1

MicroD

National Library
of Canada

Bibliothèque nationale
du Canada

Canadian Theses Service

Service des thèses canadiennes

Ottawa, Canada
K1A 0N4

NOTICE

The quality of this microform is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us an inferior photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this microform is governed by the Canadian Copyright Act, R S C 1970, c C-30

AVIS

La qualité de cette microforme dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de qualité inférieure.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, tests publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de cette microforme est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c C-30.

RETROGRESSION OF GRANULITE FACIES GNEISSES
IN THE HUNTSVILLE AREA,
SOUTHWESTERN GRENVILLE PROVINCE

by

PETER MICHAEL MANOJLOVIC B.Sc

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Master of Science

Department of Geology

Carleton University

Ottawa, Ontario

April 2, 1987

© copyright

1987, Peter Michael Manojlovic

Permission has been granted to the National Library of Canada to microfilm this thesis and to lend or sell copies of the film.

The author (copyright owner) has reserved other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without his/her written permission.

L'autorisation a été accordée à la Bibliothèque nationale du Canada de microfilmer cette thèse et de prêter ou de vendre des exemplaires du film.

L'auteur (titulaire du droit d'auteur) se réserve les autres droits de publication; ni la thèse ni de longs extraits de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation écrite.

ISBN 0-315-39379-3

The undersigned hereby recommend to the Faculty of Graduate Studies
and Research acceptance of the thesis

RETROGRESSION OF GRANULITE FACIES GNEISSES

IN THE HUNTSVILLE AREA,

SOUTHWESTERN GRENVILLE PROVINCE

submitted by Peter Michael Manojlovic, B.Sc

in partial fulfilment of the requirements for

the degree of Master of Science

These
.....
Thesis Supervisor

[Signature]
.....
Chairman, Department of Geology

Carleton University

ABSTRACT

Granulite facies gneisses in the Huntsville area of the southwestern Grenville Province are retrograded to amphibolite facies rocks in dekametre wide shear zones, adjacent to pegmatites. Excellent exposures provided a unique opportunity to examine the controlling petrogenetic variables involved in the transformation. Retrogression occurred at estimated conditions of 600 - 700°C and 600 - 700 MPa subsequent to peak metamorphic conditions of 700 - 750°C and 800 - 900 MPa. Bulk rock compositions did not change appreciably during retrogression and changes in mineral assemblages and continuous variation of mineral compositions and modal proportions reflect changes in pressure, temperature and the activity of water. Retrogression is attributed to an infiltration of water derived from the crystallization of pegmatite dikes.

ACKNOWLEDGEMENTS

I would like to thank Dr. E. Froese for his supervision, support and perceptive criticisms throughout this research. I thank L. Nadeau for the opportunity to work in the Huntsville area, for proposing this research, and especially for his enthusiastic interest in this work.

The help of P. Jones in instruction and assistance in the use of the electron microprobe is greatly acknowledged. I thank M. Bonardi and R. Hartree for carrying out microprobe analyses and whole rock analyses, respectively.

A. Davidson, R. Kretz and J. Moore are thanked for support and supervision during this study.

And lastly, I would like to thank family and friends for encouragement.

I would like to dedicate this thesis in memory of the late B.J. Baxter.

Table of Contents

Abstract	iii
Acknowledgements	iv
Table of Contents	v
List of Figures	vii
List of Tables	x
List of Plates	xi
Symbols and Abbreviations	xii
I Introduction	1
I.1 Purpose of study	2
I.2 Previous work	3
I.3 Location and access	5
I.4 Geological setting	5
II Field Relationships and Petrography	11
II.1 Felsic gneiss	11
II.1-1 Field relationships	11
II.1-2 Petrographic relationships	12
II.2 K feldspar augen gneiss	18
II.2-1 Field relationships	18
II.2-2 Petrographic relationships	19
II.3 Mafic gneiss	22
II.3-1 Field relationships	22
II.3-2 Petrographic relationships	23
III Whole Rock Geochemistry, Mineral Chemistry and Reactions	26
III.1 Whole rock geochemistry	26

III.1-1 XRF analysis	26
III.1-2 Whole rock compositional variations	27
III.2 Mineral chemistry	31
III.2-1 Electron microprobe analysis	31
III.2-2 Pyroxene	32
III.2-3 Garnet	38
III.2-4 Amphibole	43
III.2-5 Biotite	45
III.2-6 Plagioclase	52
III.3 Mineral reactions	54
III.3-1 Graphical representation	54
III.3-2 Felsic gneiss	57
III.3-3 K feldspar augen gneiss	59
III.3-4 Mafic gneiss	61
IV Conditions of Metamorphism	62
IV.1 Phase relationships	62
IV.2 Geothermobarometry	66
IV.2-1 Peak conditions	66
IV.2-1a Garnet - clinopyroxene thermometry	66
IV.2-1b Garnet - orthopyroxene thermometry	67
IV.2-1c Two pyroxene thermometry	69
IV.2-1d Summary of thermometry	70
IV.2-1e Garnet - pyroxene - plagioclase - quartz barometry	71
IV.2-1f Other barometers	72
IV.2-1g Summary of barometry	74
IV.2-2 Retrograde conditions	75

IV.2-2a Garnet - biotite thermometry	75
IV.2-2b Amphibole - plagioclase equilibria	76
IV.2-2c Summary of thermometry and barometry	77
IV.3 Summary of the Conditions of Metamorphism	79
V Discussion	81
V.1 The Process of Retrogression	81
V.1-1 Source of retrograde fluids	81
V.1-2 Mineral - fluid equilibria	82
V.1-3 Internal versus external controls on fluid composition	84
V.2 Variation in Mineral Chemistry Across the Amphibolite Facies to Granulite Facies Transition	86
VI Conclusions	89
VI.1 Proposal For Future Work	90
References	91
Appendix I Sketches of sample localities.	99
Appendix II Whole rock chemistry	106
Appendix III Mineral chemistry	114
Appendix IV Miscellaneous	127

List of Figures

Figure 1-1. Lithostructural subdivisions of the Central Gneiss Belt in the Huntsville region.	7
Figure 1-2. Sample localities.	8
Figure 3-1. Variation in Ca:Mg:Fe ²⁺ proportions among pyroxenes.	35
Figure 3-2. Mg/(Mg + Fe ²⁺) ratios in orthopyroxenes versus Mg/(Mg + Fe ²⁺) ratios in host rocks.	35

Figure 3-3.	Mg/(Mg + Fe ²⁺) ratios in clinopyroxenes versus Mg/(Mg + Fe ²⁺) ratios in host rocks.	35
Figure 3-4.	Ca/(Ca + Mg + Fe ²⁺) ratios versus Mg/(Mg + Fe ²⁺) ratios in orthopyroxenes.	35
Figure 3-5.	Ca/(Ca + Mg + Fe ²⁺) ratios versus Mg/(Mg + Fe ²⁺) ratios in clinopyroxenes.	35
Figure 3-6.	Mn/(Mg + Fe ²⁺ + Mn) ratios in orthopyroxenes versus MnO/(MgO + FeO + MnO) ratios in host rocks.	37
Figure 3-7.	Mn/(Mg + Fe ²⁺ + Mn) ratios in clinopyroxenes versus MnO/(MgO + FeO + MnO) ratios in host rocks.	37
Figure 3-8.	Mn/(Mg + Fe ²⁺ + Mn) ratios versus Mg/(Mg + Fe ²⁺) ratios in orthopyroxenes.	37
Figure 3-9.	Mn/(Mg + Fe ²⁺ + Mn) ratios versus Mg/(Mg + Fe ²⁺) ratios in clinopyroxenes.	37
Figure 3-10.	Na ₂ O content of clinopyroxenes versus NaO content of host rocks.	37
Figure 3-11.	Al ₂ O ₃ contents of orthopyroxenes versus Al ₂ O ₃ contents of host rocks.	37
Figure 3-12.	Al ₂ O ₃ contents of clinopyroxenes versus Al ₂ O ₃ contents of host rocks.	41
Figure 3-13.	Mg/(Mg + Fe ²⁺) ratios in garnets versus MgO/(MgO + FeO) ratios in host rocks.	41
Figure 3-14.	Ca/(Ca + Mg + Fe ²⁺ + Mn) ratios in garnets versus CaO/(CaO + MgO + FeO + MnO) ratios in host rocks.	41
Figure 3-15.	Normalized Ca/(Ca + Mg + Fe ²⁺ + Mn) ratios in garnets versus modal proportions of garnets.	41

Figure 3-16. Mn/(Mg + Fe ²⁺ + Mn) ratios in garnets versus MnO/(MgO + FeO + MnO) ratios in host rocks.	42
Figure 3-17. Normalized Mn/(Mg + Fe ²⁺ + Mn) ratios in garnets versus modal proportions of garnets.	42
Figure 3-18. Mg/(Mg + Fe ²⁺) ratios in hornblendes versus MgO/(MgO + FeO) ratios in host rocks.	47
Figure 3-19. Ca/(Ca + Mg + Fe ²⁺) ratios in hornblendes versus CaO/(CaO + MgO + FeO) ratios in host rocks.	47
Figure 3-20. Mn/(Mg + Fe ²⁺ + Mn) ratios in hornblendes versus MnO/(MgO + FeO + MnO) ratios in host rocks.	47
Figure 3-21. Normalized Mn/(Mg + Fe ²⁺ + Mn) ratios in hornblendes versus modal proportions of hornblendes.	47
Figure 3-22. Ti contents of hornblendes versus TiO ₂ contents of host rocks.	47
Figure 3-23. Ti contents of hornblendes versus modal proportions of hornblendes.	48
Figure 3-24. Al contents of hornblendes versus Al ₂ O ₃ contents of host rocks.	48
Figure 3-25. Fe ³⁺ /(Fe ³⁺ + Fe ²⁺) ratios in hornblendes versus Fe ₂ O ₃ + FeO ratios in host rocks.	48
Figure 3-26. Mg/(Mg + Fe ²⁺) ratios in biotites versus MgO/(MgO + FeO) ratios in host rocks.	51
Figure 3-27. Mn/(Mg + Fe ²⁺ + Mn) ratios in biotites versus MnO/(MgO + FeO + MnO) ratios in host rocks.	51
Figure 3-28. Normalized Mn/(Mg + Fe ²⁺ + Mn) ratios in biotites versus modal proportions of biotite.	51

Figure 3-29. Ti contents of biotites versus TiO_2 contents of host rocks.	51
Figure 3-30. Ti contents of biotites versus modal proportions of biotite.	51
Figure 3-31. Anorthite content versus Fe abundances in plagioclases.	53
Figure 3-32. Anorthite content versus orthoclase content in plagioclases.	53
Figure 3-33. Representative AFM diagram of mineral assemblages.	56
Figure 4-1. Phase equilibria delimiting the pressure and temperature conditions of granulite facies and amphibolite facies mafic gneisses.	65
Figure 5-1. Calculated a_{H_2O} versus temperature equilibrium.	85

List of Tables

Table 1-1. UTM coordinates of sample localities and sample number sequences.	10
Table 2-1. Estimated modes of selected felsic, K feldspar and mafic gneisses.	13
Table 3-1. Whole rock compositional variations.	30
Table 4-1. Summary of garnet-clinopyroxene thermometry.	68
Table 4-2. Summary of garnet-orthopyroxene thermometry.	68
Table 4-3. Summary of garnet-pyroxene-plagioclase-quartz barometry.	73
Table 4-4. Summary of garnet-biotite thermometry.	78
Table 4-5. Summary of amphibole-plagioclase equilibria.	78

List of Plates

- Plate 2-1. Photomicrographs of retrogression of felsic gneisses
and K feldspar augen gneisses. 17
- Plate 2-2. Photomicrographs of retrogression of K feldspar
augen gneisses and mafic gneisses. 21

Symbols, Units and Abbreviations

Symbols

K_D	Distribution Coefficient
X	Mole Fraction

Units

$^{\circ}\text{C}$	Degrees Celsius
kV	Kilovolts
mol	Mole
MPa	Megapascal (100 MPa = 1.0 Kb)
nA	Nanoampere
μm	Micrometre

Abbreviations

C	Confidence Limits
COV	Coefficient of Variation
EDS	Energy Dispersive Spectrometry
P	Pressure
SD	Standard Deviation
T	Temperature
UTM	Universal Transverse Mercator
WDS	Wavelength Dispersive Spectrometry
XRF	X-ray Fluorescence Spectrography

CHAPTER I

Introduction

The petrogenetic relationship between amphibolite facies and granulite facies rocks has sparked considerable debate since the introduction of the facies concept (Eskola, 1920). Originally, granulite facies rocks were considered to have formed by the progressive metamorphism of amphibolite facies precursors (Eskola, 1939). Recently, however, other petrogenetic models have been suggested. Granulites may be generated by the metamorphism of initially dry rocks containing insufficient H_2O to form hydrous minerals typical of the amphibolite facies (Buddington, 1963). Alternatively, granulites may be formed as a result of a lowering of fH_2O under high pressure and temperature conditions. Such a drop in fH_2O may be caused by the removal of H_2O -rich fluid into granitic liquids formed by partial melting to leave a residual, predominantly anhydrous, mineral assemblage (Brown and Fyfe, 1970). Lowering of fH_2O may also be due to the dilution of H_2O in deep crustal rocks by the addition of CO_2 -rich fluids expelled from the mantle (Newton et al., 1980). Anhydrous mineral assemblages characteristic of the granulite facies may also result from crystallization of H_2O -depleted magmas (Wendlandt, 1981). Each of these petrogenetic hypotheses has been applied in the interpretation of specific areas. These models are not mutually exclusive; they reflect rather the variability of geological conditions. Controlling petrogenetic variables must consequently be ascertained for each area.

Petrogenetic field studies of transitional amphibolite facies to granulite facies terranes are difficult, because of the rarity of well exposed, complete and gradational transitions from amphibolite facies to granulite facies mineral assemblages in rocks of the same composition. Compositional controls and effects are particularly difficult to decipher because extensive changes in fluid composition during granulitization may be accompanied by significant modifications of the bulk composition. Moreover, late retrograde metamorphic reequilibration is widespread and commonly obscures prograde effects.

Retrogression has recently received considerable attention in high grade rocks, particularly with regard to corona formation in metabasic rocks and in the zoning of minerals. Most of these studies have concentrated on evaluating pressure and temperature changes, particularly in relation to time, using solid-solid reactions. Hydration reactions during retrogression have, however, received somewhat less attention.

I.1 Purpose of Study

A number of road sections in the vicinity of Huntsville, Ontario, in the southwestern Grenville Province exhibit, on a dekametre scale, superb transition zones from amphibolite facies to granulite facies rocks. These transitions, developed in rocks of various distinct compositions, can be shown to be of retrograde origin. Rocks of a given bulk composition can be traced gradationally across these transition zones and can

be sampled easily. These rocks, therefore, provide a unique opportunity to determine the controlling petrogenetic variables involved in the transition from the amphibolite facies to the granulite facies.

The goals of this study are:

(1) to deduce the pressure and temperature conditions as well as the water activity, in rocks of a given bulk composition, prevailing at peak metamorphic conditions and during retrogression;

(2) to determine the effects of retrograde reequilibration in establishing new mineral assemblages and mineral compositions;

(3) to gain insight into processes of reequilibration; and

(4) to compare the mineral compositional variations present in these rocks with those found in other studies.

These goals have been accomplished by use of petrographic studies to determine compatible mineral assemblages and phase equilibria, by bulk rock analyses to determine whether significant chemical changes accompanied retrogression, and by microprobe analyses to determine pressure and temperature conditions.

1.2 Previous Work

Fyfe (1976) predicted that, upon cooling, intergranular fluid in metamorphic rocks will be combined into retrograde minerals. However, the amount of such fluid in high grade rocks

is very small and unlikely to account for a significant proportion of the fluid required in retrograde reactions. This view was corroborated by Yardley (1981) who pointed out that, during cooling, fH_2O in equilibrium with typical metamorphic assemblages drops drastically within 100°C of peak metamorphic conditions, demonstrating that extensive retrogression requires an influx of fluid. Vernon (1972), from his study of a reaction involving the hydration of orthopyroxene to anthophyllite, concluded that the reaction was essentially isochemical and that the introduction of water was the only additional reactant required. A similar conclusion was reached by Sen and Oliver (1981) for the formation of hornblende after hypersthene. In contrast Beach (1974, 1980), Sills (1982) and Paquet (1985) proposed reactions for the replacement of orthopyroxene and/or clinopyroxene by minerals such as hornblende, biotite, garnet and epidote involving considerable movement of ions in solution. The amount of dissolved species may be large enough to cause extensive alteration of the whole rock composition. Yardley and Baltatzis (1985) concluded that fluid would be internally buffered by the retrograde phases until a relatively large volume of fluid has passed through the rock. Influx of additional fluid becomes externally controlled after this critical volume has been reached and cation metasomatism then becomes possible.

I.3 Location and Access

Six, dekametre-wide transition zones of amphibolite facies to granulite facies rocks, exposed in blasted road sections in the Huntsville area, were selected for detailed study. Huntsville is located on Ontario Highway 11 approximately 200 km north of Toronto. It can also be reached from Ottawa via Ontario Highways 17 and 60. The transition zones studied are located along paved roads except for locality D (Figs. 1-1, 1-2). UTM coordinates of sample localities and sample number sequences are given in Table 1-1.

I.4 Geological Setting

The Grenvillian Orogeny in the Ontario Gneiss Segment, southwestern Grenville Province (Fig. 1-1), has been described as a 1000-1100 Ma crustal thickening event (Davidson, 1986). This event was the result of northwest-directed deep crustal thrusting, during which large segments of crust having contrasting rock assemblages, structural styles, metamorphic grade and geophysical signatures (thereafter referred to as domains and subdomains) were juxtaposed by movement along thin but continuous zones of highly strained flaggy L-S tectonites. These tectonites are interpreted as ductile shear zones (Davidson et al., 1982; Davidson, 1984; Davidson, 1986).

In Huntsville area, Algonquin domain has been subdivided into Huntsville and Novar subdomains. To the north,

rocks and structures of Huntsville subdomain structurally overlie and truncate those of Novar subdomain whereas they are overlain and truncated by those of Muskoka domain and Seguin subdomain to the south and along their western flank. Huntsville subdomain is underlain by thin, interleaved, shallowly southward-dipping, sheets of fine grained, commonly flaggy, strongly lineated and in part mylonitic gneisses of both supracrustal and plutonic origin (Fig.1-2)(Culshaw et al., 1982; Nadeau, 1985).

Huntsville subdomain is dominantly within the granulite facies; however, upper amphibolite facies assemblages are also common. Granitic and granodioritic gneisses at granulite grade are typically waxy, olive-green to brown on fresh surface and contain orthopyroxene with variable proportions of clinopyroxene, garnet, hornblende, biotite, accessory iron oxides and titanite. Perthitic feldspar is common. Pelitic gneisses commonly contain sillimanite, garnet, biotite, plagioclase, quartz and graphite (Culshaw et al., 1982; Nadeau, 1985). L. Anovitz (study in progress)*estimates peak metamorphic conditions to be near 1000 MPa and 750°C for several localities in the Huntsville area (Anovitz and Essene, 1986).

The widespread occurrence of transposed and flattened granitic metamorphic segregations, in virtually all rock types, and the presence of amphibolite facies mineral assemblages along the boundary of Muskoka domain indicate that peak metamorphic conditions were reached before deformation ceased (Nadeau, 1985). U-Pb zircon data on a granulite, just east of locality F, yielded a metamorphic age of 1030 $\pm$ 50 Ma (van Breemen et al., 1986). U-Pb zircon data from a sheared pegmatite and from a granulite

Figure 1-4 Lithostructural subdivisions of the Central Gneiss Belt in the Huntville region (after Davidson, 1984).

Figure 1-2. Sample locations and geology of the Huntsville area
(after Nadeau, 1985).

facies supracrustal rock from the northwest margin of the Parry Sound domain yielded a metamorphic age of 1161 ± 3 Ma indistinguishable from that of ductile shearing dated at 1159 ± 5 Ma (van Breemen et al., 1986).

The structure of the Huntsville subdomain is commonly disrupted by late metre-wide extensional shears geometrically similar to ductile and brittle-ductile shears described by Ramsay (1980) (Nadeau, 1985). The shears are typically steep with well developed deflection folds having gently plunging axes. They can rarely be traced for more than a few tens of metres and show offsets of a metre or so. Late granitic pegmatites are commonly associated with these late shear zones and exhibit, in places, wedge-shaped terminations. Although locally cutting fairly massive rocks such as metagabbro, these shears are preferentially developed in well layered rocks. Where transecting granulite facies gneisses, these shears are commonly associated with dekametre wide zones of retrogression. Felsic granulites forming the wall rocks of such shears exhibit a characteristic green to olive-brown colour. Some localities display the same green to pink-grey transition where no deflection folds or pegmatite dikes are present. In a few localities, deflection folds and associated pegmatite dikes cut mafic rocks. No colour change is evident in these rocks; however, a mineralogical change can be observed. The transitions described above form the basis of this study.

Numerous late pegmatites have been dated across the Grenville Province by K-Ar methods (Stockwell, 1962). An age of 950 Ma appears to be the youngest possible for these dikes and

suggests that retrogression occurred prior to this time. Whether retrogression occurred along a single (i.e. non-perturbed) pressure - temperature - time path or as a result of a clearly separate event is not known.

Table 1-1 Sample Localities and Number Sequences

Locality		UTM coord	Sample Nos.
A	North of Harp Lake	458269	A11
B	Junction Highways 60 and 11	385225	B3, B4, B5
C	Huntsville Lookout	455207	C7, C8
D	New Subdivision south Huntsville	435160	D9
E	Portage Road	506153	E15, E16, E17
F	Highway 60 5 km east of Dwight	605219	F18

Chapter II

Field Relationships and Petrography

The transitions studied in the field may be classified on the basis of their composition and mineralogical transformations. The most common type of transition occurs in felsic gneisses: localities B, C and F have transitions of this type. The next most abundant type of transition occurs in K feldspar augen gneisses. This type of transition occurs at localities D and E. One example of the transition in mafic gneiss was studied at locality A. Sketch maps of each locality are given in Appendix I.

II.1 Felsic Gneiss

II.1-1 Field Relationships

In outcrop exposures, the transition occurs in deflection folds which have associated subvertical pegmatite dikes. At locality B, the amount of deflection is very minor to absent. The transition at localities B, C and F commences with a green to olive-green, moderately foliated gneiss, composed essentially of feldspars and quartz with fine grained mafic minerals. Deep red porphyroblasts of garnet stud the rock. Rare lozenge-shaped K feldspar augen and thin, highly attenuated leucosomes can be seen in many of the localities. Some quartz blades are present. The foliation in the rock is defined by thin (1-2 mm wide) attenuated

mafic layers. Quartz and mafic minerals commonly define a lineation. The layered aspect of the gneiss is commonly masked by the colouration of the feldspars.

The first stages of retrogression involve a discolouration of the green feldspars to a pink (localities B and F) or a grey colour (locality C). As the colour changes, garnets decrease in abundance and become replaced by hornblende and/or biotite. In the most strongly discoloured zone, garnets are rare, having been replaced in most rocks by elongated aggregates of hornblende, biotite and pale grey feldspar. The spotted appearance imparted by garnet to the greenish rocks becomes obliterated by the foliation developed in the late folds. A slight coarsening of grain size accompanies this transformation.

II.1-2 Petrographic Relationships

In thin section, green coloured gneisses contain variable proportions of orthopyroxene, clinopyroxene, garnet, K feldspar, plagioclase, quartz and opaque oxides (Table 2-1). The grain size is fine, with an average size of less than 1 mm. Feldspars are the most abundant minerals and typically have shapes approaching granoblastic with some flattening, which helps define the foliation. K feldspar is predominant and in some rocks occurs as augen up to 3 mm in size. Perthite is developed in some grains. Plagioclase is present throughout, but is most abundant as granoblastic grains which partially surround garnets. Antiperthite is only very locally developed. Elongated blades of

Table 2-1 Typical Modal Proportions

Min.★ Rock†	Locality														
	MG		FG			FG		KAG		KAG		FG			
	G#	A	G	I	A	G	I	G	A	G	A	G	I	A	
Ksp	5	5	45	40	40	40	40	35	40	30	30	35	35	40	
Pl	45	55	20	15	20	15	15	20	20	25	25	20	20	20	
Qtz	20	10	25	35	30	30	30	30	30	30	25	25	25	25	
Opx	tr	-	1	-	-	tr	-	5	-	10	-	4	-	-	
Cpx	6	-	1	-	-	2	-	-	-	-	-	5	-	-	
Grt	8	2	5	3	-	5	3	-	2	-	5	7	5	-	
Bio	2	5	1	3	2	1	5	5	6	3	10	tr	5	1	
Hbd	15	23	-	2	5	6	7	2	tr	tr	-	1	5	10	
Mag	tr	tr	3	3	2	2	1	4	3	3	2	3	3	4	
Ilm	tr	tr	-	-	-	-	-	-	-	-	-	tr	-	-	
Pyr	tr	tr	tr	tr	-	tr	tr	tr	-	tr	tr	tr	-	-	
Py	-	-	-	-	-	-	-	tr	tr	tr	tr	-	tr	-	
Cpy	-	-	-	-	-	-	-	tr	-	tr	-	-	-	-	
Apa	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	
Ti	-	-	-	-	-	-	-	-	-	-	-	-	tr	tr	
All	-	-	-	-	-	-	-	-	-	-	-	-	tr	tr	

Visual Estimates

* Ksp=K feldspar, Pl=plagioclase, Qtz=quartz, Opx=orthopyroxene, Cpx=clinopyroxene, Grt=garnet, Bio=biotite, Hbd=hornblende, Mag=magnetite, Ilm=ilmenite, Pyr=pyrrhotite, Py=pyrite, Cpy=chalcopyrite, Apa=apatite, Ti=titanite, All=allanite;

G=granulite, I=intermediate, A=amphibolite.

+ MG=mafic gneiss, FG=felsic gneiss, KAG=K feldspar augen gneiss. Based on samples AllaG, AllfA, B5cG, B5pI, B5mA, C7dG, C7aI, D9hG, D9aA, E15bG, E15cA, F18aG, F18nI, and F18sA. Additional modal proportions in Appendix IV.

Note: Biotite in samples marked G probably is retrograde; Hornblende in samples marked G probably is retrograde, except in AllaG; Garnet in samples marked A is retrograde.

recrystallized quartz help define the foliation. Quartz aggregates have typical length to width ratios of 5:1. Orthopyroxene is commonly present (at locality C only traces occur)(Plate 1-A). Grains are equant, have only a very weak pleochroism and have an average grain size $<0.5\text{mm}$. Fresh grains are typically in contact with magnetite. A number of grains have a fine intergrowth of orange-red biotite, probably of retrograde origin, which grows along cleavage traces. Clinopyroxene is present at all localities and occurs as equant, light green grains with an average grain size of 0.5 mm . Magnetite is commonly in contact with the clinopyroxene. Clinopyroxene is, generally isolated from all Fe-Mg minerals. In some thin sections, yellow-green to dark-green amphibole, probably retrograde in origin, is in contact with the clinopyroxene. Garnet is common and occurs as larger, isolated grains with an average grain size around 2 mm . Grains are rounded to bulbous in shape with common inclusions of quartz, plagioclase and magnetite (Plate 1-B). Grains typically have a discontinuous rim of granoblastic plagioclase, quartz, orthopyroxene and clinopyroxene. Orthopyroxene, clinopyroxene and garnet rarely occur in mutual contact; however, they appear to represent the stable high grade assemblage.

As the zone of retrogression is approached, the first notable change is the disappearance of orthopyroxene, as biotite becomes more abundant. This biotite is an orange-red variety which commonly occurs as feathery blades (average size 0.5 mm) near and surrounding magnetite (Plate 1-D). In places this biotite clearly cuts the foliation. With increased retrogression, dark green to yellow-green amphibole becomes more abundant.

This amphibole occurs as fine grains (<0.5 mm) and typically has quartz inclusions in it (Plate 1-C). Cores of amphiboles commonly contain magnetite. As amphibole increases, there is a concomitant decrease in the volume of clinopyroxene. The two minerals may occur in mutual contact with apparent textural equilibrium. Garnets appear to be little changed during the initial stages of retrogression; however, inclusions of amphibole and biotite become increasingly more abundant as retrogression proceeds (Plate 1-C). At site C retrogression does not advance farther than this stage. At sites B and F garnets are smaller in size and more irregular in shape. Moats of plagioclase and quartz develop between the garnet and biotite while amphibole becomes more abundant during this stage of retrogression, until all garnet has disappeared. The most retrograded varieties have dark green amphibole and brown biotite defining the foliation (Plate 1-E). Amphibole no longer have quartz inclusions in them. Magnetite is present but appears to be less abundant. The exsolution found in some feldspars decreases throughout the retrogression. Grain size and shape appear to be little affected; however, the tartan twinning of microcline becomes more abundant. At locality F, spots of allanite and titanite surrounding opaque oxides are common in the most retrograded rocks. The final assemblage is composed of hornblende, biotite, K feldspar, plagioclase, magnetite, quartz, allanite and titanite, all of which appear to be in equilibrium.

Plate 2-1

Felsic Gneiss

- 2-1A Section F18aG. Anhydrous granulite with orthopyroxene, clinopyroxene, K feldspar, plagioclase, quartz and magnetite. Note 120° boundaries at plagioclase contacts; plane polarized light; scale bar is 0.25 mm.
- 2-1B Section F18aG. Anhydrous granulite showing typical rounded garnet with inclusions of quartz and magnetite. Note orthopyroxene and clinopyroxene separated from garnet by plagioclase and quartz; plane polarized light; scale bar is 0.25 mm.
- 2-1C Section F18nI. Intermediate stage of retrogression with garnet showing only minor reaction where in contact with biotite. Pyroxenes have reacted to form hornblende with quartz inclusions; plane polarized light; scale bar is 0.25 mm.
- 2-1D Section F18nI. Intermediate stage of retrogression showing biotite surrounding magnetite. Note biotites, with quartz inclusions, have a variety of orientations; plane polarized light; scale bar is 0.25 mm.
- 2-1E Section F18rA. Completely retrograded felsic gneiss with sharp grain boundaries between biotite and hornblende. Note 120° grain boundaries of hornblendes; plane polarized light; scale bar is 0.25 mm.

K feldspar Augen Gneiss

- 2-1F Section E15bG. Anhydrous portion of K feldspar augen gneiss with orthopyroxene, magnetite, plagioclase, K feldspar and quartz; plane polarized light; scale bar is 0.25 mm.

A

B

C

D

E

F

II.2 K Feldspar Augen Gneiss

II.2-1 Field Relationships

In the K feldspar augen gneisses, the transition is again marked by a colour change of feldspars from olive green to pinkish-grey. Two types of transition were studied. At locality D, the transition occurs along the edges of pegmatite dikes which are associated with deflection folds. Locality E does not have any deflection folds associated with the transition. The transition zones at this locality are narrow, up to a half a metre wide. Subhorizontal zones of retrogression interfinger on the outcrop.

The transition commences from the olive green rock which contains variable proportions of K feldspar augen up to 2 cm across. These augen are flattened and contain cores of unrecrystallized greenish-grey feldspar surrounded by a mantle of recrystallized feldspar. The matrix of these rocks is composed of very fine grained mafic minerals among which biotite is megascopically identifiable. In outcrop, the rock is fairly homogeneous; with K feldspar augen make up about 10%. On a finer scale, the rocks are heterogeneously deformed with zones of high strain having a more layered appearance. The layers are thin (1-2 mm) and composed of alternating mafic-rich and feldspar-rich zones. The feldspar rich zones are predominantly produced by recrystallization of K feldspar augen. Macroscopically there is little mineralogical change as the feldspars change colour. The only notable change, seen in outcrop, is the appearance of very

tiny red garnets in the pinkish zones. The colour change is not accompanied by any coarsening of grain size.

II.2-2 Petrographic Relationships

In thin section, dark greenish samples from both localities are composed of variable proportions of orthopyroxene, biotite, K feldspar, plagioclase, quartz, amphibole, and opaque minerals (Table 2-1). Feldspars, the predominant minerals, occur as fine grained (average grain size 0.5 mm) granoblastic aggregates. Larger aggregates of feldspars, with lozenge-shaped outlines (2-4 mm in length) also occur. The finer grained feldspars appear to be recrystallized varieties of the larger augen. K feldspar is more abundant than plagioclase and exhibits perthitic exsolution. Antiperthite is locally developed in plagioclase. Quartz occurs as recrystallized elongate blades which have a broad undulose extinction and average length-to-width ratios of 8:1. Orthopyroxene occurs as fresh, elongate aggregates of grains which help define the foliation (Plate 1-F). Grains are typically 0.25 mm in size and have weak pleochroism. Orthopyroxene is commonly in contact with magnetite. Biotite is very common and occurs in more or less confined streaky zones as individual orange-brown to red-brown grains (average grain size 0.25 mm). Some biotite grains are in contact with opaque oxides and orthopyroxene with no apparent reaction relationship (Plate 2-A). Amphibole occurs sporadically in thin layers. This amphibole, on average, is 0.25 mm in size and has a light yellow-green

Plate 2-2

K feldspar Augen Gneiss

2-2A Section E15bG. Granulite showing sharp grain boundaries between biotite and orthopyroxene; plane polarized light; scale bar is 0.25 mm.

2-2B Section E15cA. Completely retrograded K feldspar augen gneiss showing sharp grain boundaries between garnet and biotite; plane polarized light; scale bar is 0.25 mm.

Mafic Gneiss

2-2C Section AllaG. Granulite with orthopyroxene, clinopyroxene, garnet and hornblende; Note 120° grain boundaries between plagioclases; plane polarized light; scale bar is 0.25 mm.

2-2D Section AllaG. Granulite with clinopyroxene, garnet and hornblende; plane polarized light; scale bar is 0.25 mm.

2-2E Section AllfA. Retrograded mafic gneiss with hornblende, plagioclase, quartz and magnetite. Note quartz blebs in hornblendes; plane polarized light; scale bar is 0.25 mm.

2-2F Section AllfA. Completely retrograded mafic gneiss with euhedral garnet, biotite and bleb free hornblende; plane polarized light; scale bar is 0.25 mm.

A

B

C

D

E

F

to light green colour. Magnetite is commonly associated with this amphibole.

The green-to-pink transition in hand specimen is sharply defined in thin section as well. The first stage of retrogression involves a growth of orange biotite along cleavage traces in the orthopyroxene. As retrogression proceeds this alteration becomes more fibrous and orthopyroxene becomes engulfed in streaky layers of biotite grains. In many places these biotite streaks can be traced along strike into garnet - bearing zones. Garnets are subhedral to euhedral in shape and commonly occur as grains up to 0.5 mm situated with biotite blades (Plate 2-B). In places garnets can be seen in contact with fibrous biotite aggregates which have replaced orthopyroxene. Garnet was only rarely seen in contact with orthopyroxene and it appears that the orthopyroxene is unstable with garnet. Some anhedral garnets occur in orthopyroxene- and biotite-free zones. Amphiboles and feldspars appear to be little affected by the retrogression. Overall it appears that the retrogression involves conversion of orthopyroxene to biotite and garnet. This transformation occurs over a short distance of 3-4 cm.

II.3 Mafic Gneiss

II.3-1 Field Relationships

Only one very small outcrop of mafic gneiss was found which unequivocally displayed the retrograde transition. Two pegmatites cut through this outcrop and, although there are no

distinct colour change or deflection folds associated with the transition, a mineralogical change is evident.

The transition in the mafic rocks initiates from a dark, fine grained gneiss which is composed of fine grained mafic minerals and green feldspars. The sugary rock is studded with red garnets. Throughout this rock irregular leucosomes, up to one centimetre thick, cut across the foliation. These leucosomes are foliated, the foliation being defined by quartz blades and thin streaks of mafic minerals. Leucosomes make up about 10% of the rock.

Retrogression near the pegmatites results in a decrease in the amount of garnet present. The original gneiss contains about 10% garnet, whereas the retrograded variety contains only a small percentage. The grain size of the mafic minerals increases slightly and biotite becomes obvious with increased retrogression. Throughout the retrogression, the foliation and the leucosomes appear to be little affected.

II.3-2 Petrographic Relationships

In thin section, the mafic gneiss is seen to be composed of orthopyroxene, clinopyroxene, garnet, amphibole, biotite, plagioclase, K feldspar, quartz and magnetite (Table 2-1). Plagioclase is the dominant feldspar and makes up roughly half of the rock. Feldspars form equigranular, granoblastic grains with an average grain size of 0.75 mm. No perthitic or antiperthitic exsolution was found in these grains. Quartz occurs in thin

streaky zones (leucosomes) which cut through the rock. These grains are approximately 0.75 mm in width and have length to width ratios of 8:1. Orthopyroxene is very sparse in this rock, occurring as isolated equant grains 0.5 mm in size (Plate 2-C). A bit of biotite occurs along the cleavage traces. Clinopyroxene occurs as 0.75 mm anhedral aggregates which have a weak green colour (Plate 2-D). Garnets occur as larger (2 mm) grains which are typically in contact with the amphibole and magnetite (Plate 2-D). These grains are subhedral to euhedral and do not appear to be in contact with clinopyroxene. Amphibole is brown-green to yellow-green in colour and is granoblastic in shape (average grain size 0.75 mm). Magnetite commonly occurs in the cores of amphiboles (Plate 2-D). Where amphibole is in contact with clinopyroxene, it commonly contains worm-like inclusions of quartz which may be an indication that the amphibole is replacing clinopyroxene and in part separating it from garnet (Plate 2-E). Red-brown to yellow-brown biotite, up to 1.0 mm in size, is associated with amphibole and garnet, and in places cuts the foliation. Some biotite grains have quartz inclusions in them; possibly this is an indication that they have replaced orthopyroxene. It appears that no place in the outcrop has escaped the effects of retrogression; however, the highest grade assemblage may be inferred to have been orthopyroxene, clinopyroxene, garnet, hornblende, plagioclase, K feldspar and magnetite.

With increasing retrogression, pyroxene disappears and garnet becomes greatly diminished in abundance. Two types of garnet are found in the most retrograded rocks present. The first type occurs as irregular clumps which are surrounded by

amphibole and appear to be relict high grade grains. The second ~~type~~ is fine grained (average size 0.25 mm), and occurs as perfectly euhedral grains associated with biotite blades (Plate 2-F). Biotite has become more abundant and occurs as orange-brown grains, 1-1.5 mm in size, some of which cut the foliation. Magnetite is common with this biotite. Amphibole is more abundant as well, and has a paler colour than primary amphibole. Amphibole is commonly euhedral (average grain size 1.5 mm) and occurs as aggregates. Many of the larger grains contain quartz inclusions. Feldspars appear to be little affected by the retrogression. The stable retrograde assemblage appears to be amphibole, biotite, garnet, plagioclase, K feldspar and magnetite.

Chapter III

Whole Rock Geochemistry, Mineral Chemistry and Reactions

III.1 Whole Rock Geochemistry

III.1-1 XRF Analysis

The chemical composition of twenty-two samples of felsic, K feldspar augen and mafic gneisses was determined by X-ray fluorescence (XRF) in order to examine the possibility that changes in whole rock composition accompanied retrogression. XRF analyses were obtained at the University of Ottawa using a Philips PW 1410/20 AHP instrument. FeO in these samples was determined by titration at the Geological Survey of Canada; Fe₂O₃ was calculated by difference. Estimates of the precision of the chemical analyses is based on the coefficient of variation values obtained by Kretz et al., (1985) in the same laboratory as this study. A 95% confidence limit of ± 1.96 SD (standard deviation) was used for each sample. H₂O was determined by a Leco RMC-100 water analyser at the Geological Survey of Canada. Estimated error is ± 0.1 wt.%.

Results are given in Appendix II. Samples are prefixed by their locality symbol and postscripted by either G, I or A, where G = granulite: unaltered to weakly retrograded; I = intermediate: dominantly retrograded with relict garnet in felsic gneiss and relict orthopyroxene in K feldspar augen gneiss; A = amphibolite: retrogression more or less complete. Analyses for

each locality are listed by increasing degree of retrogression.

III.1-2 Whole Rock Compositional Variations

All the samples studied have molecular proportions of $\text{Al}_2\text{O}_3 < \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$, suggesting that they were derived from igneous precursors or immature clastic sediments. Felsic gneisses have compositions similar to granite whereas K feldspar augen gneisses resemble granodiorite. Mafic gneisses are similar in composition to gabbro (or diabase).

In order to examine possible chemical variations due to retrograde processes, the following criteria were used to determine whether compositional differences exist. Only where there is no overlap in the confidence limits were adjacent and/or more retrograded samples considered to be different in composition, unless a series of three or more samples displays a consistent trend in composition. Only major elements considered here. Compositional variations in each suite are shown in Table 3-1.

Possible compositional variations were most extensively studied in felsic gneisses. Suites of four, three and eight samples were studied at localities B, C and F respectively. Within-suite compositional variation is extremely restricted; the majority of samples have compositions which overlap the 95% confidence limits of the suite. Compositional variations which do occur are: (1) Al_2O_3 and FeO decreased while SiO_2 and Fe_2O_3 increased with retrogression at locality B; (2) CaO and possibly K_2O decreased while MgO and H_2O increased with retrogression at

locality C, and: (3) Na_2O decreased while CaO , K_2O and H_2O increased with retrogression at locality F. If sample F18aG is ignored. A comparison of the ratios $\text{MgO}/(\text{MgO} + \text{FeO})$ and $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO})$ for samples F18aG and F18yG revealed that these ratios do not vary between samples. The somewhat divergent composition of sample F18aG is therefore probably the result of an incorporation of an anomalous proportion of mafic lens material, common at locality F, into the powder used for analysis.

Compositional variations due to retrogression in K feldspar augen gneisses were examined in suites of three and two samples at localities D and E respectively. Within-suite compositional variations are somewhat more evident than in felsic gneisses; however, this may be due to the limited number of samples analyzed. Compositional variations which occur are: (1) FeO and MgO decreased while K_2O and H_2O increased with retrogression at locality C, and; (2) SiO_2 and H_2O increased with respect to all other elements at locality E, however, it remains within the 95% confidence limits. The ratios $\text{MgO}/(\text{MgO} + \text{FeO})$ and $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO})$ are similar in the two samples analyzed at locality E suggesting little variation among major oxides other than SiO_2 .

Compositional variation is greatest within a suite of two samples of mafic gneiss from locality A. There is a decrease in SiO_2 , Na_2O and Al_2O_3 and an increase in FeO , MgO , CaO and H_2O with retrogression. However, a comparison of $\text{MgO}/(\text{MgO} + \text{FeO})$ and $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO})$ ratios of non-retrograded and retrograded samples revealed that the two samples are similar in this respect. It appears that the chemical variation between the two

samples may be due to more leucosome material being added into the powders used for analysis of sample AllaG, rather than resulting from retrogression.

The following conclusions can be made concerning possible chemical variations due to retrogression: (1) major (and trace) element concentrations of the gneisses do not appear to change with retrogression since no consistent change in composition is apparent; (2) chemical variations present within individual suites may be traced to variations in protolith composition, sampling errors or an insufficient number of samples used to correctly characterize any chemical variation; (3) H_2O -contents are higher in retrograded gneisses than in non-retrograded gneisses, having been added during retrogression.

Table 3-1 Whole Rock Compositional Variations

	Locality					
	B FG*	C FG	F FG	D KAG	E KAG	A MG
SiO ₂	+	0	0	0	+	-
Al ₂ O ₃	-	0	0	0	-	-
Fe ₂ O ₃	+	0	0	-	-	-
FeO	-	0	0	-	-	+
MgO	0	+	0	-	-	+
CaO	0	-	+	-	-	+
Na ₂ O	0	0	-	0	-	-
K ₂ O	0	-	+	+	-	+
TiO ₂	0	0	+	-	-	-
P ₂ O ₅	0	0	+	0	-	-
MnO	0	+	0	+	0	-
H ₂ O	+	0	+	+	+	+

+ increases with retrogression; - decreases with retrogression; 0 no change.

* MG = mafic gneiss; FG = felsic gneiss;

KAG = K feldspar augen gneiss.

III.2 Mineral Chemistry

III.2-1 Electron Microprobe Analysis

Quantitative wavelength dispersive (WDS) electron microprobe analysis was used to analyze selected minerals in samples B5cG, B5pI, B5mA, F18aG, F18yG, F18nI, F18rI, D9aA, D9hG, E15aG and E15bA. Analysis was carried out by the author using a Cambridge Microscan 5 electron microprobe at Carleton University. Operating conditions were: 15 kV accelerating potential, specimen current of 50 nA as measured on pure Fe and a beam diameter of 2-3 μ m. All elements were counted for 20 seconds. X-ray counts were reduced to oxide weight percents using the program EMPADR VII of Rucklidge and Gasparrini (1969). Estimated errors are $\pm$ 3% (relative) for major elements and $\pm$ 10% (relative) for minor elements.

Analyses of minerals in samples AllaG, AllfA, C7aA, C7dG, F18sA and B3aG were carried out by M. Bonardi of the Geological Survey of Canada using a Material Analysis Company electron microprobe, equipped with a Revox energy dispersive spectrometer (EDS). Operating conditions were: 20 kV accelerating voltage, specimen current of 10 nA measured on a standard kaersutite and a counting time of 100 seconds. Data were reduced by the method of Plant and Lachance (1973). Determinations have a relative accuracy of $\pm$ 1-2% for major elements and up to $\pm$ 10% for minor elements.

Results of electron microprobe analysis are given in Appendix III. Samples are prefixed by their locality symbol and

postscripted by either G, I or A as with whole rock chemical analyses. Reported compositions are averages for each thin section (number of points analyzed given in Appendix III). Initial work by EDS demonstrated that mineral compositions did not vary significantly within individual grains or thin sections except for minor variations in core to rim analyses of larger garnets. More detailed analysis is required to define this zoning and is beyond the scope of this work.

III.2-2 Pyroxene

Analyzed pyroxenes from selected samples of felsic gneiss, K feldspar-augen gneiss and mafic gneiss are presented in Appendix III.1. Pyroxene analyses were made of fresh, unaltered grains. No orthopyroxene analysis was obtained for locality C because the amount present is trivial and none was present in the thin sections used for microprobe work.

The electron microprobe analyses of pyroxene were calculated on the basis of 6 oxygens. An appropriate amount of Fe^{2+} was converted, where necessary, to Fe^{3+} or the number of cations was reduced to supply 12 charges. Calculated Fe^{3+} is strongly dependent on minor variations in major oxides, particularly SiO_2 . Results of the calculation are broadly consistent with Fe^{3+} determined for pyroxenes from other granulite localities (Jen and Kretz, 1981; Stephenson, 1984).

The variation in Ca:Mg:Fe^{2+} proportions among the py-

roxenes is illustrated in Fig. 3-1. Tie lines have been drawn between coexisting pyroxenes. Pyroxenes are consistent with the metamorphic trend of Huebner (1980). Clinopyroxenes can be classified as augite and orthopyroxenes as hypersthene. The $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios range from 0.32 to 0.57 in orthopyroxenes, from 0.32 to 0.71 in clinopyroxenes, and appear to be systematically related to the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in the host rock (Fig. 3-2, 3-3). Pyroxenes in section B5cG are somewhat more Mg-rich than the average trend.

CaO contents of the pyroxenes do not appear to be related to host rock CaO; however, they show some correlation with the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios of the pyroxenes. CaO contents decrease with increasing $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in orthopyroxene while there is a weak trend for CaO to increase in clinopyroxene as the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratio increases (Fig. 3-4, 3-5). This is consistent with observations on natural pyroxenes as well as experimental studies on pyroxenes (Stephensen, 1984).

MnO contents in the pyroxenes show no relationship to host rock MnO or the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratio of the pyroxenes (Figs. 3-6, 3-7, 3-8, 3-9).

Na_2O contents of orthopyroxenes are below detection limits in most rocks. In the clinopyroxenes it shows no correlation to host rock Na_2O (Fig. 3-10).

The Al_2O_3 contents of the pyroxenes are relatively low as compared to pyroxenes from other granulite terrains (Stephensen, 1984). Pyroxenes display a trend towards increasing Al_2O_3 with host rock Al_2O_3 content (Figs. 3-11, 3-12). Al_2O_3 contents of pyroxenes in the mafic gneiss are markedly greater

Figure 3-1. Variation in Ca:Mg:Fe^{2+} proportions among pyroxenes.

Figure 3-2. $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in orthopyroxenes versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in host rocks.

Figure 3-3. $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in clinopyroxenes versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in host rocks.

Figure 3-4. $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+})$ ratios versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in orthopyroxenes.

Figure 3-5. $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+})$ ratios versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in clinopyroxenes.

NOTE: In Figure 3 ratios in rocks refer to weight percent, ratios in minerals refer to atomic percent.

Figure 3-6. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in orthopyroxenes versus $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in host rocks.

Figure 3-7. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in clinopyroxenes versus $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in host rocks.

Figure 3-8. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in orthopyroxenes.

Figure 3-9. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios versus $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in clinopyroxenes.

Figure 3-10. Na_2O content of clinopyroxenes versus NaO content of host rocks.

Figure 3-11. Al_2O_3 contents of orthopyroxenes versus Al_2O_3 contents of host rocks.

than in the other gneisses.

III.2-3 Garnet

Garnets were analyzed in felsic gneisses, K feldspar augen gneisses and mafic gneiss from a variety of parageneses. Analyses are from granulites at localities A, B, C and F (Al1aG, B5cG, C7dG, F18aG and F18yG), from moderately retrograded gneisses at localities B, C and F (B5pI, C7aI and F18nI) and from completely retrograded gneisses at localities A, D and E (Al1fA, D9aA and E15cA). Analyses are given in Appendix III.2. Garnets have cores which have a broadly common composition while rims show some enrichment of Mn. This is typical of high grade garnets (Tracy, 1982).

Garnets were calculated on the basis of 12 oxygens. An appropriate amount of Fe^{2+} was converted to Fe^{3+} , to make a total of 24 positive charges. Calculated Fe^{3+} values are low which is consistent with the common observation that high-grade garnets have low contents of andradite component.

The analyzed samples are essentially iron-rich almandine garnets containing 51 to 72 mole % almandine, 4 to 20 mole % pyrope, 15 to 25 mole % grossular and 3 to 19 mole % spessartine. These garnets are chemically similar to those found in high-grade gneisses.

The $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios of garnets appear to be broadly controlled by the $\text{MgO}/(\text{MgO} + \text{FeO})$ ratios of their host rocks (Fig. 3-13; Corresponding garnets from each locality are

shown on this and subsequent figures by a tie line with an arrow pointing in the direction of increasing retrogression). There is no consistent trend in $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios of garnets as retrogression proceeded. Garnets at localities A and B show an increase in their $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratio with increasing degree of retrogression whereas the reverse trend is present in garnets from localities C and F.

The $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+} + \text{Mn})$ contents of the garnets show no systematic relationship to their host rock $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO} + \text{MnO})$ ratio (Fig. 3-14). No clear trend in the variation of $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+} + \text{Mn})$ content of garnets with the degree of retrogression is present; however, there appears to be some increase in this ratio in garnets from felsic gneisses as retrogression proceeded. This increase also coincides with a decrease in the modal proportion of garnet as retrogression proceeded. The reverse appears true for the mafic gneiss (Fig. 3-15).

Analyzed garnets from the partially retrograded rocks contain an appreciably higher spessartine content than the non-retrograded granulites, and Fig. 3-16 suggests that this does not merely reflect the MnO contents of the rocks. Fig. 3-17 shows that the garnet $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios tend to increase relative to the $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in their host rocks as the modal abundance of garnet decreases upon retrogression.

Figure 3-12. Al_2O_3 contents of clinopyroxenes versus Al_2O_3 contents of host rocks.

Figure 3-13. $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in garnets versus $\text{MgO}/(\text{MgO} + \text{FeO})$ ratios in host rocks.

Figure 3-14. $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in garnets versus $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO} + \text{MnO})$ ratios in host rocks.

Figure 3-15. Normalized $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in garnets versus modal proportions of garnets.

Figure 3-16. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in garnets versus $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in host rocks.

Figure 3-17. Normalized $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in garnets versus modal proportions of garnets.

III.2-4 Amphibole

Amphiboles were analyzed from granulites at localities A, B, C, D and F (AllaG, C7dG, D9hG, F18aG and F18yG). Petrography (Chapter 2) suggests that the amphibole in these samples is probably retrograde, except for AllaG in which it appears to be primary. Amphiboles were also analyzed from moderately retrograded gneisses at localities B, C and F (B5pI and C7aI) and from completely retrograded gneisses at localities A, B and F (AllfA, F18rA and F18sA). Analyses are given in Appendix III.3.

The problem of assigning Fe^{2+} and Fe^{3+} in an electron-probe analysis is well known. A number of calculation schemes have been proposed (Robinson et al., 1982). For this work it is assumed that all cations except Ca, Na and K occupy 13.000 sites and that the total cationic charge is 46. This method assumes that there is no cummingtonite or anthophyllite molecule. This calculation scheme commonly gives good results for Ca amphiboles (Robinson et al., 1982). Resulting $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratios are in agreement with analyses obtained by wet chemical analysis from granulite and amphibole grade rocks elsewhere (Stephensen, 1977). Comparison with amphiboles analyzed by Jen and Kretz (1981) suggests, however, that ferric iron may be overestimated by this method. Regardless of the normalization scheme used, the trends in amphibole composition will be apparent.

Amphiboles are mostly magnesian hastingsite (AllaG, B5mA, D9hG, F18aG and F18sA) and magnesian hastingsitic hornblende.

blende (Al1fA, B5pI, F18yG and F18rA) (Leake, 1978), the main distinction between the two is a higher Si content in the magnesian hastingsitic hornblendes. Amphibole from B5pI has low A-site occupancies and thus borders the tschermakitic hornblende field (Leake, 1978). Amphiboles in sections C7aI and C7dG have low $Mg/(Mg + Fe^{2+})$ ratios and thus are hastingsites and hastingsitic hornblendes respectively (Leake, 1978). These amphiboles will be termed hornblendes for convenience in further discussions; they are chemically similar to those found in other high-grade gneisses (Stephensen, 1977).

The $Mg/(Mg + Fe^{2+})$ ratios of hornblendes show a crude correlation with their host rock $MgO/(MgO + FeO)$ ratio (Fig. 3-18). A trend towards decreasing $Mg/(Mg + Fe^{2+})$ ratios as retrogression proceeded is evident in hornblendes from felsic gneisses.

$Ca/(Ca + Mg + Fe^{2+})$ ratios of hornblendes show no relationship to their host rock $CaO/(CaO + MgO + FeO)$ ratio and indeed do not vary significantly as their paragenesis changes upon retrogression (Fig. 3-19).

Hornblende $Mn/(Mg + Fe^{2+} + Mn)$ ratios clearly do not display a relationship to their host rock $MnO/(MgO + FeO + MnO)$ ratio. However, a very clear relationship does emerge whereby the $Mn/(Mg + Fe^{2+} + Mn)$ ratio in hornblendes increased significantly as retrogression proceeded (Fig. 3-20). This increase also corresponds to an increase in the modal proportion of hornblende with increasing retrogression (Fig. 3-21).

The Ti contents of the hornblendes does not appear to be related to the TiO_2 contents of their host rocks (Fig. 3-22).

Hornblende from B5pI appears to be anomalously low in Ti as compared to the other hornblendes. If this sample is ignored, there appears to be an overall decrease in the Ti content of hornblende as the retrogression proceeded. Fig. 3-23 suggests that this change coincides with an overall increase in the modal proportion of hornblende.

Al contents of hornblendes display only marginally a relationship to their host rock Al_2O_3 contents; however, there is a clear trend in felsic gneisses for Al in hornblendes to increase markedly with increased retrogression (Fig. 3-24). Al contents from the retrograded mafic gneiss in contrast have slightly lower Al content than their non-retrograded equivalents.

No clear relationship exists between the $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratio in hornblendes compared to their host rocks $\text{Fe}_2\text{O}_3/(\text{Fe}_2\text{O}_3 + \text{FeO})$ ratio (Fig 3-25).

III.2-5 Biotite

Biotites have been analyzed from granulites at all localities (A11aG, B5cG, C7dG, D9hG, E15bG and F18yG). Biotite in these samples appears to be of retrograde origin. Biotites from moderately retrograded gneisses were analyzed at localities B and C (B5pI and C7aI) and from completely retrograded gneisses at localities A, B, D, E and F (A11fA, B5mA, D9aA, E15cA and F18rA). Analyses are given in Appendix III.4.

Biotites have been calculated on the basis of 22 oxygen atoms. No estimate of the Fe^{3+} content is given due to the

Figure 3-18. $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in hornblendes versus $\text{MgO}/(\text{MgO} + \text{FeO})$ ratios in host rocks.

Figure 3-19. $\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+})$ ratios in hornblendes versus $\text{CaO}/(\text{CaO} + \text{MgO} + \text{FeO})$ ratios in host rocks.

Figure 3-20. $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in hornblendes versus $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in host rocks.

Figure 3-21. Normalized $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios in hornblendes versus modal proportions of hornblendes.

Figure 3-22. Ti contents of hornblendes versus TiO_2 contents of host rocks.

FIG 3-18

FIG 3-19

FIG 3-20

FIG 3-21

FIG 3-22

FIG 3-23

FIG 3-24

Figure 3-23. Ti contents of hornblendes versus modal proportions of hornblendes.

Figure 3-24. Al contents of hornblendes versus Al_2O_3 contents of host rocks.

Figure 3-25. $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ ratios in hornblendes versus $Fe_2O_3 + FeO$ ratios in host rocks.

FIG 3-25

common occurrence of vacancies in the biotite structure, particularly Ti-rich biotites (Dymek, 1983). Wet-chemical analyses of biotites from an amphibolite facies to granulite facies transition in western Australia give $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratios of 0.05-0.09 for biotites from granulite grade rocks and 0.07-0.16 for amphibolite grade biotites (Stephensen, 1977). Similar values are expected for the biotites studied here.

As with the other ferromagnesian minerals, the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratio in biotite appears to be related to their host rock. $\text{MgO}/(\text{MgO} + \text{FeO})$ ratios (Fig. 3-26). The $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios in biotite increases in K feldspar gneiss and mafic gneiss but decreases in felsic gneisses as the degree of retrogression increases.

The $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios of biotite clearly do not have any relationship to the $\text{MnO}/(\text{MgO} + \text{FeO} + \text{MnO})$ ratios in the host rocks (Fig. 3-27). Biotites at most localities show a marked increase in their $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios with increasing retrogression. This increase also coincides with a modal increase in the amount of biotite (Fig. 3-28). Localities C and E have decreasing $\text{Mn}/(\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ ratios with a modal increase in biotite.

Biotites have a high Ti content and no correlation is evident between host rock TiO_2 contents and the Ti contents of biotite (Fig. 3-29). Ti contents of biotite decrease in all cases with increased retrogression and this is clearly correlated to a modal increase in the amount of biotite present (Fig. 3-30).

III.2-6 Plagioclase

Plagioclases were analyzed in granulites from all localities (AllaG, B5cG, C7dG, D9hG, E15bG and F18aG), in moderately retrograded gneisses from localities B and C (B5pI and C7aI) and in strongly retrograded gneisses at localities A, B, D, E and F (AllfA, B5mA, D9aA, E15cA and F18sA). Analyses are given in Appendix III.5.

An contents of plagioclases range from 15 to 37 mole%. An contents display no consistent trends with increased retrogression (Figs. 3-31, 3-32). No relationship with Fe content or K-feldspar content is apparent. Overall the proportion of plagioclase relative to ferromagnesian minerals is high and slight variations in plagioclase composition upon retrogression may be masked by errors in the analyses.

Additional analyses of K feldspar and magnetite are given in Appendix IV.

Figure 3-31. Anorthite content versus Fe abundances in plagioclases.

Figure 3-32. Anorthite content versus orthoclase content in plagioclases.

III.3 Mineral Reactions

III.3-1 Graphical Representation

It was demonstrated in III.2 that the composition of the minerals garnet, hornblende and biotite change as the degree of retrogression increases. These changes can not be accounted for by variations in the composition of the host rocks, indicating that the variations in mineral compositions are due to reactions in response to changing external conditions.

It is convenient to examine the variations in mineral composition using graphical representation in order to gain insight into mineral reactions. Minerals in granulite facies gneisses can be approximately represented in the system $\text{SiO}_2 - \text{TiO}_2 - \text{Al}_2\text{O}_3 - \text{Fe}_2\text{O}_3 - \text{FeO} - \text{MnO} - \text{MgO} - \text{CaO} - \text{Na}_2\text{O} - \text{K}_2\text{O} - \text{H}_2\text{O}$. Phase relations may be shown in terms of four components:

$$A = \text{Al}_2\text{O}_3 - (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$$

$$F = \text{FeO} - (\text{Fe}_2\text{O}_3 + \text{TiO}_2)$$

$$M = \text{MgO}$$

$$N = \text{MnO}$$

provided that quartz, magnetite, ilmenite, K feldspar and plagioclase of constant composition occur in the rocks and the pressure, temperature and chemical potential of H_2O are fixed (Reinhardt and Skippen, 1970; Froese, 1978).

Mineral and bulk rock compositions may be plotted on a diagram of $A/(F+M)$ versus $M/(F+M)$, similar to the representation of Stout (1972). Minerals with MnO content would plot above this AFM plane. At constant pressure, five minerals within the

AFMN system are related by a discontinuous reaction and four minerals, defining a subtetrahedron, represent a continuous reaction.

All the rocks in this study contain quartz, magnetite, K feldspar and plagioclase. Traces of ilmenite were found only sporadically, but the main Ti-bearing phases biotite and hornblende have Ti contents which are similar to those found in amphibolite facies terranes in which ilmenite is present (Dymek, 1983; Raase, 1974). This suggests that all the minerals under study are nearly saturated in Ti with respect to ilmenite and that the component $\text{TiO}_2\cdot\text{FeO}$ may be considered an excess component (Reinhardt, 1968).

Although the compositions of orthopyroxene, clinopyroxene, hornblende and biotite differ from one locality to another, their relative positions on the AFM diagram are similar and consistent with the topology of Reinhardt and Skippen (1970) and Froese and Jen (1979) neglecting MnO as a component. They plot slightly above the AFM plane and are shown projected through MnO. However, the relative position of hornblende is controversial, reflecting the uncertainty of the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio. With a reasonable assumption of Fe_2O_3 content, it lies to the MgO side of the orthopyroxene-clinopyroxene tie line. A schematic diagram illustrating some of the pertinent phase relations observed in the rocks studied is given in Figure 3-33. Garnet is not plotted since it contains a significant proportion of spessartine component and would plot considerably above the diagram.

The position of one discontinuous reaction can be shown based on the assumption that orthopyroxene, clinopyroxene and garnet,

$A = \text{Al}_2\text{O}_3 - (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$

$F = \text{FeO} - (\text{Fe}_2\text{O}_3 + \text{TiO}_2)$

$M = \text{MgO}$

$N = \text{MnO}$

(plotted above
AFM plane)

- P constant
- + quartz
- + K feldspar
- + plagioclase of constant composition
- + magnetite
- + ilmenite

○ Bulk Rock Composition

TEMPERATURE →

Figure 3-33. Representative AFM diagrams of mineral assemblages.

produced by granulite facies metamorphism, equilibrated with the retrograde minerals hornblende and biotite. The coexistence of these minerals, in the presence of quartz, K feldspar, plagioclase of constant composition, magnetite and ilmenite, implies the reaction:

In the AFMN tetrahedron, this reaction is represented by an exchange of the orthopyroxene-clinopyroxene-biotite tie plane for the hornblende-garnet tie line. In order to derive the correct stoichiometry, the reaction was balanced by means of a computer program made available by J.A. Percival, Geological Survey of Canada. The compositions of the ferromagnesian minerals in sample F18yG, representing the initial stages of retrogression were used. The ratio $\text{Fe}_2\text{O}_3/\text{FeO}$ (in weight %) was assumed to be 0.1 and 0.2 in clinopyroxene and hornblende, respectively. These values are based on wet chemical analyses reported by Jen and Kretz (1981). Orthopyroxene, garnet and biotite were treated as ferrous phases. Cr_2O_3 was neglected. The balanced reaction is:

$$(3.2) \quad 1.501 \text{ Orthopyroxene} + 2.484 \text{ Clinopyroxene} + 0.368 \text{ Biotite} \\ + 0.709 \text{ Albite} + 1.047 \text{ Anorthite} + 0.339 \text{ Magnetite} \\ + 0.05 \text{ Ilmenite} + 1.00 \text{ H}_2\text{O} = \\ 1.736 \text{ Hornblende} + 0.173 \text{ Garnet} + 0.175 \text{ K feldspar} + 2.071 \text{ Quartz}.$$

III.3-2 Felsic Gneiss

Petrography suggests that the high-grade assemblage orthopyroxene + clinopyroxene + garnet is the stable initial

assemblage in these rocks, while hornblende + biotite is the stable retrograde assemblage. At granulite facies conditions, the composition of felsic gneisses plots on orthopyroxene-clinopyroxene-garnet tie planes. During retrogression, the subtetrahedron orthopyroxene-clinopyroxene-garnet-biotite shifts towards Fe-richer compositions causing the bulk composition to move into the subtetrahedron; biotite appears as an additional mineral. Reaction 3.1 leads to the disappearance of orthopyroxene and the formation of hornblende. Further retrogression causes a shift of the clinopyroxene-garnet-biotite-hornblende subtetrahedron to Mg-richer compositions, the bulk composition moves out of this subtetrahedron and clinopyroxene disappears. A further increase in water activity allows hornblende and biotite to incorporate greater amounts of MnO and these minerals plot progressively higher above the AFM plane. Consequently, the bulk composition moves from a hornblende-biotite-garnet tie plane into a volume of hornblende-biotite tie lines. This change cannot be shown on the AFM face in Figure 3-33. However, it can be demonstrated by a mass balance that a rock composition represented by the granulite assemblage orthopyroxene-clinopyroxene-garnet can be expressed by the retrograde assemblage biotite-hornblende. In order to do this, the compositions of the ferromagnesian minerals orthopyroxene, clinopyroxene and garnet in sample F18yG, and hornblende and biotite in sample F18rA, were used. The $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios used in clinopyroxene and hornblende are the same as above. The mass balance is:

It should be noted that there are a number of possible sequences of mineral assemblages leading to the decomposition of the pyroxenes. The sequence will depend on slight compositional differences, i.e. on whether the bulk composition falls into the subtetrahedron clinopyroxene-biotite-hornblende-garnet, orthopyroxene-biotite-hornblende-garnet or orthopyroxene-hornblende-clinopyroxene-garnet. Because of the small volume occupied by these subtetrahedra, rocks with small compositional differences may display different sequences. Limited documentation of intermediate stages of retrogression makes it difficult to determine the exact sequence for a particular rock.)

The Ti contents of biotite and hornblende are quite high (see III-2). This is the reverse of what is commonly expected for hornblendes and biotites formed during retrograde reactions involving pyroxenes (Sen and Oliver, 1981). Only the initial hornblendes in B5pI are low in Ti. In order to explain this feature, a source of Ti is needed. Pyroxenes commonly have magnetite associated with them (ilmenite is rare) indicating that Ti in biotite and hornblende may be derived from magnetite which contains some Ti in solid solution (see Appendix IV).

The changes in these rocks can, therefore, be viewed as a series of mainly continuous reactions in response to increased water activity. These are hydration reactions and require a source of water.

III.3-3 K feldspar Augen Gneiss

Petrography suggests that in the high grade assemblage of the K feldspar augen gneisses orthopyroxene is the only primary ferromagnesian mineral, while the retrograde assemblage is biotite + garnet + hornblende. This change in mineral assemblage takes place over a few centimetres, suggesting that the reactions are extremely sensitive to changes in the activity of water. Bulk rock compositions for these rocks plot in the orthopyroxene field on the AFM diagram (Fig. 3-33). A shift of the orthopyroxene-biotite-hornblende-garnet subtetrahedron to Fe-richer compositions during retrogression causes the bulk composition to fall into the subtetrahedron leading to the appearance of biotite-hornblende-garnet. The subtetrahedron eventually overrides the bulk composition and orthopyroxene disappears from the assemblage. Newly-formed garnets have high MnO contents derived mainly from the decomposition of orthopyroxene.

III.3-4 Mafic Gneiss

The high grade assemblage of the mafic gneiss is hornblende + garnet + orthopyroxene + clinopyroxene. The retrogression of these gneisses took place by reactions which are similar to those proposed for the felsic gneisses with the retrograde assemblage being hornblende + garnet + biotite. The preservation of garnet in the retrograde assemblage is most likely due to the high abundance of garnet in the original rock and to the inability

ty of hornblende (and to some extent biotite) to accept the MnO content of the garnet into its structure.

Chapter IV

Conditions of Metamorphism

IV.1 Phase Relations

Experimentally determined phase relations for rocks equilibrated under upper-amphibolite facies to granulite facies conditions are few and are generally limited to rocks having a "basaltic" or a "pelitic" composition. Experiments on "charnockitic" compositions have not been attempted and, therefore, phase relationships in these compositions must be extrapolated from those experimentally determined in mafic and pelitic systems. Most of the data pertinent to this study are derived from experimental equilibria in mafic rocks. Phase relations are strongly dependent on variables such as fluid pressure, total pressure, oxygen fugacity and whole rock composition. Some phase equilibria useful to this study are given in Figure 4-1.

Equilibria in granulite facies rocks include:

H_2O (Luth, 1967);

+ Plagioclase + Ilmenite + H_2O (Spear, 1981a);

$\text{Amphibole}_2 + \text{Plagioclase}_2 + \text{Clinopyroxene} + \text{Orthopyroxene} +$

$\text{Ilmenite} + \text{H}_2\text{O}$ (Spear, 1981a); and

Reactions 4.2, 4.3 and 4.4 are shown in Figure 4.1 on the QFM buffer. At $P_{\text{fluid}} = P_{\text{total}}$ and $P_{\text{total}} > 200$ MPa, equilibrium 4.1 is intersected by melting reactions at approximately 850°C and reactions 4.2, 4.3 and 4.4 take place at approximately 950°C, 810°C and 790°C, respectively. If P_{fluid} is less than P_{total} , these equilibria are shifted strongly to lower temperatures. At 700°C and $P = 200$ MPa, the above reactions are stable under the following conditions: (4.1) $X_{\text{H}_2\text{O}} \approx 0.2$ (Hansen et al., 1984); (4.2) $X_{\text{H}_2\text{O}} \approx 0.1$ (Spear, 1981a); (4.3) $X_{\text{H}_2\text{O}} \approx 0.4$ (Spear, 1981a); (4.4) $X_{\text{H}_2\text{O}} \approx 0.5$ (Spear, 1981a).

The pressure sensitive reaction

(4.5) $\text{Anorthite} + 2 \text{Enstatite} = \text{Pyrope} + \text{Diopside} + \text{Quartz} \quad (G)$ divides the granulite field. This reaction was initially studied by Green and Ringwood (1976) and further constrained at lower pressure and temperature conditions by Hansen (1981) to be at 1340 - 1400 MPa at 900°C with a slope of 0.8 MPa/°C. Addition of Fe to this system shifts the reaction to lower pressures.

Experimental equilibria of mafic compositions under upper-amphibolite facies conditions include the disappearance of titanite according to the reaction:

and the disappearance of chlorite by reactions such as:

(4.8) Chlorite + Albite + Titanite = Hornblende (F)
 + Plagioclase + Ilmenite (Moody et al., 1983).

Reaction 4.6 is strongly dependent on oxygen fugacity and fluid pressure. On the QFM buffer, titanite disappears at $P_{\text{fluid}} = P_{\text{total}} = 400 \text{ MPa}$ at 700°C and 200 MPa at 600°C (Spear, 1981a). Chlorite disappears from the assemblage at approximately 575°C where $P_{\text{fluid}} > 200 \text{ MPa} = P_{\text{total}}$ (Liou et al., 1974; Moody et al., 1983).

The occurrence of orthopyroxene and/or clinopyroxene and/or garnet in the unaltered to weakly retrograded gneisses at all localities suggests that peak metamorphic conditions were at temperatures of less than 1000°C and at pressures below 1250 MPa . The lack of chlorite in all retrograde assemblages and the occurrence of titanite at locality F suggests that retrograde temperatures were in excess of 575°C and that $P_{\text{fluid}} > 200 \text{ MPa}$. The fact that there is no sign of melting accompanying retrogression indicates that temperatures of retrogression were below about 650°C based on melting in the granitic system at $P_{\text{fluid}} = P_{\text{total}}$ (Luth et al., 1964)(I). Metamorphic conditions can be further constrained using geothermobarometric methods.

Figure 4-1. Experimentally determined phase equilibria at $P_{H_2O} = P_{total}$: (A) Phlogopite + Quartz = Enstatite + Sanidine + H_2O (Luth, 1967); (B,C,D,E) Amphibolite stability (Spear, 1981a); (B) Amphibole out; (C) Orthopyroxene in; (D) Clinopyroxene in; (E) Titanite out; (F) Chlorite out (Moody et al., 1983); (G) Anorthite + 2 Enstatite = Pyrope + Diopside + Quartz-(with addition of Fe) (Hansen, 1981); (H) Al_2SiO_5 relations (Holdaway, 1971); (I) Beginning of melting in the granite system (Luth et al., 1964); (P) Calculated peak metamorphic conditions from this study; (R) Calculated retrograde metamorphic conditions from this study

IV.2 Geothermobarometry

IV.2-1 Peak Conditions

IV.2-1A Garnet-Clinopyroxene Thermometry

The occurrence in apparent equilibrium of garnet and clinopyroxene in the felsic and mafic gneisses allows the use of calibrated equilibria between garnet and clinopyroxene to estimate the temperature of peak metamorphism. A number of calibrations, based on the exchange of Fe - Mg between garnet and clinopyroxene, are presently available. These include those of Rheim and Green (1974), Ellis and Green (1979), Ganguly (1979) and Dahl (1980). Results from applying these calibrations to four localities are given in Table 4-1 and are based on the calculated Fe^{2+} values and a total pressure of 800 MPa (see later). Calculations in this and in the following sections are based on average mineral compositions from a single thin section at each locality.

The Ganguly (1979) calibration yields the highest temperatures and the Rheim and Green (1974) calibration the lowest. The Ellis and Green (1979) calibration is generally considered to give the most reasonable values for granulites when systematically tested against other thermometers (Johnson et al., 1983) and, therefore, has been preferred in the present study. The application of this thermometer gives temperatures of 703C for locality A, 630 for locality B, 719C for locality C and 742C for locality F (sample F18aG). When applied to the slightly more retrograded sample F18yG at locality F a value of 680C is obtained

suggesting that this thermometer is sensitive to minor changes in mineral composition as a result of retrogression such that the results above are probably minimum values. The effect of pressure is minimal on this thermometer increasing or decreasing temperatures by only 3°C if the pressure estimate is in error by 100 MPa. If the assumption that all Fe is Fe^{2+} is adopted, then values increase in felsic gneisses by less than 15°C and by 50°C in the mafic gneiss.

IV.2-1B Garnet-Orthopyroxene Thermometry

The occurrence of garnet and orthopyroxene in apparent equilibrium at localities A, B and F allows the use of Fe-Mg exchange thermometers calibrated for these two minerals. Three calibrations have been applied: Powell (1978), Harley (1984a) and Sen and Bhattacharya (1984). Results are given in Table 4-2 using calculated Fe^{2+} values at an assumed pressure of 800 MPa.

These calibrations have not been systematically tested against other thermometers to the extent of the garnet-clinopyroxene calibrations, and as a result, the calibration of Harley (1984a) will be adopted since it gives values intermediate to the other calibrations. Temperatures are calculated to be 630°C for locality A, 620°C for locality B and 707°C for locality F (sample F18aG). The slightly retrograded sample F18yG (locality F) yielded a result of 731°C in contrast to the trend suggested by garnet-clinopyroxene thermometry. If the pressure estimate is in error by ± 100 MPa, then the temperatures are only in error by $\pm 5^{\circ}\text{C}$. When the assumption Fe total = Fe^{2+} is made, calcu-

Table 4-1 Garnet-Clinopyroxene Thermometry Based on a Pressure of 800 MPa

Calibration (Sample)	Locality A (AllaG)	Locality B (B5cG)	Locality C (C7dG)	Locality F (F18aG)
Ellis and Green (1979)	703	630*	719	742
T-Max Ganguly (1979)	816	751	831	865
T-Min Ganguly (1979)	798	734	813	846
Råheim and Green (1974)	643	579	661	696
Dahl (1980)	732	629	851	901

Table 4-2 Garnet-Orthopyroxene Thermometry Based on a Pressure of 800 MPa

Calibration (Sample)	Locality A (AllaG)	Locality B (B5cG)	Locality F (F18aG)
Powell (1978)	601	603	639
Harley (1984a)	639	620	707
Sen and Bhattacharya (1984)	697	681	795

lated temperatures are virtually identical, except for the mafic gneiss where it is some 50°C higher.

IV.2-1C Two Pyroxene Thermometry

Temperatures based on the shape of the solvus between Ca-poor and Ca-rich pyroxenes have long been considered unreliable when applied to granulite facies rocks (Bohlen and Essene, 1979). Below 850°C the limbs of this solvus are steep and minor errors in experimental data and/or analyses result in scattered temperature values when using this method. Application of the Wood and Banno (1973) and Wells (1977) calibrations yielded temperatures of 800 - 900°C for localities A and F. Values between 700 - 800°C are recorded for locality B and the slightly retrograded sample F18yG at locality F. Lindsley's (1983) graphical representation of the solvus yielded values between 800 - 1000°C at a pressure of 1000 MPa for clinopyroxenes and between 500 - 600°C for orthopyroxenes.

Recently, Kretz (1982) has proposed a thermometer based on the Fe - Mg distribution between Ca-poor and Ca-rich pyroxenes as well as a refinement for the clinopyroxene slope of the solvus which allows a comparison of the two methods. Values for localities A and F are between 640 - 660°C based on the solvus version and between 750 - 800°C for the Fe - Mg exchange version. Values obtained from locality B and the slightly retrograded sample F18yG are considerably lower, being 540 - 560°C for the solvus version and 540 - 570°C for the Fe - Mg exchange version.

These results suggest that pyroxenes are extremely

sensitive to minor amounts of retrogression and have been variably reset. The pyroxene solvus appears to be slightly more sensitive to this resetting than is the Fe - Mg exchange. The values obtained using two-pyroxene thermometry may be scattered; however, they clearly indicate that most of the pyroxenes have been reset from the peak of metamorphism to some degree and that thermometric and barometric calibrations using them record minimum values.

IV.2-1D. Summary of Thermometry.

The above methods suggest that peak metamorphic temperatures for localities A, C and F were between 700 - 750°C. Felsic gneisses at locality B gave somewhat lower peak temperatures in the order of 600 - 650°C, comparable to the temperatures calculated for the slightly retrograded sample F18yG at locality F, suggesting significant re-equilibration of mineral compositions as a result of retrogression. Mineral analyses from a non-retrograded mafic boudin at locality B (sample B3aG, mineral compositions in Appendices II.1-5) confirm this as temperatures recorded by the Ellis and Green (1979), Harley (1984a) and Kretz (1982) thermometers for this locality gave values closely clustered around 750°C. No methods are currently available to estimate peak metamorphic temperatures at localities D and E; however, a similar temperature near 750°C is likely. Calculated temperatures are shown in Figure 4-1.

These values are comparable to those obtained by

Anovitz and Essene (1986) from the Huntsville region; however, they tend to be on the low end of their 750 - 800°C estimates.

IV.2-1E Garnet-Pyroxene-Plagioclase-Quartz Barometry

Perkins and Newton (1981) and Newton and Perkins (1982) have presented two calibrated reactions that can be used as geobarometers:

- (4.9) Anorthite + Diopside = $2/3$ Grossular + $1/3$ Pyrope + Quartz
 (4.10) Anorthite + Enstatite = $1/3$ Grossular + $2/3$ Pyrope + Quartz.

Bohlen et al. (1983) experimentally calibrated the analogous reaction:

- (4.11) Anorthite + Ferrosilite = $1/3$ Grossular
 + $2/3$ Almandine + Quartz.

Pressures were calculated using calculated Fe^{2+} values. Site occupancy in pyroxenes were estimated assuming an ideal, two-site mixing model (Wood and Banno, 1973). The following activity models were used for pyroxenes: diopside: $X_{\text{Ca}}(M_2) \cdot X_{\text{Mg}}(M_1)$, enstatite: $X_{\text{Mg}}(M_1) \cdot X_{\text{Mg}}(M_2)$, ferrosilite: $X_{\text{Fe}^{2+}}(M_1) \cdot X_{\text{Fe}^{2+}}(M_2)$. Activities for grossular and pyrope were calculated assuming the interaction parameters of Newton and Perkins (1982) and the activity of almandine was calculated as in Bohlen et al., (1983). Perkins and Chipera (1985a) suggested using the garnet activity models of Ganguly and Saxena (1984a,b). These models are complex and calculations using Perkins and Chipera's (1985a) simplified barometry results in estimates which cluster near the values given here. Anorthite activities were calculated using the Al-avoidance model of Newton (1983). All

calculations were carried out at 700°C. Determined values are given in Table 4-3 for localities A, B, C and F. Values in brackets are empirical corrections of +150 MPa for reaction 4.9 and -50 MPa for reaction 4.10, which were suggested by Newton (1983).

Pressures calculated using these three calibrations and empirical corrections cluster between 770 - 820 MPa at locality A, 750 - 860 MPa at locality B, 920 - 930 MPa at locality C and 790 - 850 MPa at locality F. The effect of the slight amount of retrogression in sample Fl8yG is minimal (orthopyroxene pressures using 4.10 increase and 4.11 decrease by 50 MPa). If the assumption $\text{Fe total} = \text{Fe}^{2+}$ is adopted, then pressures are increased by less than 40 MPa. The relatively high pressures obtained at locality C are probably the result of a long extrapolation to end member compositions in the relatively Fe-rich and Mg-poor rocks at this locality. Calculated pressures are similar at all localities suggesting that they equilibrated under similar conditions of 800 MPa at 700°C. Pressures increase by 200 MPa for every 100°C in temperature, such that at 750°C the calculated pressure would be 900 MPa.

IV.2-1F Other Barometers

The pressure sensitive solubility of aluminum in pyroxene has led to the development of barometers based on the reactions:

Table 4-3 Garnet-Pyroxene-Plagioclase-Quartz Barometry
Based on a Temperature of 700°C

Calibration (Sample)	Locality A (Al1aG)	Locality B (B5cG)	Locality C (C7dG)	Locality F (F18aG)
Cpx-Newton and Perkins (1982)	701 (851)	723 (873)	799 (949)	647 (797)
Opx-Newton and Perkins (1982)	857 (807)	806 (756)	- -	892 (842)
Bohlen et al. (1983)	843	853	-	798

Values in brackets empirical correction of Newton (1983)

(4.12) Enstatite + Mg Tschermak's component (MgTs in opx)

= Pyrope; and

(4.13) Anorthite = Ca Tschermak's component (CaTs in cpx)

+ Quartz

(Wood, 1974, 1979). Calibrations of these barometers, when applied to the samples in this study, gave inconsistent results. Pressure estimates range from 510 - 1320 MPa using the Wood (1974) calibration, 700 - 1200 MPa using the Harley and Green (1982) calibration and 500 - 1600 MPa using the Harley (1984b) calibration of reaction 4.12. Pressures using the Ellis (1980) calibration of reaction 4.13 range from 970 - 1350 MPa. Calculations were carried out at 700°C using the activity models and site occupancies as in the previous section. MgTs and CaTs activities were calculated as in Wood and Banno (1973) and Wood (1979), with MgTs: $X_{Mg}(M_2) \cdot X_{Al}(M_1)$ and CaTs: $4X_{Ca}(M_2) \cdot X_{Al}(M_1)$. Inconsistencies in the pressures determined by these methods may be traced to the low Al contents of the pyroxenes so that slight errors in analyses and uncertainties in site assignments, especially Al in M_1 , are responsible for large differences in pressure.

IV.2-1G Summary of Barometry

Pressures of peak metamorphism as suggested by garnet-pyroxene-plagioclase-quartz equilibria appear to be the most reliable. Based on this, pressure estimates are similar for all localities, being 800 MPa at 700°C and near 900 MPa at 750°C.

Anovitz and Essene (1986) reported pressures to be near 1000 MPa for this region; however, their determinations were made at assumed temperatures closer to 800°C. It was shown earlier that the temperatures calculated here are sensitive to the resetting of mineral compositions by minor amounts of retrogression. If a temperature of 800°C were used in the barometric calculations performed here, a pressure of approximately 1000 MPa would result.

IV.2-2 Retrograde Conditions

IV.2-2A Garnet-Biotite Thermometry

The occurrence of biotite and garnet in apparent equilibrium in the retrograde assemblage at localities A, D and E permits an estimation of retrograde temperatures using the exchange equilibrium of Fe and Mg between garnet and biotite. Of the numerous calibrations present, four have been applied here (Table 4-4).

Thompson's (1976) empirical thermometer gave temperatures which range from 575 - 650°C. Temperatures using Ferry and Spear's (1978) calibration are slightly higher with values being 693°C for locality A, 603°C for locality D and 653°C for locality E assuming a pressure of 600 MPa (see later). Temperatures vary only by 4°C for every 100 MPa variation in assumed pressure. The effect of using Fe total = Fe^{2+} vs calculated Fe^{2+} values is minimal, as calculated Fe^{3+} values are low in garnets. The Indares and Martignole (1985) calibration is essentially a cor-

rection applied to the Ferry and Spear (1978) thermometer, taking into account the effects of Al and Ti in biotite. Their corrections decrease the calculated temperatures relative to the Ferry and Spear (1978) thermometer. Perkins and Chipera (1985b) suggested that the experimental calibration of Perchuk and Lavrent'eva (1983), which takes into account the Mn content of garnet and biotite, yields the most accurate results when applied to granulites. This thermometer gave temperatures of 606°C for locality A, 548°C for locality D and 609°C for locality E.

IV.2-2B Amphibole-Plagioclase Equilibria

The occurrence of hornblende and plagioclase in the retrograde assemblage in apparent equilibrium at localities A, B, C and F allows the use of amphibole-plagioclase equilibria to estimate temperatures and pressures of retrogression (Table 4-5).

Spear (1980, 1981b) formulated empirical thermometers using exchange equilibria between amphibole and plagioclase based on the reactions:

These thermometers are dependent on the amount of Na assigned to the A-site of the amphibole, which can only be approximated in this study. Amphibole formulae were calculated as in III.2-4. Temperatures are estimated to be between 500 - 550°C for localities B, C and F, whereas locality A gave a temperature between

600 - 650°C.

Plyusnina (1982) experimentally calibrated a thermobarometer based on the Al content of amphibole and the anorthite content of coexisting plagioclase. She suggested that this thermobarometer should be used only with epidote-bearing assemblages; however, she presented an example in which epidote is absent and garnet is present that gave reasonable results. Temperatures estimated by this method are similar to those based on Spear (1980, 1981b) and range from 520- 560°C, the highest temperatures being estimated for locality A. Pressures are estimated to be between 560 - 640 MPa for localities A, B and C. A value of 750 MPa is estimated for locality F. Estimated pressures are somewhat higher using the Hammarstrom and Zen (1986) calibration, initially derived for amphiboles crystallized from melts, which is based on the Al content of amphibole. Calculated pressures range from 660 - 704 MPa.

IV.2-2C Summary of Thermometry and Barometry

Temperatures of retrogression estimated from garnet-biotite thermometry are somewhat higher than those estimated from amphibole-plagioclase equilibria. Comparison of temperatures estimated by both methods for locality A gave a temperature range of 560 - 700°C. Spear's (1980) calibration gave a value between 600 - 650°C, suggesting that amphibole-plagioclase equilibria based on Spear's (1980, 1981b) methods are consistent, but may underestimate temperatures in the order of 50-100°C relative to the more extensively tested garnet-biotite thermometers. Based

Table 4-4 Garnet-Biotite Thermometry Based on a
Pressure of 600 MPa

Calibration (Sample)	Locality A (Al1fA)	Locality D (D9aA)	Locality E (E15cA)
Thompson (1976)	643	578	615
Ferry and Spear (1978)	693	603	653
Indares and Martignole (1985)	655	626	580
Perchuk and Lavrent'eva (1983)	606	548	609

Table 4-5 Amphibole-Plagioclase Equilibria

Calibration (Sample)	Locality A (Al1fA)	Locality B (B5mA)	Locality C (C7aI)	Locality F (F18sA)
Temperatures (°C)				
Spear (1980)	650	520	510	510
Spear (1981b)	600	500	550	500
Plyusnina (1982)	560	530	530	520
Pressures (MPa)				
Plyusnina (1982)	600	640	560	750
Hammarstrom and Zen (1985)	660	704	688	660

on this assumption, temperatures of retrogression are estimated to be 600 - 700°C for locality A and 550 - 650°C for localities B, C, D, E and F.

Estimates of the pressure under which retrogression occurred are somewhat tenuous due to the poorly tested nature of the currently formulated methods. Results based on an average of the two methods used suggest that pressure was between 600 - 700 MPa. Temperatures and pressures estimated from thermometry and barometry are shown in Figure 4-1.

IV.3 Summary of the Conditions of Metamorphism

Experimentally determined phase relations and geothermobarometric methods suggest that the peak metamorphic conditions were at temperatures of 700 - 750°C and at pressures of 800 - 900 MPa. At these temperatures and pressures the aluminosilicate polymorph kyanite is expected to be locally present in the area. Isolated occurrences have been reported by Anovitz and Essene (1986). In order for orthopyroxene and clinopyroxene to be stable under these conditions P_{fluid} must be less than P_{total} . For the complete dehydration of amphibole at 750°C, $X_{\text{H}_2\text{O}} < 0.4$ based on experimental work by Spear (1981a).

Retrogression is estimated to have occurred at temperatures near 600°C for most localities and possibly as high as 700°C at locality A, and at pressures of 600 - 700 MPa (temperatures must be above 600°C, because chlorite is not present, and below 700°C, because there is no melting associated with retrogression). Under these conditions P_{fluid} is approximately 0.5

P_{total} based on the occurrence of titanite at locality F and $X_{\text{H}_2\text{O}}$
> 0.4 based on the absence of pyroxenes in the retrograde assem-
blage (Spear, 1981a).

Chapter V

Discussion

V.1 The Process of Retrogression

V.1-1 Source of Retrograde Fluids

Retrogression refers to the process whereby a metamorphic assemblage is altered at some time after the peak of metamorphism, under lower grade conditions. This process commonly involves the formation of minerals which are more hydrated than the minerals from which they were derived, suggesting that it involves the participation of a fluid phase. This fluid phase may be the residual fluid which was present in the rock at the peak of metamorphism, that becomes combined into the retrograde minerals (Fyfe, 1976). However, it is probable that this fluid is limited in amount and has a patchy rather than pervasive distribution (Yardley, 1981). Although the extensive retrogression described in this study does not appear to be predominantly of this type, the initial stages of retrogression may be. The estimate of $X_{H_2O} < 0.4$ for the granulite facies gneisses is less than that of $X_{H_2O} > 0.4$ for the retrogression and suggests that an H_2O -rich fluid had to be introduced into the rocks.

The occurrence of retrogression around late deflection folds which commonly have pegmatites associated with them indicates that fluid was introduced along these zones. The temperatures estimated for retrogression of around 600 - 700°C are

consistent with minimum temperatures for the production of granitic liquids under H_2O - saturated conditions (Luth et al., 1964). These are the conditions under which most pegmatites are thought to form (Jahns and Burnham, 1969). Therefore it appears likely that most of the retrogression studied here was the result of fluids generated from the crystallization of the commonly associated pegmatite dikes which were intruded during uplift of the region when pressures were approximately 200 MPa less than those of peak metamorphic conditions. A similar setting for retrogression has been described on a larger scale by Shearer et al. (1986). Under such circumstances, it is expected that there would have been a gradient in X_{H_2O} from the non-retrograded to weakly-retrograded granulites toward the pegmatite dikes.

V.1-2 Mineral - Fluid Equilibria

Rocks are essentially dry after they pass through the peak of metamorphism, because the fluid content in them has been absorbed into initial retrograde minerals (Fyfe 1976, Yardley 1981). The peak metamorphic assemblages are, therefore, metastable at the lower grade conditions under which the influx of retrograde fluid occurs.

The first stages of retrogression in the rocks studied here involved the conversion of pyroxenes to hornblende and biotite. Thermodynamic data on the decomposition of orthopyroxene to biotite indicates that X_{H_2O} must be greater than 0.2 at 700°C and 600 MPa for retrogression to occur (Hansen et al.,

1984). Reaction of orthopyroxene to hornblende requires $X_{H_2O} > 0.4$ while reaction of clinopyroxene to hornblende requires $X_{H_2O} > 0.5$ (Spear, 1981a) under similar conditions.

It is possible to calculate a_{H_2O} - T conditions for reaction (3.1) shown in Figure 3-33 by considering equilibria among components in minerals taking part in the reaction. For instance, the coexistence of quartz, K feldspar, biotite and orthopyroxene implies the equilibrium

Activities were obtained from analyses of minerals in F18yG assuming ideal mixing on lattice sites:

$$a_{KMg_3Si_3AlO_{10}} = [K] [Mg/3]^3$$

$$a_{MgSiO_3} = [Mg]$$

The equilibrium was calculated by S.W. Adcock, Geological Survey of Canada, using the computer programs of Perkins et al. (1986) and the data base of Berman et al. (1985). The activities are true only for one point along the curve at the temperature of equilibration. At 700°C , the activity of water is about 0.1 (Fig. 5.1).

The decomposition of garnet, in the felsic gneisses, is probably much more complex than the decomposition of pyroxenes. The MnO enrichment of garnet as retrogression proceeds, in these rocks, is a good indicator that MnO does not enter the hornblende or biotite structure readily. Higher H_2O activities than those required for the decomposition of pyroxenes are probably required to distribute the MnO derived from garnet into hornblende and biotite. The disappearance of garnet restricted to the vicinity of pegmatite dikes, confirms that the pegmatite dikes were the

2

of/de

2

MicroD

probable source of the retrograde fluids, that these fluids were H_2O -rich, and that a gradient in X_{H_2O} existed.

V.1-3 Internal versus External Controls of Fluid Composition

A number of authors have suggested that a large amount of dissolved species, in the fluid phase of retrogression, may externally control the retrograde mineral composition (Beach, 1980; Sills, 1982; Yardley and Baltatzis, 1985). It has been shown in this study that the mineral composition of the retrograde phases is strongly dependent on external conditions rather than on variations in the host rock composition. This is particularly evident when comparing the MnO contents of hornblende, biotite and garnet with the whole rock composition of their host rocks. Yardley and Baltatzis (1985) have suggested that whole rock compositions are probably not changed during retrogression and that the fluid composition is internally controlled by the retrograde phases until a large volume of fluid has passed through the rock, at which time the rock may change in composition, which becomes externally controlled by the fluid phase and the dissolved species in it. It is strongly suggested here that the amount of introduced H_2O was the critical step in the retrogression and that little chemical change was produced by the fluid. The retrograde assemblages, therefore, internally controlled the fluid composition. It should be noted, however, that a fluid in equilibrium with granitic pegmatite is probably very similar to fluids which were in equilibrium with the chemically

Figure S-1.

Calculated activity of water versus temperature conditions for
 reaction 3.1 based on activities obtained from analyses of
 minerals in sample Fl8yG.

similar gneisses studied here. For this reason, a large chemical exchange between pegmatites and host gneisses is unlikely.

V.2 Variations in Mineral Composition across the Amphibolite Facies to Granulite Facies Transition

A number of grade changes in mineral composition have been proposed which characterize the prograde transition from the upper amphibolite facies to the granulite facies. It is worthwhile to compare the retrograde changes in mineral composition studied here with the proposed prograde changes. This discussion is limited to compositional changes of hornblende, biotite and garnet, as these minerals occur across the transitions discussed in the literature.

The most commonly reported changes in hornblende composition include increasing Ti and decreasing Mn content with rising temperature (Raase, 1974; Stephensen, 1977). Pressure-related changes include an increase in Al^{VI} and a decrease in Ti with increasing pressure (Raase, 1974; Hynes, 1982). The mechanisms by which these changes occur are not generally discussed and parageneses are not commonly reported. However, Stephensen (1977) suggested that Ti-enrichment and Mn-depletion in hornblende results from the decomposition of hornblende to form pyroxenes, with Ti being favoured by the hornblende structure and Mn being favoured by the pyroxene structure.

Similar grade-related changes have been proposed for biotite, whereby Ti increases and Mn decreases with increasing grade. Decomposition of biotite to form pyroxene is the most

commonly cited mechanism (Stephensen, 1977). Stephensen (1979) also suggested that the relative increase of garnet over biotite, with increasing grade, partitions Mn into the garnet structure while Ti is concentrated in the biotite.

Garnets are commonly considered to decrease in Mn and Ca while increasing in Fe and particularly Mg with increasing grade (Binns, 1969; Stephensen, 1979). Many of these changes are related to exchange reactions with other minerals in the paragenesis; however, the decrease in Mn is usually accounted for by a modal increase in garnet with increasing grade, Mn becoming diluted in this mineral as host rock Mn remains constant.

The systematic chemical changes observed in the minerals studied here indicate a decrease in Ti content and an increase in Mn content of hornblende and biotite as retrogression proceeded and as the modal proportion of these minerals increased, except in the K feldspar augen gneisses where Mn decreased. However, garnet is formed along with biotite in these gneisses and Mn is, therefore, partitioned selectively into this mineral. Al^{VI} contents of hornblende increased with increased retrogression in response to the decomposition of garnet rather than as a consequence of decreased pressure. The only clear trend in the composition of garnet is an increase in the Mn content of this mineral as retrogression proceeded and as the modal amount decreased.

Most of these changes can clearly be ascribed to reaction relationships amongst the minerals involved and are thus strongly related to paragenesis and the modal proportion of the minerals present. This is illustrated by the propensity of Ti to

favour hornblende and biotite, and Mn to favour garnet, such that variations in the modal proportions of these minerals greatly affect the mineral composition. Many compilations of grade-related changes in mineral composition ignore this point and broad generalizations given in these papers (i.e. increasing Al^{VI} with increasing pressure, in hornblende) should not be considered reliable. It is of interest to note that many of the changes in mineral composition found here are similar to those found in metabasic dikes described by Glassley and Sorensen (1980). They ascribe the changes in composition to be the result of prograde metamorphism.

Chapter VI

Conclusions

The following conclusions can be made concerning the retrogression of granulite facies gneisses from the Huntsville area:

1. The samples equilibrated, at the peak of metamorphism, at temperatures of 700 - 750°C and pressures of 800 - 900 MPa. Water activity was estimated to be < 0.4 . Retrogression occurred at temperatures between 600 - 700°C and pressures between 600 - 700 MPa. Water activity was estimated to be > 0.4 .

2. Changes in mineral composition are related to imposed conditions of pressure, temperature and water activity, because bulk rock compositions did not change significantly as a result of retrogression. The formation and decomposition of minerals can be modelled as mainly continuous reactions.

3. Retrogression occurred as the result of infiltration of the fluids generated by the crystallization of pegmatite dikes into the surrounding gneisses, as fluid gradients were formed around the pegmatite dikes. The amount of fluid which infiltrated these rocks was not high enough to alter the bulk rock compositions.

4. Comparison of the variations in mineral composition from published studies of prograde amphibolite facies to granulite facies transitions elsewhere, with those resulting from retrogression, suggests that water activity plays a more impor-

tant role than pressure and temperature.

. VI.1 Proposal For Future Work

This study has provided some reconnaissance information on the retrogression of granulite facies gneisses. A number of aspects have not been fully explored. These include:

1. A more comprehensive examination of fluid composition, including the determination of volatiles other than H_2O , an examination of variations in oxygen fugacity, and the analysis of fluid inclusions.
 2. The tectonic significance of zones of retrogression in the Huntsville area.
 3. The relationship between the width of the zones of retrogression and the source of the fluids causing retrogression.
-

REFERENCES

- Anovitz, L.M. and Essene, E.J., 1986, Thermobarometry in the Grenville Province, Ontario; G.A.C.- M.A.C., Program with Abstracts, v. 11, p. 41.
- Beach, A., 1974, Amphibolitisation of Scourian granulites; Scott. J. Geol., v. 10, p. 35-43.
- Beach, A., 1980, Retrogressive metamorphic processes in shear zones, with special reference to the Lewisian complex; J. Struct., Geol. v. 2, p. 257-263.
- Berman, R.G., Brown, T.H. and Greenwood, H.J., 1985, An internally consistent thermodynamic data base for minerals in the system $\text{Na}_2\text{O}-\text{K}_2\text{O}-\text{CaO}-\text{MgO}-\text{FeO}-\text{Fe}_2\text{O}_3-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{TiO}_2-\text{H}_2\text{O}-\text{CO}_2$; Atomic Energy of Canada Ltd. Technical Report 377, 62 p.
- Binns, R.A., 1969, Ferromagnesian minerals in high-grade metamorphic rocks, Geol. Soc. Aust. Spec. Publ. 2, p. 323-332.
- Bohlen, S.R. and Essene, E.J., 1979, A critical evaluation of two-pyroxene thermometry in Adirondack granulites; Lithos, v. 12, p. 335-345.
- Bohlen, S.R., Wall, V.J. and Boettcher, A.L., 1983, Experimental investigation and application of garnet granulite equilibria; Contr. Mineral. Petrol., v. 83, p. 52-61.
- Brown, G.C. and Fyfe, W.S., 1970, The production of granitic melts during ultrametamorphism; Contr. Mineral. Petrol., v. 28, p. 310-318.
- Buddington, A.F., 1963, Isograds and the role of H_2O in metamorphic facies orthogneisses of the northwest Adirondack area, New York; Bull. Geol. Soc. Amer., v. 74, p. 1155-1182.
- Culshaw, N.G., Davidson, A. and Nadeau, L., 1983, Structural subdivisions of the Grenville Province in the Parry Sound-Algonquin region, Ontario; in Current Research, Part B, Geological Survey of Canada, Paper 83-1B, p. 243-252.
- Dahl, P.S., 1980, The thermal-compositional dependence of Fe^{2+} -Mg distributions between coexisting garnet and pyroxene: applications to geothermometry; Am. Min., v. 65, p. 852-866.
- Davidson, A., 1984, Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield; in Kröner, A. and Greiling, R. (Eds.), Precambrian Tectonics Illustrated, E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, p. 263-279.

- Davidson, A., 1986, New interpretations in the southwestern Grenville Province; in Moore, J.M., Davidson, A. and Baer, A.J. (Eds.), The Grenville Province, Geological Association of Canada Special Paper 31, p. 61-74.
- Davidson, A., Culshaw, N.G. and Nadeau, L., 1982, A tectono-metamorphic framework for part of the Grenville Province, Ontario; in Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p. 175-190.
- Davidson, A., Culshaw, N.G. and Nadeau, L., 1984, Grenville traverse B: cross-section of the Central Gneiss Belt; Geological Association of Canada, Field Trip Guidebook 9B/10B, 79 p.
- Dymek, R.F., 1983, Titanium, aluminum and interlayer cation substitutions in biotite from high-grade gneisses, West Greenland; Am. Min., v. 68, p. 880-899.
- Ellis, D.J., 1980, Osumilite-sapphirine-quartz granulites from Enderby Land, Antarctica: P-T conditions of metamorphism, implications for garnet-cordierite equilibria and evolution of the deep crust; Contr. Mineral. Petrol., v.74, p.201-210.
- Ellis, D.J. and Green, D.H., 1979, An experimental study of the effect of Ca upon garnet-clinopyroxene exchange equilibria; Contr. Mineral. Petrol., v. 71, p. 13-22.
- Eskola, P., 1920, The mineral facies of rocks; Norsk. Geol. Tidsskr., v. 6, p. 143-194.
- Eskola, P., 1939, Die metamorphen Gesteine; in Barth, T.F.W., Correns, C.W. and Eskola, P. (Eds.) Die Entstehung der Gesteine, Berlin: Springer, p. 263-407.
- Ferry, J.M. and Spear, F.S., 1978, Experimental calibration of Fe and Mg exchange between biotite and garnet; Contr. Mineral. Petrol., v., p.113-117.
- Froese, E., 1978, The graphical representation of mineral assemblages in biotite - bearing granulites; in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, p. 323-325.
- Froese, E. and Jen, L.S., 1979, A reaction grid for biotite-bearing mafic granulites; in Current Research, Part A, Geological Survey of Canada, Paper 79-1A, p. 83-85.
- Fyfe, W.S., 1976, Chemical aspects of rock deformation; Royal Soc. of London Phil. Trans., v. A283, p. 221-228.

- Ganguly, J., 1979, Garnet and clinopyroxene solid solutions and geothermometry based on Fe-Mg distribution coefficient; *Geochim. Cosmochim. Acta* v. 43, p. 1021-1029.
- Ganguly, J. and Saxena, S.K., 1984, Mixing properties of aluminosilicate garnets: constraints from natural and experimental data, and implications to geothermo-barometry; *Am. Min.*, v. 69, p. 88-97.
- Ganguly, J. and Saxena, S.K., 1984, Mixing properties of aluminosilicate garnets: constraints from natural and experimental data, and implications to geothermo-barometry: clarifications; *Am. Min.*, v. 70, p. 1320.
- Glassley, W.E. and Sorensen, K., 1980, Constant P_g - T amphibolite to granulite facies transition in Agto (West Greenland) metadolerites: implications and applications; *J. of Petrol.*, v.21, p. 69-105.
- Green, D.H. and Ringwood, A.E., 1967, An experimental investigation of the gabbro to eclogite transformation and its petrological application; *Geochim. Cosmochim. Acta*, v.31, p. 767-833.
- Hammarstrom, J.M. and Zen, E-an, 1986, Aluminum in hornblende: An empirical igneous geobarometer; *Am. Min.*, v.71, p. 1297-1313.
- Hansen, B., 1981, The transformation from pyroxene granulite facies to garnet clinopyroxene granulite facies. Experiments in the system $\text{CaO} - \text{MgO} - \text{Al}_2\text{O}_3 - \text{SiO}_2$; *Contr. Mineral. Petrol.*, v. 76, p. 234-242.
- Hansen, E.C., Newton, R.C. and Janardhan, A.S., 1984, Fluid inclusions in rocks from the amphibolite-facies gneiss to charnockite progression in southern Karnataka, India: direct evidence concerning fluids of granulite metamorphism; *J. Metamorphic Geol.*, v.2, p. 249-264.
- Harley, S.L., 1984a, An experimental study of the partitioning of Fe and Mg between garnet and orthopyroxene; *Contr. Mineral. Petrol.*, v. 86, p. 359-373.
- Harley, S.L., 1984b, The solubility of alumina in orthopyroxene coexisting with garnet in $\text{FeO} - \text{MgO} - \text{Al}_2\text{O}_3 - \text{SiO}_2$ and $\text{CaO} - \text{FeO} - \text{MgO} - \text{Al}_2\text{O}_3 - \text{SiO}_2$; *J. of Petrol.*, v. 25, p. 665-696.
- Harley, S.L. and Green, D.H., 1982, Garnet-orthopyroxene barometry for granulites and peridotites; *Nature*, v. 300, p. 697-701.

- Holdaway, M.J., 1971, Stability of andalusite and the aluminium silicate phase diagram; *Am. J. Sci.*, v. 271, p. 97-131.
- Huebner, J.S., 1980, Pyroxene phase equilibria at low pressure; in Prewitt, C.T. (Ed.) *Pyroxenes*, Mineralogical Society of America, *Reviews in Mineralogy*, v. 7, p. 213-288.
- Hynes, A., 1982, A comparison of amphiboles from medium- and low-pressure metabasites; *Contr. Mineral. Petrol.*, v. 81, p. 119-125.
- Indares, A. and Martignole, J., 1985, Biotite-garnet geothermometry in the granulite facies: the influence of Ti and Al in biotite; *Am. Min.*, v. 70, p. 272-278.
- Jahns, R.H. and Burnham, C.W., 1969, Experimental studies of pegmatite genesis: I. A model for derivation and crystallization of granite pegmatites; *Econ. Geol.*, v. 64, p. 843-864.
- Jen, L.S. and Kretz, R., 1981, Mineral chemistry of some mafic granulites from the Adirondack Region; *Can. Min.*, v. 19, p. 479-491.
- Johnson, C.A., Bohlen, S.R. and Essene, E.J., 1983, An evaluation of garnet-clinopyroxene thermometry in granulites; *Contr. Mineral. Petrol.*, v. 84, p. 191-198.
- Kretz, R., 1982, Transfer and exchange equilibria in a portion of the pyroxene quadrilateral as deduced from natural and experimental data; *Geochim. Cosmochim. Acta*, v. 46, p. 411-421.
- Kretz, R., Hartree, R., Garrett, D. and Cermignani, C., 1985, Petrology of the Grenville swarm of gabbro dikes, Canadian Precambrian Shield; *Can. J. Earth Sci.*, v. 22, p. 53-71.
- Leake, B.E., 1978, Nomenclature of amphiboles; *Can. Min.*, v. 16, p. 501-520.
- Lindsley, D.H., 1983, Pyroxene thermometry; *Am. Min.*, v. 68, p. 477-493.
- Liou, J.G., Kuniyoshi, S. and Ito, K., 1974, Experimental studies of the phase relations between greenschist and amphibolite in a basaltic system; *Am. J. of Sci.*, v. 613-632.

- Luth, W.C., 1967, System $KAlSiO_4 - Mg_2SiO_4 - SiO_2 - H_2O$. I. Inferred phase relations and petrologic applications; *J. Petrol.*, v. 8, p. 372-416.
- Luth, W.C., Jahns, R.H. and Tuttle, O.F., 1964, The granite system at pressures of 4 to 10 kilobars; *J. of Geophys. Res.*, v. 69, p. 759-773.
- Moody, J.B., Meyer, D. and Jenkins, J.E., 1983, Experimental characterization of the greenschist/amphibolite boundary in mafic systems; *Am. J. of Sci.*, v.283, p. 48-92.
- Nadeau, L., 1985, Characterization of lithotectonic units and their boundaries in the Huntsville region, Central Gneiss Belt; in Davidson, A., Nadeau, L., Grant, S.M. and Pryer, L.L., 1985, Studies in the Grenville Province of Ontario; in Current Research, Part A, Geological Survey of Canada, Paper 85-1A, p. 464-474.
- Newton, R.C., 1983, Geobarometry of high-grade metamorphic rocks; *Am. J. of Sci.*, v. 283-A, p. 1-28.
- Newton, R.C., Smith, J.V. and Windley, B.F., 1980, Carbonic metamorphism, granulites and crustal growth; *Nature*, v. 288, p. 45-49.
- Newton, R.C. and Perkins, D.III, 1982, Thermodynamic calibration of geobarometers based on the assemblages garnet-plagioclase-orthopyroxene (clinopyroxene)-quartz; *Am. Min.*, v. 67, p. 203-222.
- Paquet, J., 1985, Petrology and geothermometry of retrograded Hercynian charnockites and host gneisses, Agly Massif, French Pyrenees; *J. Metamorphic Geol.*, v. 3, p. 43-58.
- Perchuk, L.L. and Lavrent'eva, 1983, Experimental investigation of exchange equilibria in the system cordierite - garnet - biotite; In Saxena, S.K. (ed.) Kinetics and equilibrium in mineral reactions; *Advances in Physical Geochemistry*, p. 199-239.
- Perkins, D.III and Chipera, S.J., 1985a, Garnet - orthopyroxene - plagioclase - quartz barometry: refinement and application to the English River subprovince and the Minnesota River valley; *Contr. Mineral. Petrol.*, v. 89, p. 69-80.
- Perkins, D.III and Chipera, S.J., 1985b, Evaluation of garnet - biotite geothermometers by trend surface analysis; applications to the English River subprovince, Ontario; *G.S.A. abstracts with programs*, 1985, p. 686.
- Perkins, D.III and Newton, R.C., 1981, Charnockite geobarometers based on coexisting garnet-orthopyroxene-plagioclase-quartz; *Nature*, v. 292, p. 144-146.

- Perkins, E.H., Brown, T.H. and Berman, R.G., 1986, PTX-SYTEM: three programs for calculation of pressure-temperature-composition phase diagrams; *Computers & Geosciences*, v. 12, p. 749-755.
- Plant, A.G. and Lachance, G.R., 1973, Quantitative electron microprobe analysis using an energy dispersive spectrometry; *Proceedings of the Eighth National Conference on Electron Probe Analysis*, New Orleans, Paper 13.
- Plyusnina, L.P., 1982, Geothermometry and geobarometry of plagioclase-hornblende bearing assemblages; *Contr. Mineral. Petrol.*, v. 80, p. 140-146.
- Powell, R., 1978, The thermodynamics of pyroxene geotherms; *Phil. Trans. R. Soc. London*, v. A288, p. 457-469.
- Raase, P., 1974, Al and Ti contents of hornblende: indicators of pressure and temperature of regional metamorphism; *Contr. Mineral. Petrol.*, v. 45, p.231-236.
- Råheim, A. and Green, D.H., 1974, Experimental determination of the temperature and pressure dependence of the Fe-Mg partition coefficient for coexisting garnet and clinopyroxene; *Contr. Mineral. Petrol.*, v. 48, p. 179-203.
- Ramsay, J.G., 1980, Shear zone geometry: a review; *J. Struct. Geol.*, v. 2, p. 83-99.
- Reinhardt, E.W., 1968, Phase relations in cordierite - bearing gneisses from the Gananoque area, Ontario; *Can. J. Earth Sci.*, v. 5, p. 455-482.
- Reinhardt, E.W. and Skippen, G.B., 1970, Petrochemical study of Grenville granulites, in *Report of Activities, Part B, Geological Survey of Canada*, Paper 70-1B, p. 48-54.
- Robinson, P., Spear, F.S., Schumacher, J.C., Laird, J., Klein, C., Evans, B.W. and Doolan, B.L., 1982, Phase relations of metamorphic amphiboles: natural occurrence and theory; in *Veblen, D.R. and Ribbe, P.H.(Eds.), Mineralogical Society of America, Reviews in Mineralogy*, b.9B, p. 1-211.
- Rucklidge, J.C. and Gasparrini, E.L., 1969, Specifications of a computer program for processing electron microprobe analytical data: EMPADR VII; *Department of Geology, University of Toronto, Toronto, Canada*.
- Sen, S.K. and Oliver, R.L., 1981, Secondary hornblendes in some granulite facies rocks from the Mann Ranges, Australia; *J. Geol. Soc. Aust.*, v. 28, p. 137-140.

- Sen, S.K. and Bhattacharya, A., 1984, An orthopyroxene-garnet thermometer and its application to the Madras charnockites; *Contr. Mineral. Petrol.*, v. 88, p. 64-71.
- Shearer, C.K., Papike, J.J., Simon, S.B. and Laul, J.C., 1986, Pegmatite-wallrock interactions, Black Hills, South Dakota: Interaction between pegmatite-derived fluids and quartz-mica schist wallrock; *Am. Min.*, v. 71, p. 518-539.
- Sills, J.C., 1983, Mineralogical changes occurring during the retrogression of Archaean gneisses from the Lewisian complex of NW Scotland; *Lithos*, v. 16, p. 113-124.
- Spear, F.S., 1980, NaSi=CaAl exchange equilibria between plagioclase and amphibole: an empirical model; *Contr. Mineral. Petrol.*, v. 72, p. 33-41.
- Spear, F.S., 1981a, An experimental study of hornblende stability and compositional variability in amphibolite; *Am. J. of Sci.*, v. 281, p. 697-734.
- Spear, F.S., 1981b, Amphibole-plagioclase equilibria: an empirical model for the relation albite + tremolite = edenite + 4quartz; *Contr. Mineral. Petrol.*, v. 77, p. 355-364.
- Stephensen, N.C.N., 1977, Coexisting hornblendes and biotites from Precambrian gneisses of the south coast of Western Australia; *Lithos*, v. 10, p. 9-27.
- Stephensen, N.C.N., 1979, Coexisting garnets and biotites from Precambrian gneisses of the south coast of Western Australia; *Lithos*, v. 12, p. 73-87.
- Stephensen, N.C.N., 1984, Two-pyroxene thermometry of Precambrian granulites from Cape Riche, Albany-Fraser Province, Western Australia; *J. Metamorphic Geol.*, v. 2, p. 297-314.
- Stockwell, C.H., 1962, Second report on structural provinces, orogenesis, and time classification of rocks of the Canadian Precambrian Shield; in Lowdon J.A. et al.(Eds.), *Age Determinations and Geological Studies*, Geological Survey of Canada, Paper 62-17, p.123-133.
- Stout, J.H., 1972, Phase petrology and mineral chemistry of coexisting amphiboles from Telemark, Norway; *J. of Petrol.*, v. 13, p. 99-145.
- Thompson, A.B., 1976, Mineral reactions in pelitic rocks: II calculations of some P-T-X(Fe-Mg) phase relations; *Am. J. of Sci.*, v. 276, p.425-454.

- Tracy, R.J., 1982, Compositional zoning and inclusions in metamorphic minerals; in Ferry, J.M. (Ed.), Mineralogical Society of America, Reviews in Mineralogy, v. 10, p. 355-397.
- van Breemen, O., Davidson, A., Loveridge, W.D. and Sullivan, R.W., 1986, U-Pb zircon geochronology of Grenville tectonites, granulites and igneous precursors, Parry Sound, Ontario; in Moore, J.M., Davidson, A. and Baer, A.J. (Eds.), The Grenville Province, Geological Association of Canada, Special Paper 31, p.191-207.
- Vernon, R.H., 1972, Reactions involving hydration of cordierite and hypersthene; Contr. Mineral. and Petrol., v. 35, p. 125-137.
- Wells, P.R.A., 1977, Pyroxene thermometry in simple and complex systems; Contr. Mineral. Petrol., v. 62, p. 129-139.
- Wendlandt, R.F., 1981, Influence of CO₂ on melting of model granulite facies assemblages: a model for the genesis of charnockites; Am. Min., v. 66, p. 1164-1174.
- Wood, B.J., 1974, The solubility of alumina in orthopyroxene coexisting with garnet; Contr. Mineral. Petrol., v. 46, p. 1-15.
- Wood, B.J., 1979, Activity-composition relations in Ca(Mg,Fe)Si₂O₆-CaAl₂SiO₆ clinopyroxene solid solutions; Am. J. of Sci., v. 279, p.854-875.
- Wood, B.J. and Banno, S., 1973, Garnet-orthopyroxene and orthopyroxene-clinopyroxene relationships in simple and complex systems; Contr. Mineral. Petrol., v. 42, p. 109-124.
- Yardley, B.W.D., 1981, Effect of cooling on the water content and mechanical behavior of metamorphosed rocks; Geology, v. 9, p. 405-408.
- Yardley, B.W.D. and Baltatzis, E., 1985, Retrogression of staurolite schists and the sources of infiltrating fluids during metamorphism; Contr. Mineral. and Petrol., v. 89, p. 59-68.

APPENDIX I

Sketches of Sample Localities

Locality B - Felsic Gneiss

vertical section

Sample

Distance N from peg. (1)

B5aG	25'3"	B5kI	15'2"
B5bG	20'10"	B5lI	10'0"
B5cG	20'11"	B5mA	0'8"
B5dG	20'6"	B5nA	1'0"
B5eG	20'0"	B5oA	2'3"
B5fI	14'0"	B5pI	3'3"
B5gI	15'1"	B5qI	8'3"
B5hI	12'9"	B5rI	8'3"

approx. scale:

2"

approx. limit

of retrogression:

attitude of peg: (1) 300/72

pegmatite:

fol: 30/30; line 55

foliation:

Locality C - Felsic Gneiss

vertical section

approx. scale: 10"
 approx. limit of retrogression:
 attitude of peg:(1)
 307/ 73
 fol: 57/ 28
 pegmatite:
 foliation:

Sample	Distance from peg. (1)	
C7aI	0'8" E	C8cG 11'4" W
C7bI	2'0" E	C8dG 18'4" W
C7cG	3'9" E	C8eG 21'0" W
C7dG	6'6" E	C8fG 29'0" W
C7eG	10'0" E	C8gG 34'2" W
C7fG	5'4" E	C8hG 38'8" W
C8aG	3'1" W	C8iG 41'8" W
C8bG	4'8" W	C8jG 42'9" W

Locality D - K feldspar Augen Gneiss

vertical section

Sample

Distance S from peg.(1)

D9aA 0'2"

D9bA 0'8"

D9cA 2'7"

D9dA 3'2"

D9eI 4'4"

D9fI 5'8"

D9gG 6'5"

D9hG 7'0"

approx. scale:

2 ft

approx. limit
of retrogression:

pegmatite:

foliation:

attitude of peg:(1) 276 / 60

fol. 144 / 18; lin. 162 / 08

Locality E - K feldspar Augen Gneiss

vertical section

Sample

Distances S from contact of mafic body

E15aG	-16'5"	3'5"	F.B.O.	E15jG	61'9"	2'6"	F.B.O.
E15bG	2'0"	2'0"	F.B.O.	E15kA	65'6"	3'0"	F.B.O.
E15cA	11'9"	1'0"	F.B.O.	E15lA	71'1"	6'0"	F.B.O.
E15dI	26'5"	2'6"	F.B.O.	E15mI	68'8"	6'6"	F.B.O.
E15eI	30'6"	5'0"	F.B.O.	E15nI	78'5"	6'0"	F.B.O.
E15fG	42'2"	4'0"	F.B.O.	E15oI	80'3"	5'5"	F.B.O.
E15gG	44'2"	5'6"	F.B.O.	E15pA	84'8"	5'0"	F.B.O.
E15hG	55'2"	2'0"	F.B.O.	E15qA	84'9"	4'9"	F.B.O.
E15iG	56'7"	6'0"	F.B.O.				

approx. scale:

approx. limit
of retrogression:

fol. 04/14

mafic body:

foliation:

F.B.O. = from base of outcrop

Locality F - Felsic Gneiss

vertical section

E

W

approx. limit
of retrogression:

pegmatite:

foliation:

F.B.O. = from base of outcrop

Sample	Distance E from peg. (1)	Distance W from peg. (1)	Distance W from peg. (2)	Distance W from peg. (5)	attitude of peg:	attitude of peg:	attitude of peg:	attitude of peg:	attitude of peg:	attitude of peg:
F18aG	97'0"	1'0"	F.B.O.	F18nI	68'7"	1'6"	F.B.O.	(1)	176	✓68
F18bG	77'2"	2'0"	F.B.O.	F18oA	69'9"	1'6"	F.B.O.	(2)	178	✓63
F18cG	53'8"	3'0"	F.B.O.	F18pA	72'2"	2'0"	F.B.O.	(3)	182	✓78
F18dG	32'0"	5'6"	F.B.O.	F18qA	73'8"	1'6"	F.B.O.	(4)	182	✓78
F18eG	32'0"	5'7"	F.B.O.	F18rA	78'0"	3'0"	F.B.O.	(5)	176	✓72
F18fG	18'7"	3'6"	F.B.O.	F18sA	2'0"	3'0"	F.B.O.	(6)	167	✓71
F18gG	15'8"	0'6"	F.B.O.	F18tA	7'4"	3'0"	F.B.O.			
Distance W from peg. (1)				F18uI	11'9"	4'0"	F.B.O.			
F18hG	10'8"	2'6"	F.B.O.	F18vI	15'5"	4'0"	F.B.O.			
F18iG	25'9"	1'0"	F.B.O.	Distance W from peg. (5)						
F18jG	42'7"	1'6"	F.B.O.	F18wA	5'4"	4'0"	F.B.O.			
F18kI	65'3"	4'6"	F.B.O.	F18xG	40'8"	2'0"	F.B.O.			
F18lG	65'9"	1'0"	F.B.O.	F18yG	54'2"	2'6"	F.B.O.			
F18mG	68'2"	0'6"	F.B.O.	F18zG	71'0"	2'6"	F.B.O.			

fol. 04 / 14

Appendix II

Whole Rock Chemistry

Note: Oxides given in weight %

Trace elements given in ppm

Precision of chemical analysis*

XRF	COV
SiO ₂	0.9
Al ₂ O ₃	0.8
TiO ₂	0.8
Fe	0.8
MnO	4.0
MgO	1.0
CaO	0.8
Na ₂ O	2.0
K ₂ O	1.2
P ₂ O ₅	4.0
Ba	10
Cr	10
Zr	10
Sr	4
Rb	40
Y	20
Titration	COV
FeO	1.0

* Expressed as coefficient of variation.

COV = 100(standard deviation / mean)

+/- c values = 95% confidence limits based on precision of analyses

XRF Analyses of Whole Rocks

Locality A, Mafic Gneiss

<u>Sample Number</u>	<u>AllaG</u>	<u>AllfA</u>	<u>Mean</u>	<u>+/-</u>	<u>c*</u>
SiO ₂	53.22	49.00	51.11	+/-	0.90
Al ₂ O ₃	18.91	16.37	17.64	+/-	0.28
Fe ₂ O ₃	3.54	3.37	3.46	+/-	0.07
FeO	6.7	9.0	7.85	+/-	0.15
MgO	3.20	5.74	4.47	+/-	0.09
CaO	7.90	9.10	8.50	+/-	0.13
Na ₂ O	4.22	3.12	3.67	+/-	0.14
K ₂ O	0.66	0.85	0.76	+/-	0.02
TiO ₂	1.43	1.17	1.30	+/-	0.02
P ₂ O ₅	0.41	0.12	0.27	+/-	0.02
MnO	0.20	0.17	0.19	+/-	0.01
Ba	266 +/-52	416 +/-82			
Cr	66 +/-13	78 +/-15			
Zr	168 +/-33	26 +/-5			
Sr	611 +/-48	484 +/-38			
Rb	<10	<10			
Y	39 +/-15	19 +/-4			

*c = +/- 1.96 SD for mean value

H ₂ O	0.6	1.4
	+/-0.1	+/-0.1

XRF Analyses of Whole Rocks

Locality B, Felsic Gneiss

<u>Sample Number</u>	<u>B5cG</u>	<u>B5pI</u>	<u>B5nA</u>	<u>B5mA</u>	<u>Mean +/-c*</u>
SiO ₂	71.16	71.19	71.75	72.87	71.74 +/- 1.27
Al ₂ O ₃	13.78	13.56	13.51	12.84	13.42 +/- 0.21
Fe ₂ O ₃	0.80	1.18	1.31	1.51	1.20 +/- 0.02
FeO	2.6	2.6	2.1	1.9	2.3 +/- 0.04
MgO	0.42	0.62	0.49	0.37	0.48 +/- 0.01
CaO	1.62	1.76	1.57	1.29	1.56 +/- 0.02
Na ₂ O	2.82	2.73	3.03	2.63	2.80 +/- 0.11
K ₂ O	5.28	4.98	4.60	5.43	5.07 +/- 0.12
TiO ₂	0.43	0.51	0.43	0.41	0.45 +/- 0.01
P ₂ O ₅	0.06	0.03	0.05	0.07	0.05 +/- 0.01
MnO	0.07	0.09	0.05	0.06	0.07 +/- 0.01
Ba	1039 +/-204	1014 +/-199	1175 +/-230	1454 +/-285	
Cr	118 +/-23	89 +/-17	112 +/-22	98 +/-19	
Zr	429 +/-84	501 +/-98	413 +/-81	491 +/-96	
Sr	135 +/-11	173 +/-14	216 +/-17	219 +/-17	
Rb	163 +/-128	149 +/-117	107 +/-84	114 +/-89	
Y	73 +/-29	73 +/-29	49 +/-19	48 +/-19	

*c = +/- 1.96 SD for mean value

H ₂ O	0.3 +/-0.1	0.3 +/-0.1	0.3 +/-0.1	0.3 +/-0.1
------------------	---------------	---------------	---------------	---------------

XRF Analyses of Whole Rocks

Locality C, Felsic Gneiss

<u>Sample Number</u>	<u>C81G</u>	<u>C7dG</u>	<u>C7aI</u>	<u>Mean +/-c*</u>
SiO ₂	65.14	65.45	64.80	65.13 +/- 1.15
Al ₂ O ₃	14.87	14.71	14.90	14.83 +/- 0.23
Fe ₂ O ₃	1.02	0.71	1.17	0.97 +/- 0.02
FeO	5.4	5.3	5.3	5.33 +/- 0.10
MgO	0.45	0.68	0.95	0.69 +/- 0.01
CaO	3.04	2.87	2.59	2.83 +/- 0.04
Na ₂ O	3.14	3.59	3.05	3.26 +/- 0.13
K ₂ O	5.14	4.96	4.86	4.99 +/- 0.12
TiO ₂	0.74	0.85	0.83	0.81 +/- 0.01
P ₂ O ₅	0.22	0.27	0.26	0.25 +/- 0.02
MnO	0.12	0.12	0.15	0.13 +/- 0.01
Ba	2201 +/-431	1955 +/-383	2036 +/-399	
Cr	114 +/-22	98 +/-19	90 +/-18	
Zr	798 +/-156	486 +/-95	776 +/-152	
Sr	293 +/-23	212 +/-17	282 +/-22	
Rb	95 +/-74	82 +/-64	115 +/-90	
Y	62 +/-24	45 +/-18	64 +/-25	
*c = +/- 1.96 SD for mean value				
H ₂ O	0.3 +/-0.1	0.5 +/-0.1	0.9 +/-0.1	

XRF Analyses of Whole Rocks

Locality D, K feldspar Augen Gneiss

<u>Sample Number</u>	<u>D9hG</u>	<u>D9fI</u>	<u>D9aA</u>	<u>Mean +/-c*</u>
SiO ₂	64.78	66.85	65.83	65.82 +/- 1.16
Al ₂ O ₃	14.60	14.15	14.61	14.45 +/- 0.23
Fe ₂ O ₃	2.17	1.38	1.71	1.75 +/- 0.03
FeO	4.0	3.9	3.7	3.87 +/- 0.08
MgO	1.71	1.62	1.44	1.59 +/- 0.03
CaO	3.35	2.74	2.89	2.99 +/- 0.05
Na ₂ O	3.03	2.87	3.01	2.97 +/- 0.12
K ₂ O	3.52	3.82	4.04	3.79 +/- 0.09
TiO ₂	1.12	0.96	1.02	1.03 +/- 0.02
P ₂ O ₅	0.39	0.30	0.36	0.35 +/- 0.03
MnO	0.10	0.10	0.12	0.11 +/- 0.01
Ba	1525 +/-299	1515 +/-297	1537 +/-301	
Cr	115 +/-23	133 +/-26	151 +/-30	
Zr	508 +/-100	460 +/-90	500 +/-98	
Sr	282 +/-22	264 +/-21	277 +/-22	
Rb	82 +/-64	97 +/-76	95 +/-74	
Y	57 +/-22	48 +/-19	45 +/-18	
*c = +/- 1.96 SD for mean value				
H ₂ O	0.4 +/-0.1	0.7 +/-0.1	0.6 +/-0.1	

XRF Analyses of Whole Rocks

Locality E, Augen Gneiss

<u>Sample Number</u>	<u>E15bG</u>	<u>E15cA</u>	<u>Mean +/-c*</u>
SiO ₂	64.56	67.14	65.85 +/- 1.16
Al ₂ O ₃	15.01	14.02	14.52 +/- 0.23
Fe ₂ O ₃	1.61	0.91	1.26 +/- 0.02
FeO	4.1	3.9	4.00 +/- 0.08
MgO	1.69	1.49	1.59 +/- 0.03
CaO	3.51	3.07	3.29 +/- 0.05
Na ₂ O	3.26	3.00	3.13 +/- 0.12
K ₂ O	3.89	3.44	3.67 +/- 0.09
TiO ₂	1.07	0.87	0.97 +/- 0.02
P ₂ O ₅	0.41	0.28	0.35 +/- 0.03
MnO	0.11	0.10	0.11 +/- 0.01
Ba	1613 +/-316	1054 +/-207	
Cr	108 +/-21	124 +/-24	
Zr	379 +/-74	467 +/-91	
Sr	319 +/-25	243 +/-19	
Rb	87 +/-68	97 +/-76	
Y	55 +/-22	63 +/-25	

*c = +/- 1.96 Sd for mean value

H ₂ O	0.3 +/-0.1	0.5 +/-0.1
------------------	---------------	---------------

XRF Analyses of Whole Rocks

Locality F, Felsic Gneiss

<u>Sample Number</u>	<u>F18aG</u>	<u>F18yG</u>	<u>F18mG</u>	<u>F18uI</u>	<u>F18vI</u>	<u>F18nI</u>
SiO ₂	61.64	67.94	69.27	68.34	69.37	68.82
Al ₂ O ₃	13.81	13.35	13.48	13.19	13.58	13.48
Fe ₂ O ₃	2.69	1.72	1.50	1.81	1.42	1.19
FeO	5.8	2.8	3.0	3.1	3.1	3.4
MgO	1.49	0.68	0.64	0.64	0.63	0.84
CaO	4.24	2.14	1.99	2.12	1.95	2.16
Na ₂ O	3.00	2.75	2.81	2.58	2.68	2.56
K ₂ O	3.90	5.28	5.18	5.03	5.25	5.20
TiO ₂	1.59	0.56	0.56	0.60	0.55	0.59
P ₂ O ₅	0.66	0.09	0.11	0.13	0.11	0.08
MnO	0.16	0.09	0.09	0.10	0.10	0.09
Ba	1168 +/-229	827 +/-162	778 +/-152	773 +/-152	809 +/-159	840 +/-165
Cr	70 +/-14	80 +/-16	94 +/-18	93 +/-18	89 +/-17	112 +/-22
Zr	628 +/-123	573 +/-112	669 +/-131	701 +/-137	617 +/-121	632 +/-124
Sr	229 +/-18	97 +/-8	109 +/-9	79 +/-6	83 +/-7	137 +/-11
Rb	119 +/-93	176 +/-138	169 +/-132	172 +/-135	181 +/-142	170 +/-133
Y	84 +/-33	92 +/-36	98 +/-38	106 +/-42	107 +/-42	93 +/-36
H ₂ O	0.2 +/-0.1	0.3 +/-0.1	0.3 +/-0.1	0.3 +/-0.1	0.3 +/-0.1	0.3 +/-0.1

XRF Analyses of Whole Rocks

Locality F continued

<u>Sample</u>	<u>Number</u>	<u>F18rA</u>	<u>F18sA</u>	<u>Mean</u>	<u>+/-</u>	<u>c*</u>
SiO ₂		68.32	68.57	68.66	+/-	1.21
Al ₂ O ₃		13.44	13.35	13.44	+/-	0.21
Fe ₂ O ₃		1.35	1.61	1.51	+/-	0.05
FeO		3.2	3.2	3.11	+/-	0.05
MgO		0.74	0.61	0.68	+/-	0.01
CaO		2.37	2.22	2.14	+/-	0.03
Na ₂ O		2.56	2.49	2.63	+/-	0.10
K ₂ O		5.53	5.64	5.30	+/-	0.12
TiO ₂		0.63	0.62	0.59	+/-	0.01
P ₂ O ₅		0.16	0.13	0.12	+/-	0.01
MnO		0.08	0.09	0.09	+/-	0.01
Ba		1080	1308			
		+/-212	+/-256			
Cr		71	74			
		+/-14	+/-15			
Zr		736	640			
		+/-144	+/-125			
Sr		129	183			
		+/-10	+/-14			
Rb		127	111			
		+/-100	+/-87			
Y		112	103			
		+/-44	+/-40			

*c = +/- 1.96 SD for mean value excluding F18aG

H ₂ O	0.5	0.3
	+/-0.1	+/-0.1

APPENDIX III

Electron Microprobe Analyses

- III.1 Pyroxene
- III.2 Garnet
- III.3 Amphibole
- III.4 Biotite
- III.5 Plagioclase

III.1 Electron Microprobe Analyses of Orthopyroxene

Sample Number	AllaG (3)	B5cG * (8)	D9hG (6)	E15bG (7)	F18aG (4)	F18yG (7)	B3aG (2)
SiO ₂ (wt.%)	50.25	50.04	51.58	50.79	50.40	47.85	49.96
TiO ₂	0.16	0.04	0.05	0.07	0.07	0.02	0.09
Al ₂ O ₃	1.53	1.10	1.20	1.02	0.87	0.50	1.69
Cr ₂ O ₃	0.06	-	-	-	-	-	0.11
FeO*	27.99	34.27	27.08	31.32	35.63	38.93	29.72
MnO	0.65	0.70	1.38	1.16	0.81	0.79	0.50
MgO	19.05	13.26	17.95	15.40	12.59	10.00	17.55
CaO	0.36	0.56	0.45	0.51	0.60	0.62	0.41
Na ₂ O	-	0.02	-	-	-	0.11	-
K ₂ O	-	-	-	-	-	-	-
Total	100.05	99.99	99.69	100.27	100.97	98.82	100.05

*Fe as FeO Cations on the basis of 6 oxygens

Si	1.915	1.979	1.979	1.977	1.985	1.964	1.922
Al(4)	0.069	0.021	0.021	0.023	0.015	0.024	0.077
Al(6)	-	0.030	0.033	0.024	0.026	-	-
Ti	0.005	0.001	0.001	0.002	0.002	0.001	0.003
Cr	0.002	-	-	-	-	-	0.003
Fe(3+)	0.090	-	-	-	-	0.056	0.071
Fe(2+)	0.802	1.133	0.869	1.019	1.174	1.281	0.885
Mn	0.021	0.023	0.045	0.038	0.027	0.028	0.016
Mg	1.082	0.782	1.027	0.894	0.739	0.612	1.006
Ca	0.015	0.024	0.019	0.021	0.025	0.027	0.017
Na	-	0.002	-	-	-	0.009	-
K	-	-	-	-	-	-	0.001

III.1 Electron Microprobe Analyses of Clinopyroxene

Sample Number	AllaG	B5cG	C7dG	F18aG	F18yG	B3aG
(no. anal.)	(4)	(4)	(9)	(5)	(6)	(6)
SiO ₂ (wt.%)	50.57	51.88	49.22	51.07	50.83	51.23
TiO ₂	0.23	0.11	0.07	0.16	0.19	0.22
Al ₂ O ₃	2.79	1.54	1.40	1.86	1.51	3.07
Cr ₂ O ₃	0.09	-	0.10	-	-	0.08
FeO*	11.75	14.57	22.99	15.54	16.11	10.99
MnO	0.29	0.26	0.42	0.30	0.44	0.17
MgO	12.22	9.94	5.62	9.48	8.93	12.32
CaO	21.24	20.52	19.69	21.03	20.98	21.72
Na ₂ O	0.50	1.52	0.46	-	0.80	-
K ₂ O	0.07	-	0.09	-	-	0.07
Total	99.75	100.34	100.06	99.44	99.76	99.87
*Fe as FeO	Cations on the basis of 6 oxygens					
Si	1.905	1.959	1.948	1.970	1.957	1.929
Al(4)	0.095	0.041	0.052	0.030	0.043	0.072
Al(6)	0.029	0.028	0.014	0.055	0.024	0.064
Ti	0.007	0.003	0.002	0.005	0.006	0.006
Cr	0.003	-	0.003	-	-	0.002
Fe(3+)	0.091	0.119	0.071	-	0.067	-
Fe(2+)	0.279	0.342	0.690	0.501	0.452	0.346
Mn	0.009	0.008	0.014	0.010	0.013	0.005
Mg	0.686	0.560	0.332	0.545	0.512	0.692
Ca	0.857	0.830	0.835	0.869	0.865	0.876
Na	0.037	0.111	0.035	-	0.060	-
K	0.003	-	0.005	-	-	0.003

III.2 Electron Microprobe Analyses of Garnet

Sample Number	AllaG	AllfA	B5cG	B5pI	C7dG	C7aI
(no. anal.)	(9)	(6)	(7)	(5)	(18)	(32)
SiO ₂ (wt.%)	38.03	37.75	37.49	37.58	37.06	37.10
TiO ₂	-	-	-	-	-	-
Al ₂ O ₃	22.15	22.10	21.43	21.05	21.47	21.42
Cr ₂ O ₃	0.07	0.03	-	-	0.04	0.06
FeO*	27.99	23.83	28.92	26.61	31.74	28.76
MnO	1.54	5.56	1.76	3.16	1.40	2.95
MgO	5.12	4.90	2.77	2.85	1.33	0.96
CaO	7.32	5.89	7.29	7.87	6.91	8.67
Na ₂ O	-	-	-	-	-	-
K ₂ O	-	-	-	-	-	-
Total	100.10	100.06	99.66	99.12	99.95	99.91
*Fe as FeO	Cations on the basis of 12 oxygens					
Si	2.915	2.959	2.988	3.006	2.979	2.979
Al(4)	0.085	0.041	0.012	-	0.021	0.021
Al(6)	1.916	2.000	2.001	1.985	2.013	2.005
Ti	-	-	-	-	-	-
Cr	0.004	0.002	-	-	0.003	0.004
Fe(3+)	0.165	0.039	0.010	0.003	0.005	0.014
Fe(2+)	1.629	1.523	1.918	1.778	2.130	1.917
Mn	0.100	0.369	0.119	0.214	0.095	0.201
Mg	0.585	0.572	0.329	0.340	0.159	0.115
Ca	0.601	0.495	0.623	0.675	0.595	0.746
Na	-	-	-	-	-	-
K	-	-	-	-	-	-

III.2 Electron Microprobe Analyses of Garnet

Sample Number	D9aA	E15cA	F18aG	F18yG	F18nI	B3aG
<u>(no. anal.)</u>	<u>(7)</u>	<u>(9)</u>	<u>(5)</u>	<u>(6)</u>	<u>(6)</u>	<u>(6)</u>
SiO ₂ (wt.%)	37.30	37.07	37.34	36.75	36.93	37.97
TiO ₂	-	-	-	-	-	-
Al ₂ O ₃	21.36	21.32	21.34	20.98	21.34	22.19
Cr ₂ O ₃	-	-	-	-	-	0.09
FeO*	23.15	28.91	29.10	30.63	30.16	26.06
MnO	8.35	3.36	2.15	1.75	2.42	1.11
MgO	3.49	3.48	3.22	2.59	1.86	5.18
CaO	5.99	5.37	6.19	7.22	6.50	7.19
Na ₂ O	-	-	-	-	-	-
K ₂ O	-	-	-	-	-	-
Total	99.64	99.51	99.34	99.92	99.20	99.78
*Fe as FeO	Cations on the basis of 12 oxygens					
Si	2.968	2.961	2.985	2.935	2.982	2.968
Al(4)	0.032	0.039	0.015	0.065	0.018	0.032
Al(6)	1.971	1.968	1.996	1.910	2.013	2.012
Ti	-	-	-	-	-	-
Cr	-	-	-	-	-	0.006
Fe(3+)	0.060	0.071	0.020	0.155	0.006	0.015
Fe(2+)	1.481	1.860	1.925	1.891	2.030	1.689
Mn	0.563	0.227	0.146	0.118	0.166	0.074
Mg	0.414	0.414	0.384	0.308	0.224	0.604
Ca	0.511	0.460	0.530	0.618	0.562	0.602
Na	-	-	-	-	-	-
K	-	-	-	-	-	-

III.3 Electron Microprobe Analyses of Amphibole

Sample Number	AllaG	AllfA	B5pI	B5mA	C7dG	C7aI
(no. anal.)	(6)	(6)	(5)	(4)	(12)	(10)
SiO ₂ (wt.%)	42.41	42.99	42.48	40.06	40.81	39.69
TiO ₂	1.42	1.19	0.49	1.72	1.73	1.54
Al ₂ O ₃	12.29	12.13	11.18	12.02	10.32	11.61
Cr ₂ O ₃	0.08	0.07	-	-	0.07	0.06
FeO*	16.35	15.79	19.26	20.24	28.07	28.64
MnO	0.08	0.39	0.19	0.87	0.11	0.19
MgO	11.36	11.81	10.22	7.85	3.80	3.06
CaO	11.28	11.05	11.28	10.88	10.44	10.57
Na ₂ O	1.77	2.06	1.66	1.74	1.85	1.73
K ₂ O	1.40	1.13	1.25	1.52	1.48	1.67
Total	98.42	98.62	98.01	96.90	98.68	98.76
*Fe as FeO	23 oxygens - all cations except Ca, Na, K = 13					
Si	6.216	6.253	6.309	6.135	6.349	6.198
Al(4)	1.784	1.747	1.691	1.865	1.651	1.802
Al(6)	0.339	0.332	0.266	0.305	0.241	0.335
Ti	0.157	0.130	0.055	0.198	0.202	0.181
Cr	0.009	0.008	-	-	0.009	0.007
Fe(3+)	0.817	0.912	1.010	0.780	0.664	0.706
Fe(2+)	1.187	1.009	1.382	1.812	2.988	3.034
Mn	0.010	0.048	0.024	0.113	0.015	0.025
Mg	2.482	2.561	2.263	1.792	0.881	0.712
Ca	1.771	1.722	1.795	1.785	1.740	1.768
Na	0.503	0.581	0.478	0.517	0.558	0.524
K	0.262	0.210	0.237	0.297	0.294	0.333

III.3 Electron Microprobe Analyses of Amphibole

Sample Number	D9hG	F18aG	F18yG	F18rA	F18sA	B3aG
(no. anal.)	(4)	(4)	(4)	(3)	(7)	(7)
SiO ₂ (wt%)	41.24	41.04	41.36	38.67	40.45	41.82
TiO ₂	1.89	1.75	1.41	1.50	1.47	1.85
Al ₂ O ₃	12.15	11.63	10.96	12.23	11.61	13.35
Cr ₂ O ₃	-	-	-	-	0.03	0.08
FeO*	17.20	21.08	23.22	24.61	24.06	16.79
MnO	0.33	0.16	0.26	0.57	0.49	0.02
MgO	10.37	8.45	7.33	5.58	6.17	10.77
CaO	11.63	11.21	11.04	11.12	10.68	11.43
Na ₂ O	1.39	1.65	1.71	1.76	1.68	1.08
K ₂ O	1.65	1.73	1.47	1.71	1.69	1.50
Total	97.85	98.70	98.76	97.75	98.32	98.70
*Fe as FeO 23 Oxygens - all cations except Ca, Na, K = 13						
Si	6.155	6.173	6.259	6.014	6.187	6.100
Al(4)	1.845	1.827	1.741	1.986	1.813	1.900
Al(6)	0.292	0.235	0.214	0.255	0.278	0.523
Ti	0.212	0.198	0.161	0.175	0.169	0.203
Cr	-	-	-	-	0.004	0.009
Fe(3+)	0.693	0.769	0.841	0.805	0.864	0.932
Fe(2+)	1.454	1.883	2.098	2.396	2.214	1.116
Mn	0.042	0.020	0.033	0.075	0.064	0.003
Mg	2.307	1.895	1.654	1.294	1.407	2.342
Ca	1.860	1.807	1.790	1.853	1.750	1.786
Na	0.402	0.481	0.502	0.531	0.498	0.305
K	0.314	0.332	0.284	0.339	0.330	0.279

III.4 Electron Microprobe Analyses of Biotite

Sample Number	AllaG	AllfA	B5cG	B5pI	B5mA
<u>(no. anal.)</u>	<u>(8)</u>	<u>(7)</u>	<u>(2)</u>	<u>(6)</u>	<u>(3)</u>
SiO ₂ (wt.%)	36.03	36.16	36.99	36.68	35.34
TiO ₂	5.69	5.46	5.09	4.58	4.71
Al ₂ O ₃	14.44	15.43	14.11	14.84	14.95
Cr ₂ O ₃	0.03	0.01	-	-	-
FeO*	17.07	15.81	20.61	20.21	20.90
MnO	0.02	0.12	-	0.04	0.36
MgO	13.14	13.71	10.42	10.98	10.03
CaO	-	-	-	-	-
Na ₂ O	0.12	-	0.04	0.06	0.09
K ₂ O	9.46	9.50	9.07	9.30	9.23
Total	95.99	96.20	96.33	96.69	95.64
*Fe as FeO	Cations on the basis of 22 oxygens				
Si	5.422	5.388	5.605	5.459	5.439
Al(4)	0.578	0.612	0.395	0.541	0.561
Al(6)	1.983	2.098	2.126	2.136	2.151
Ti	0.644	0.612	0.581	0.527	0.545
Cr	0.004	0.001	-	-	-
Fe	2.148	1.970	2.613	2.587	2.690
Mn	0.002	0.015	-	0.005	0.048
Mg	2.947	3.045	2.353	2.504	2.302
Ca	-	-	-	-	-
Na	0.035	-	0.012	0.019	0.028
K	1.817	1.805	1.755	1.815	1.812

III.4 Electron Microprobe Analyses of Biotite

Sample Number	C7dG	C7aI	D9hG	D9aA	E15bG	E15cA
<u>(no. anal.)</u>	<u>(7)</u>	<u>(8)</u>	<u>(7)</u>	<u>(6)</u>	<u>(4)</u>	<u>(7)</u>
SiO ₂ (wt.%)	34.26	33.53	37.18	37.01	36.77	36.30
TiO ₂	5.59	5.34	4.95	4.40	5.82	5.35
Al ₂ O ₃	12.87	13.31	14.98	15.51	14.47	14.50
Cr ₂ O ₃	0.02	0.02	-	-	-	-
FeO*	30.50	31.15	18.08	16.47	21.15	20.19
MnO	0.01	0.07	0.14	0.21	0.39	0.05
MgO	3.86	3.54	12.16	13.18	9.97	11.29
CaO	-	0.01	-	-	-	-
Na ₂ O	-	0.02	0.01	0.04	0.01	0.03
K ₂ O	8.87	8.73	9.11	9.34	8.96	9.23
Total	95.99	95.71	96.61	96.16	97.54	96.94
*Fe as FeO	Cations on the basis of 22 oxygens					
Si	5.516	5.439	5.546	5.515	5.531	5.472
Al(4)	0.484	0.561	0.454	0.485	0.469	0.528
Al(6)	1.959	1.984	2.181	2.238	2.096	2.049
Ti	0.677	0.651	0.555	0.493	0.659	0.606
Cr	0.003	0.002	-	-	-	-
Fe	4.107	4.226	2.256	2.054	2.661	2.545
Mn	0.001	0.010	0.017	0.027	0.013	0.006
Mg	0.925	0.855	2.704	2.926	2.237	2.536
Ca	-	0.001	-	-	-	-
Na	-	0.006	0.002	0.010	0.002	0.009
K	1.823	1.807	1.734	1.775	1.721	1.774

III.4 Electron Microprobe Analyses of Biotite

Sample Number	F18yG	F18rA	B3aG
(no. anal.)	(4)	(6)	(6)
SiO ₂ (wt.%)	35.81	34.64	35.12
TiO ₂	5.17	4.66	6.44
Al ₂ O ₃	14.05	14.41	14.23
Cr ₂ O ₃	-	-	0.03
FeO*	24.05	25.51	19.19
MnO	0.09	0.33	0.01
MgO	8.95	7.39	11.73
CaO	-	0.01	-
Na ₂ O	0.07	0.10	-
K ₂ O	9.36	9.22	9.06
Total	97.55	96.27	95.81
*Fe as FeO	22 oxygen basis		
Si	5.479	5.429	5.349
Al(4)	0.521	0.571	0.651
Al(6)	2.012	2.091	1.903
Ti	0.595	0.549	0.738
Cr	-	-	0.004
Fe	3.078	3.344	2.444
Mn	0.012	0.051	0.002
Mg	2.040	1.725	2.664
Ca	-	0.002	-
Na	0.021	0.032	-
K	1.827	1.843	1.761

III.5 Electron Microprobe Analyses of Plagioclase

Sample Number	AllaG	AllfA	B5cG	B5pI	B5mA
<u>(no. anal.)</u>	<u>(9)</u>	<u>(8)</u>	<u>(3)</u>	<u>(5)</u>	<u>(7)</u>
SiO ₂ (wt.%)	58.59	57.82	62.21	61.15	62.28
Al ₂ O ₃	26.38	26.73	23.86	24.65	23.87
FeO*	0.19	0.12	0.50	0.12	0.06
CaO	7.38	7.73	4.72	5.63	4.68
Na ₂ O	7.29	7.18	8.77	8.49	8.60
K ₂ O	0.36	0.26	0.24	0.27	0.21
Total	100.20	99.83	100.30	100.31	99.70
*Fe as FeO	Cations on the basis of 32 oxygens				
Si	10.472	10.377	11.013	10.846	11.053
Al	5.518	5.653	4.980	5.154	4.994
Fe	0.014	0.004	0.075	0.018	0.009
Ca	1.414	1.486	0.895	1.070	0.890
Na	2.528	2.498	3.012	2.920	2.961
K	0.082	0.061	0.053	0.060	0.049
An	0.351	0.367	0.226	0.264	0.228
Ab	0.628	0.618	0.761	0.721	0.759

III.5 Electron Microprobe Analyses of Plagioclase

Sample Number	C7dG	C7aI	D9hG	D9aA	E15bG	E15cA
<u>(no. anal.)</u>	<u>(9)</u>	<u>(16)</u>	<u>(7)</u>	<u>(8)</u>	<u>(5)</u>	<u>(5)</u>
SiO ₂ (wt.%)	63.86	62.84	59.95	60.11	60.14	60.10
Al ₂ O ₃	22.63	22.94	25.99	25.04	24.75	25.10
FeO*	0.07	0.08	0.16	0.26	0.11	0.10
CaO	3.17	3.87	6.76	6.35	6.40	6.35
Na ₂ O	10.27	9.67	7.67	7.72	7.81	7.84
K ₂ O	0.20	0.21	0.21	0.31	0.26	0.25
Total	100.20	99.61	100.74	99.79	99.47	99.74

*Fe as FeO

Cations on the basis of 32 oxygens

Si	11.277	11.178	10.611	10.733	10.768	10.730
Al	4.711	4.810	5.423	5.271	5.223	5.283
Fe	0.004	0.006	0.024	0.038	0.016	0.016
Ca	0.599	0.738	1.281	1.216	1.228	1.215
Na	3.515	3.336	2.632	2.674	2.712	2.712
K	0.043	0.048	0.048	0.071	0.060	0.058
An	0.144	0.179	0.323	0.307	0.307	0.305
Ab	0.846	0.809	0.665	0.675	0.678	0.681

III.5 Electron Microprobe Analyses of Plagioclase

Sample Number	F18aG	F18yG	F18sA	B3aG
(no. anal.)	(4)	(4)	(5)	(10)
SiO ₂ (wt.%)	62.44	62.65	62.90	57.94
Al ₂ O ₃	24.07	23.65	23.25	26.77
FeO*	-	-	0.11	0.13
CaO	4.94	4.63	3.94	7.75
Na ₂ O	8.77	9.01	9.63	7.14
K ₂ O	0.20	0.22	0.21	0.26
Total	100.40	100.16	100.04	99.99

*Fe as FeO Cations on the basis of 32 oxygens

Si	11.017	11.080	11.147	10.382
Al	5.006	4.929	4.857	5.653
Fe	-	-	-	0.010
Ca	0.935	0.879	0.750	1.489
Na	3.001	3.091	3.309	2.479
K	0.045	0.048	0.047	0.059
An	0.235	0.219	0.183	0.370
Ab	0.754	0.769	0.806	0.616

APPENDIX IV

Miscellaneous

IV.1 Modal Proportions

Min.* Rock ⁺	Locality							
	FG	FG	KAG	FG	FG	FG	FG	FG
	B A#	C G	D I	F G	F G	F I	F I	F A
Ksp	40	45	40	35	35	35	35	35
Pl	20	15	15	20	20	15	20	15
Qtz	30	20	25	25	25	30	25	30
Opx	-	1	1	1	1	-	-	-
Cpx	-	5	-	2	3	-	-	-
Grt	-	6	4	5	4	5	5	-
Bio	3	1	10	2	3	3	4	4
Hbd	4	4	2	7	6	7	6	8
Mag	2	2	2	4	3	4	3	2
Ilm		tr		-				tr
Pyr		tr		tr				-
Py		-		-				tr
Cpy		-		-				-
Apa	tr	tr	tr	tr	tr	tr	tr	3
Ti	-	-	-	-	-	-	-	tr
All	-	-	-	-	-	-	-	-

Visual Estimates

* Ksp=K feldspar, Pl=plagioclase, Qtz=quartz, Opx=orthopyroxene, Cpx=clinopyroxene, Grt=garnet, Bio=biotite, Hbd=hornblende, Mag=magnetite, Ilm=ilmenite, Pyr=pyrrhotite, Py=pyrite, Cpy=chalcopyrite, Apa=apatite, Ti=titanite, All=allanite;

G=granulite, I=intermediate, A=amphibolite.

+ FG=felsic gneiss, KAG=K feldspar augen gneiss.

Samples (from left to right) B5nA, C81G, D9fI, F18yG, F18mG, F18uI, F18vI, F18rA.

Note: Opaque minerals in B5nA, D9fI, F18mG, F18uI and F18vI not determined - all opaque minerals assumed to be magnetite; Biotite and Hornblende in samples marked G probably is retrograde; Garnet in samples marked A are retrograde.

IV.2 Electron Microprobe
Analysis of Magnetite

Sample Number F18yG

(no. anal.) (8)

SiO₂ (wt.%) 0.38

TiO₂ 0.75

Al₂O₃ 0.53

FeO^{*} 89.61

MnO 0.08

MgO 0.04

Total 91.39

^{*}Fe as FeO

Si 0.156

Ti 0.232

Al 0.258

Fe 30.785

Mn 0.028

Mg 0.025

IV.3 Electron Microprobe Analyses of
K feldspars

Sample Number	C7dG	C7aI	F18sA
<u>(no. anal.)</u>	<u>(1)</u>	<u>(5)</u>	<u>(2)</u>
SiO ₂ (wt.%)	65.81	64.88	64.63
Al ₂ O ₃	19.07	19.20	19.01
FeO*	0.03	0.02	-
CaO	0.00	0.20	0.23
Na ₂ O	1.87	1.80	1.87
K ₂ O	14.03	13.38	13.74
Total	100.81	99.47	99.49

*Fe as FeO Cations based on 32 oxygens

Si	11.957	11.918	11.906
Al	4.084	4.157	4.129
Ca	-	0.039	0.046
Na	0.655	0.643	0.668
K	3.254	3.136	3.228
An	-	1.0	1.2
Ab	16.8	16.8	16.9
Or	81.9	83.2	82.1

END

20/06/88

FIN