n.d.:n.d.	n.d.:n.d.	-0.58:1.58	0.01:0.99	0.03:0.97	-0.03:1.03
Nb	1.28:-0.28	1.37:-0.37	0.48:0.52	0.49:0.51	-0.11:1.11	-0.40:1.40	-0.21:1.21
Hf	1.18:-0.18	1.18:-0.18	1.13:-0.13	-0.05:1.05	0.16:0.84	-0.08:1.08	0.06:0.94
Zr	1.12:-0.12	1.09:-0.09	1.20:-0.20	-0.04:1.04	0.27:0.73	-0.11:1.11	-0.14:1.14
Y	1.79:-0.79	1.55:-0.55	0.73:0.27	1.08:-0.08	0.27:0.73	0.74:0.26	0.21:0.79
Th	1.33:-0.33	4.00:-3.00	1.42:-0.42	4.50:-3.50	0.13:0.87	0.78:0.22	0.87:0.13
U	0.75:0.25	0.67:0.33	1.00:0.00	0.44:0.56	-3.50:4.50	2.33:-1.33	0.64:0.36
La	0.38:0.62	0.37:0.63	0.34:0.66	0.08:0.92	0.26:0.74	0.16:0.84	-0.43:1.43
Ce	0.47:0.53	0.48:0.52	0.40:0.59	0.11:0.89	0.20:0.80	0.17:0.83	-0.28:1.28
Pr	n.d.:n.d.	n.d.:n.d.	n.d.:n.d.	-0.51:1.51	0.17:0.83	0.12:0.88	-0.16:1.16
Nd	0.37:0.63	0.35:0.65	0.30:0.70	0.04:0.96	0.12:0.88	0.06:0.94	-0.06:1.06
Sm	0.39:0.61	0.38:0.62	0.27:0.73	0.03:0.97	0.00:1.00	0.00:1.00	0.00:1.00
Eu	0.61:0.39	0.60:0.40	0.54:0.46	-0.16:1.16	0.51:0.49	3.18:-2.18	0.41:0.59
Gd	0.50:0.50	0.48:0.52	0.33:0.67	0.06:0.94	-0.30:1.29	-0.16:1.16	0.05:0.94
Tb	n.d.:n.d.	n.d.:n.d.	n.d.:n.d.	-0.62:1.62	-0.04:1.04	-0.04:1.04	0.02:0.98
Dy	4.47:-3.47	4.33:-3.33	2.52:-1.52	-7.67:8.67	0.04:0.96	0.08:0.92	-0.03:1.03
Ho	0.85:0.15	0.78:0.22	0.47:0.53	0.34:0.66	0.11:0.89	0.31:0.69	-0.13:1.13
Er	n.d.:n.d.	n.d.:n.d.	n.d.:n.d.	-1.73:2.73	0.14:0.86	0.42:0.58	-0.20:1.20
Tm	0.99:0.01	0.88:0.12	0.55:0.45	0.74:0.26	0.16:0.84	0.62:0.38	-0.33:1.33
Yb	0.82:0.18	0.74:0.26	0.47:0.53	0.65:0.35	0.19:0.81	0.88:0.12	-0.44:1.44
Lu	0.63:0.37	0.58:0.42	0.40:0.60	0.53:0.47	0.13:0.87	0.54:0.46	-0.28:1.28

Appendix VIII.1: Melt compositions generated by partial melting model: a) melt compositions generated by (i) modal, and (ii) nonmodal melting of non-enriched trial source-rocks (Data plotted on Figs. 8.5a - d, and 8.7a - d).
b) Enrichments of trial source-rock compositions for 20 and 60% (i) modal, and (ii) nonmodal melting to yield target melt composition (Data plotted on Figs. 8.6a - d, and 8.8a - d).

a: Melt compositions

	MODAL MELTING						NONMODAL MELTING					
	AV. L. CRUST		GRANULITE		GRANODIORITE		AV. L. CRUST		GRANULITE		GRANODIORITE	
	ALC20MM	ALC60MM	Gr120MM	Gr160MM	Grd20MM	Grd60MM	ALC20MM	ALC60MM	Gr120MM	Gr160MM	Grd20MM	Grd60MM
Rb	7.98	6.37	62.61	60.22	228.00	194.78	8.34	7.10	65.63	69.43	236.37	214.21
Sr	97.62	133.07	70.23	112.68	99.89	157.49	79.68	70.34	63.97	76.63	91.92	111.66
Ba	179.62	163.47	416.09	447.06	511.79	662.24	181.91	169.34	430.17	499.79	534.78	794.92
Zr	76.82	73.25	193.86	214.87	160.23	174.27	125.22	-693.07	469.42	-225.70	413.96	-174.43
La	13.97	0.31	39.54	34.75	47.57	40.98	16.94	22.94	81.84	95.80	103.39	-103.63
Ce	27.51	25.05	95.04	81.93	101.43	85.57	31.80	39.68	177.95	-399.56	203.30	-318.97
Nd	13.20	12.94	54.09	43.23	50.33	42.65	14.16	16.18	87.59	521.74	79.81	700.86
Sm	3.05	3.11	7.31	8.36	9.01	7.49	3.10	3.27	24.34	26.51	11.68	17.40
Eu	1.02	1.09	3.29	3.66	1.58	1.66	0.91	0.80	3.10	3.05	1.60	1.74
Tb	0.48	0.53	2.76	1.65	1.29	1.11	0.47	0.50	3.16	2.15	1.48	1.66
Yb	2.14	2.17	9.04	5.28	5.21	3.98	2.10	2.06	10.17	6.57	5.95	5.54
Lu	0.32	0.31	2.24	1.28	1.27	0.86	0.32	0.29	2.58	1.67	1.50	1.23

b: Enrichments required for: i) modal, and b) nonmodal melting of selected source-rocks to produce target melt

	MODAL MELTING						NONMODAL MELTING					
	L. CRUST		GRANULITE		GRANODIORITE		L. CRUST		GRANULITE		GRANODIORITE	
	ENLC20MM	ENLC60MM	ENGr120MM	ENGr160MM	ENGrd20MM	ENGrd60MM	ENLC20MM	ENLC60MM	ENGr120MM	ENGr160MM	ENGrd20MM	ENGrd60MM
Rb	100.30	125.70	139.90	145.50	(170.00)	(170.00)	96.00	113.00	139.90	126.50	(170.00)	(170.00)
Sr	(230.00)	(230.00)	(285.00)	(285.00)	(372.00)	(372.00)	(230.00)	(230.00)	(285.00)	(285.00)	(372.00)	(372.00)
Ba	281.50	309.30	(483.00)	(483.00)	(938.00)	(938.00)	273.00	298.70	(483.00)	(483)	(938.00)	(938.00)
Zr	452.00	474.00	616.75	556.50	591.40	544.00	277.40	(P=2.25)	(241)	(241)	229.00	(P=2.43)
La	106.30	120.70	105.90	120.60	102.20	118.70	87.80	64.80	51.20	(31)	47.10	(P=1.72)
Ce	229.90	252.50	208.40	241.80	200.80	238.00	199.00	159.60	110.20	(72)	100.20	(P=1.66)
Nd	115.50	117.80	80.00	105.00	88.30	104.30	107.80	94.50	49.50	(36)	55.80	(P=1.60)
Sm	28.10	27.60	36.20	31.60	19.25	23.30	27.80	26.25	10.90	10.00	15.00	10.05
Eu	1.67	1.56	(4.13)	(4.13)	(1.75)	(1.75)	1.86	2.13	(4.13)	(4.13)	(P=0.66)	(P=0.85)
Tb	4.78	4.33	1.66	2.77	2.92	3.40	4.87	4.60	1.45	2.13	2.55	2.28
Yb	10.72	10.57	4.30	7.36	6.44	8.43	10.91	11.12	3.82	5.92	5.65	6.06
Lu	1.32	1.40	(0.90)	0.997	0.75	1.11	1.35	1.42	(0.90)	(0.90)	(P=0.73)	0.75

N.B.: Concentrations given in brackets are above those required to generate that target composition. Even if the P value is zero, the amount of these elements is still above that of the target melt. The P values in brackets indicate instances where the target melt content of a given element can be arrived at by adjusting p_i values (i.e., the fraction of minerals going into the melt) at this P-value, without enrichment of the element. In this case, the element content does not require further enrichment at the given percentage of melting.

Appendix VIII.2: a) p_i values compiled by Rämö (1991), b) calculated D and adjusted P values (i.e., bulk melt distribution coefficients) for nonmodal melting, and c) inferred "appropriate" p_i values that will yield target melt composition from selected source rock compositions.

MM(ALC)	0.06	0.13	0	0	0.16	0.03	0	0.24	0.005	0.005	0.0005	0.001	0.0005	0.00
MM(Gr1)	0.00	0.20	0.31	0.00	0.04	0.00	0.00	0.45	0.001	0.00	0.001	0.001	0.001	0.00
MM(Grd)	0.06	0.13	0.20	0.25	0.00	0.00	0.00	0.36	0.00	0.00	0.001	0.001	0.001	0.00

Calculated bulk D values (DALC, DGr1, & DGrd), and adjusted P values for nonmodal melting of selected source rocks.

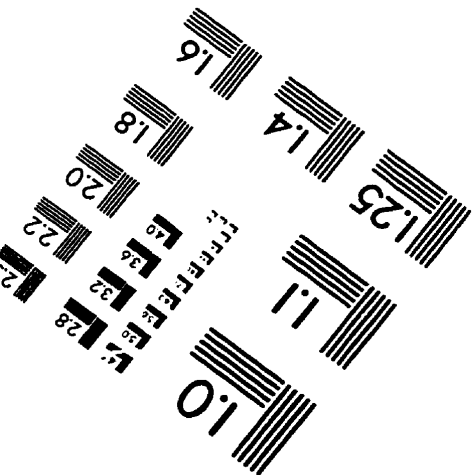
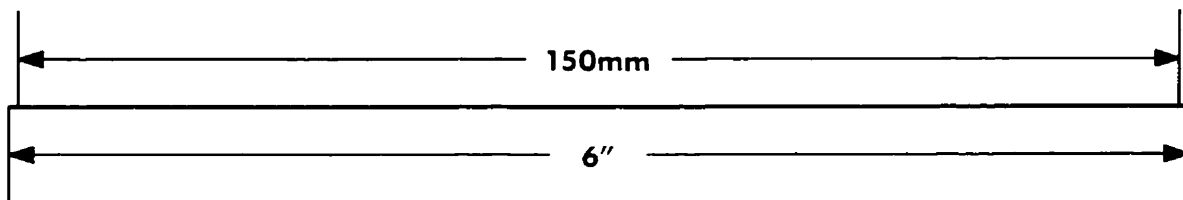
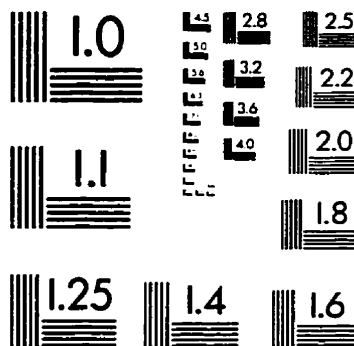
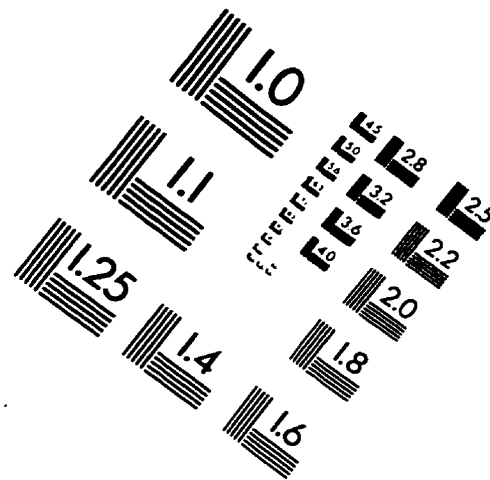
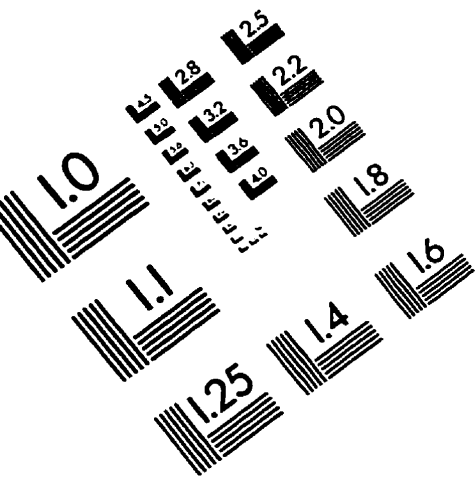
	L/CRUST				GRANULITE				GRANODIORITE			
	DALC	PALC1	PALC2	PALC3	DGr1	PGr11	PGr12	PGr13	DGrd	PGrd1	PGrd2	PGrd3
Rb	0.580	0.723	2.146	1.385	0.908	1.121	3.263	1.875	0.682	0.814	0.000	0.265
Sr	2.695	0.042	2.750	1.582	4.823	2.84	1.391	0.615	4.405	2.789	0.000	0.000
Ba	0.794	0.847	0.000	0.000	1.201	1.391	2.840	0.000	2.041	2.435	0.000	0.000
Zr	0.869	2.651	4.740	2.247	1.304	4.953	5.092	2.364	1.240	4.891	5.275	2.425
La	0.592	1.424	4.265	2.088	0.730	2.756	3.502	1.835	0.696	2.739	3.150	1.716
Ce	0.639	1.359	4.557	2.186	0.697	2.462	3.178	1.726	0.662	2.490	2.967	1.655
Nd	0.801	1.280	5.236	2.413	0.582	1.855	2.410	1.470	0.669	2.027	2.805	1.602
Sm	0.911	1.131	5.638	2.555	0.417	1.082	1.300	1.095	0.639	1.451	3.020	1.674
Eu	1.135	0.529	2.900	1.631	1.321	0.945	0.000	0.000	1.138	1.216	0.680	0.886
Tb	1.160	1.177	6.690	2.897	0.285	0.559	0.905	0.969	0.691	1.175	3.205	1.735
Yb	0.954	0.948	5.125	2.375	0.266	0.497	0.541	0.847	0.522	0.902	2.065	1.355
Lu	0.800	0.786	0.000	0.000	0.253	0.521	0.000	0.404	0.387	0.762	0.762	0.900

Average lower crust: bulk D of average lower crust (DALC), bulk melt distribution coefficient value based on melt modes from literature (PALC1), and adjusted to yield target melt at 20% (PALC2) and 60% (PALC3) nonmodal melting. **Granulite:** bulk D of granulite (DGr1), bulk melt distribution coefficients from literature (PGr11), and adjusted to yield target melt at 20% (PGr12) and 60% (PGr13) nonmodal melting. **Granodiorite:** bulk D of granodiorite (DGrd), bulk melt distribution coefficients from literature (PGrd1), and adjusted to yield target melt at 20% (PGrd2) and 60% (PGrd3) nonmodal melting].

Adjusted p_i values (NLC, NGr1, and NGrd in parentheses give the starting p_i values determined by Rämö (1991)

	(NLC)	ALC20	ALC60	(NGr1)	AGr120	AGr160	(NGrd)	AGrd20	AGrd60
Hbl	0.060	1.904	0.882	0.000	0.202	0.171	0.060	0.533	0.176
Bt	0.130	0.426	0.284	0.179	0.561	0.378	0.116	0.090	0.090
Plag	0.002	0.520	0.326	0.303	0.186	0.225	0.194	0.013	0.044
San	0.000	-0.403	-0.273	0.023	-0.141	-0.363	0.265	0.017	-0.059
Opx	0.160	1.701	0.675	0.006	0.549	0.441	0.005	-0.827	-0.250
Cpx	0.032	-7.560	-3.267	0.029	-1.579	-0.624	-0.009	-0.222	0.327
Qtz	0.131	45.301	20.586	0.382	8.269	-5.729	0.437	-23.311	-7.275
Mag	0.005	-1.391	-0.835	-0.005	0.163	-0.151	-0.002	-0.360	-0.274
Ilm	0.004	-0.680	-0.337	0.005	-0.090	-0.062	0.003	-0.297	-0.054
Zrn	0.000	0.001	0.001	0.001	0.001	0.000	0.001	0.001	0.000
Apat	0.001	0.108	0.050	0.001	0.016	0.010	0.001	0.011	0.001
Aln	0.001	0.003	0.002	0.001	0.001	0.001	0.001	0.002	0.001

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